Rethinking Late Weichselian ice-sheet dynamics in coastal NW Svalbard

JON Y. LANDVIK, ÖLAFUR INGÖLFSSON, JÜRGEN MIENERT, SCOTT J. LEHMAN, ANDERS SOLHEIM, ANDERS ELVERHØI AND DAG OTTESEN


New marine geological evidence provides a better understanding of ice-sheet dynamics along the western margin of the last Svalbard/Barents Sea Ice Sheet. A suite of glacial sediments in the Kongsfjordrenna cross-shelf trough can be traced southwards to the shelf west of Prins Karls Forland. A prominent moraine system on the shelf shows minimum Late Weichselian ice extent, indicating that glacial ice also covered the coastal lowlands of northwest Svalbard. Our results suggest that the cross-shelf trough was filled by a fast-flowing ice stream, with sharp boundaries to dynamically less active ice on the adjacent shelves and strandflats. The latter glacial mode favoured the preservation of older geological records adjacent to the main pathway of the Kongsfjorden glacial system. We suggest that the same model may apply to the Late Weichselian glacier drainage along other fjords of northwest Svalbard, as well as the western margin of the Barents Ice Sheet. Such differences in glacier regime may explain the apparent contradictions between the marine and land geological record, and may also serve as a model for glaciation dynamics in other fjord regions.

Jon Y. Landvik (e-mail: jon.landvik@umb.no), Norwegian University of Life Sciences, Department of Plant and Environmental Sciences, P.O. Box 5003, NO-1432 As, Norway; Olafur Ingólfsson, Department of Geology and Geography, University of Iceland, IS-101 Reykjavik, Iceland; Jürgen Mienert, University of Tromso, Department of Geology, Dramsvæien 201, NO-9037 Tromso, Norway; Scott J. Lehman, University of Colorado, INSTAAR, Campus Box 436, Boulder, CO 80309, USA; Anders Solheim, Norwegian Geotechnical Institute, P.O. Box 3930 Ullevaal Stadion, NO-0806 Oslo, Norway; Anders Elverhøi, University of Oslo, Department of Geosciences, P.O. Box 1047 Blindern, NO-0316 Oslo, Norway; Dag Ottesen, Geological Survey of Norway, Leiv Eiriksons vei 39, NO-7491 Trondheim, Norway; received 19th April 2004, accepted 3rd August 2004.

The size and dynamics of former ice sheets remain the subject of recurrent debate, especially with regard to areas close to the former ice margins. An understanding of the temporal and spatial changes of late Quaternary ice sheets is fundamental to an improved understanding of their interaction with global sea level, climate change (including possible climate/ice-sheet feedback mechanisms) and local and regional sediment distribution.

There is growing consensus among scientists on the larger-scale extent of the Northern Hemisphere late Quaternary ice sheets (see review in Brochmann et al. 2003). However, the detailed reconstruction of ice geometry, especially in coastal and shelf regions, remains a challenge to glacial geologists because of the often fragmentary and complex geological record left by the ice sheets. The main reasons are the transgressive nature of glacier build-up and decay; rapid oscillations of ice-sheet margins and associated complexity of ice-flow patterns; and variation in substratum conditions from, for example, solid bedrock to soft and deformable sediments in fjords, troughs and on the shelves. A better understanding of the geological record in the marginal areas, and of the associated ice geometries and dynamics, has recently been obtained from coastal, ice-marginal settings of Greenland (Funder & Hansen 1996; Funder et al. 1998), the Canadian Arctic (Steig et al. 1998; Jennings et al. 1998; England 1999; Kaplan et al. 2001; Miller et al. 2002) and western Scandinavia (Sejrup et al. 1998, 2003; King et al. 1998; Larsen et al. 2000).

We have studied a similar setting in the Kongsfjorden area, at the west coast of Svalbard (Fig. 1). During periods when the Barents Sea was covered by large ice sheets (Landvik et al. 1998; Mangerud et al. 1998), the coastal areas of Svalbard and the adjacent continental shelf experienced conditions that left complex, and apparently contradictory, geological records (see discussion in the next paragraph).

Despite a number of geomorphic and stratigraphic studies during the past two decades, as well as accumulating marine geological data, there remains ambiguity regarding ice-sheet dynamics and glacial pathways during the Last Glacial Maximum (LGM). Ice-free conditions in certain areas on the west coast of Svalbard during the LGM have been interpreted from stratigraphical and morphological observations (Salvigsen 1977; Miller 1982; Forman & Miller 1984; Forman 1989; Miller et al. 1989; Andersson et al. 1999), whereas marine geological data suggest that Late Weichselian glaciers inundated the major fjord troughs and parts of the shallow shelf off western Svalbard (Mangerud et al. 1992; Svendsen et al. 1992, 1996). A recent compilation of the western Svalbard land and
shelf stratigraphic record, together with ice-sheet modelling (Landvik et al. 1998), suggests that the Barents Sea Ice Sheet extended to the continental shelf edge, and that ice-free areas along the west coast were restricted to nunataks protruding above the >800-m-thick ice sheet. A recent study from Amsterdamøya, northwest Svalbard, $^{10}$Be exposure age dating shows that the Late Weichselian glaciers covered the coastal areas, whereas the mountain plateau >300 m a.s.l. had been ice-free for >80 000 years (Landvik et al. 2003).

We propose that these apparently contradictory field data from the terrestrial and marine environments may be reconciled. Our hypothesis is that the rather complicated geologic record found along the west coast of Svalbard reflects different physical and basal conditions in the coastal flanks of the ice sheet during the glacial cycles. The present paper discusses this hypothesis in the light of both the existing data set, and new data from the cross-shelf trough and shelf off Kongsfjorden (Fig. 1).

Previous studies

The land record
On land, well-dated Late Weichselian glacier expansions have been recorded from two areas along the west
coast. At the south coast of Bellsund (Landvik et al. 1992) and in Linnédale (Lønne & Mangerud 1991; Mangerud et al. 1992), till fabric indicates glacier flow controlled by local topography, where expanding valley and fjord glaciers entered the main fjords of Bellsund and Isfjorden. The Linnédalen valley was deglaciated before 12.3 ka BP (Mangerud & Svendsen 1990), whereas sediments above till on the south coast of Bellsund indicate deglaciation by 12.8–12.6 ka BP (Landvik et al. 1992). A well-preserved beach terrace at 87 m a.s.l. on the eastern slope of the Linnédalen valley was discussed by Mangerud et al. (1987, 1992). The terrace is capped by a thin diamicton, interpreted as a subglacial till, deposited by a glacier that moved down the Linnédalen valley. Luminescence dates of the beach sediment were 40 to 60 ka, and radiocarbon dates on shells were c. 36 ka BP (Mangerud et al. 1998). However, amino acid ratios of associated shells overlap with those of early Holocene samples. Based on the inferred digenetic temperature, Mangerud et al. (1992) calculated the duration of the Late Weichselian ice cover to have been 3000–6000 years at this site.

Multiple generations of raised beach sequences on Svalbard, with pre-Late Weichselian beach terraces occurring higher than the postglacial marine limit, have been described from Prins Karls Forland (Fig. 1) (Salvigsen 1977; Salvigsen & Nydal 1981; Forman 1990; Andersson et al. 1999, 2000; Andersson 2000), along the east coast of Forlandsundet (Forman 1989), Brøggerhalvoya (Miller 1982; Forman & Miller 1984; Tolgensbakk & Sollid 1987) and Kongsfjordsletta (Lehman & Forman 1992). It has been suggested that the presence of glacially undisturbed pre-Late Weichselian raised beaches indicates that the landforms were not covered by glaciers during the LGM (e.g. Salvigsen 1977; Forman 1989). However, Landvik et al. (1998) argued that preserved pre-Late Weichselian beach terraces alone cannot prove non-glaciation, since observations of morphologically well-preserved terraces overrun by fjord and valley glaciers have been presented from Kongsfjordsletta (Lehman & Forman 1992) and Linnédalen (Mangerud et al. 1987, 1992, 1998).

Morphologic and stratigraphic studies in the Kongsfjorden area (Fig. 1) (Forman 1989; Lehman & Forman 1992; Houmark-Nielsen & Funder 1999) suggested that the Late Weichselian glaciation was primarily confined to the Kongsfjorden trough. Lehman & Forman (1992) suggested that Kongsfjorden might have drained a Late Weichselian ice cap on the central part of Spitsbergen that was contiguous with the Late Weichselian Barents Sea Ice Sheet. Forman (1989) studying Late Weichselian glacier expansion in the Forlandsundet area, between Kongsfjorden and Isfjorden, suggested that the Late Weichselian glaciation was primarily a local event, with fjord and cirque glaciers expanding only a few kilometres compared to their neoglacial limits.

In a recent study, Andersson et al. (1999) showed that local glaciers on central Prins Karls Forland (Fig. 1) extended into Forlandsundet at the same time as the formation of the 34–36 m a.s.l. marine limit dated to 11.4 ka BP. They did not find evidence of regional glacial overriding, and suggested that the area remained largely ice-free during the Late Weichselian. Similarly, Andersson (2000) suggested a limited Late Weichselian expansion of local glaciers in the McVitiepynten area, northern Prins Karls Forland. Palaeotemperature estimates derived from the low amino acid ratios of fossil shells from Poolepynten and McVitiepynten were taken to indicate that the area had been ice-free for a long period due to long exposure to sub-zero temperatures (Andersson et al. 1999, 2000).

The offshore record

The marine record related to the Late Weichselian ice extent comprises sediment cores from the fjords and cross-shelf troughs, seismostratigraphy of the middle and outer shelf, and studies of the trough mouth fans off the major fjords (see Landvik et al. 1998). Sediment core data from the Isfjorden trough revealed a stiff diamicton, interpreted as a till deposit, underlying Holocene marine sediments (Svendsen et al. 1992, 1996). Fossil shells from the diamicton yielded infinite radiocarbon ages and ages in the interval 12.6–12.4 ka BP, suggesting a grounded glacier in the Isfjorden cross-shelf trough during the Late Weichselian. Radiocarbon ages of foraminifera and molluscs above the till indicate initial ice retreat from the outer shelf prior to 15 ka BP (Elverhøi et al. 1995; Svendsen et al. 1996). Further south, sediment cores from the outer part of the Bellsund trough contain a diamicton, interpreted as a subglacially deposited till overlain by glaciomarine sediments, radiocarbon dated to between 16.4 and 13.0 ka BP (Cadam 1996). These ages suggest that an outlet glacier in the Bellsund trough expanded onto the shelf during the Late Weichselian, and that deglaciation of the outer part of the shelf west of Bellsund may have started as early as 16.4 ka BP. The clearest indications of Late Weichselian ice extending to the continental shelf edge are thick debris lobe deposits on the continental slope off the Isfjorden trough (Andersen et al. 1996). As shown from the Barents Sea shelf, such lobes have been interpreted to reflect high sediment discharge and release from a grounded glacier at the shelf margin (Laberg & Vorren 1995, 1996). Two radiocarbon dates of 19.2 ka BP from marine sediments below one of the debris lobes on the Isfjorden fan suggest that deposition of the uppermost part of the pro-glacial fan occurred during the Late Weichselian (Andersen et al. 1996).

Seismic profiles across the outer shelf between the Isfjorden and Bellsund cross-shelf troughs (Svendsen et al. 1992; Solheim et al. 1996) show that the upper units have been truncated by up to 150 m of erosion.
A ridge, inferred to represent ice-marginal deposition, occurs close to the shelf break (Solheim et al. 1996). There is no direct chronostratigraphic control on the age of the erosional episode and the ridge formation. However, both Solheim et al. (1996) and Landvik et al. (1998) proposed that these features were formed by a Late Weichselian glacier that extended to the shelf break. Off the northwest coast of Spitsbergen, on the inner shelf off Magdalenafjorden, in Smeerenburg-fjorden and Raudfjorden, as well as between Amsterdamøya and Danskøya (Fig. 1), submarine moraine ridges were deposited in front of fjord glaciers (Liestøl 1972). According to Salvigsen (1977) and Landvik et al. (1998) these moraines are of Late Weichselian age. They may represent recessional stages rather than the maximum extent of the Late Weichselian glaciation (Andersen 1981; Landvik et al. 1998).

Methods and data acquisition

Seismic stratigraphy

Seismic data were obtained by an airgun array of two 0.6 litre sleeve guns used as a source and a single channel streamer with hydrophones as a recorder. The array was towed 25–30 m behind RV ‘Jan Mayen’ at a water depth of about 4 m. A firing pressure of 130 to 140 bar and a shooting rate of 10 s were used. The values for the trigger offset varied between 0.1 and 0.7 ms. The records revealed a penetration of 0.5 s TWT (Two-Way Travel Time). The signal-to-noise ratio was very good, and the spectral analysis of the seismic records showed that the frequency range of the sleeve gun array was from 30 to 500 Hz. A vertical resolution of less than 3 m can be achieved.

Sediment cores

The marine sediment cores were recovered during three different cruises (see Table 2). In all cases, sediment coring targets were identified from associated acoustic surveys of the sub-bottom stratigraphy. Cores were split and X-rayed, and lithological descriptions were prepared in the laboratory. Main grain size fractions were determined by wet sieving on 63 μm, 100 μm, 500 μm, 1 mm and 2 mm screens. Undrained shear strength on the JM99 cores was determined by fall cone test in the laboratory.

Radiocarbon dating

Age control (Table 3) for ‘NP87’ and ‘NP90’ cores is provided by AMS 14C dating of benthic foraminifera. Monospecific measurements were performed on local abundance maxima of Cibicides lobatulus and Nonion labradoricum. In two cases, the basal ages from cores NP87-136-GC2 and NP90-11-PC3, low abundances did not permit monospecific measurement and a mixed benthic assemblage was dated. Foraminiferal samples were graphitized at the Woods Hole Oceanographic Institution (WHOI) and 14C measurement was performed at the US-NSF AMS Facility of the University of Arizona.

Age control for ‘JM99’ cores is provided whenever possible by AMS 14C dating of articulated mollusc valves. These samples were graphitized at INSTAAR’s Laboratory for AMS Radiocarbon Preparation and

**Table 1. Seismic lines presented in this study.**

<table>
<thead>
<tr>
<th>Line</th>
<th>Start Latitude (N)</th>
<th>Start Longitude (E)</th>
<th>Stop Latitude (N)</th>
<th>Stop Longitude (E)</th>
</tr>
</thead>
<tbody>
<tr>
<td>JM99-1</td>
<td>78°29.969'</td>
<td>10°52.272'</td>
<td>78°27.918'</td>
<td>09°16.264'</td>
</tr>
<tr>
<td>JM99-2</td>
<td>78°41.912'</td>
<td>09°09.141'</td>
<td>78°49.021'</td>
<td>10°17.451'</td>
</tr>
<tr>
<td>JM99-5</td>
<td>79°03.335'</td>
<td>11°29.911'</td>
<td>78°51.936'</td>
<td>07°36.896'</td>
</tr>
<tr>
<td>JM99-16 leg 1</td>
<td>79°02.12'</td>
<td>09°30.08'</td>
<td>78°40.19'</td>
<td>09°30.08'</td>
</tr>
<tr>
<td>JM99-16 leg 2</td>
<td>79°30.12'</td>
<td>10°02.15'</td>
<td>78°17.97'</td>
<td>10°02.15'</td>
</tr>
</tbody>
</table>

**Table 2. Sediment cores. GC = gravity corer; PC = piston corer.**

<table>
<thead>
<tr>
<th>Core no.</th>
<th>Latitude (N)</th>
<th>Longitude (E)</th>
<th>Water depth</th>
<th>Core length</th>
<th>Corer type and diameter</th>
<th>Cruise</th>
</tr>
</thead>
<tbody>
<tr>
<td>NP87-136-GC2</td>
<td>79°01.06'</td>
<td>11°06.74'</td>
<td>279</td>
<td>475</td>
<td>GC 11 cm</td>
<td>‘Lance’ 1987</td>
</tr>
<tr>
<td>NP90-9-PC3</td>
<td>79°01.32'</td>
<td>11°06.24'</td>
<td>276</td>
<td>520</td>
<td>PC 11 cm</td>
<td>‘Håkon Mosby’ 1990</td>
</tr>
<tr>
<td>NP90-11-PC3</td>
<td>78°59.25'</td>
<td>10°07.36'</td>
<td>264</td>
<td>430</td>
<td>PC 11 cm</td>
<td>‘Håkon Mosby’ 1990</td>
</tr>
<tr>
<td>JM99-583</td>
<td>78°43.18'</td>
<td>09°21.26'</td>
<td>308</td>
<td>75</td>
<td>GC 11 cm</td>
<td>‘Jan Mayen’ 1999</td>
</tr>
<tr>
<td>JM99-591</td>
<td>78°58.59'</td>
<td>09°51.89'</td>
<td>239</td>
<td>105</td>
<td>GC 11 cm</td>
<td>‘Jan Mayen’ 1999</td>
</tr>
<tr>
<td>JM99-592</td>
<td>79°00.16'</td>
<td>10°24.32'</td>
<td>266</td>
<td>285</td>
<td>GC 11 cm</td>
<td>‘Jan Mayen’ 1999</td>
</tr>
</tbody>
</table>
Research at the University of Colorado and 14C measurement was performed at the National Ocean Sciences AMS Facility at WHOI.

All ages used in this paper are in radiocarbon years. In all cases, 440 years was subtracted from the laboratory-reported age to account for the local surface water reservoir age (Mangerud & Gulliksen 1975) (Table 3). However, studies from western Norway suggest (Bondevik et al. 1999) that the reservoir age is 440 years lower in the North Atlantic water periodically was several hundred years higher during the Late Weichselian.

Sea-floor morphology and seismic stratigraphy

The Kongsfjordrenna cross-shelf trough

The seismic line JM99-5 (Fig. 2A) along the Kongsfjordrenna cross-shelf trough (Fig. 1) shows a thickening of glaciogenic sediments from a few ms (TWT) at the mid-shelf (15–25 km from the coast) to approximately 100 ms TWT at the outer shelf. An upper unit of glaciogenic sediments shows several strong internal reflectors and is separated from the lower acoustic bedrock unit by a major unconformity (Fig. 2A). The acoustic energy of the sleeve gun array did not penetrate significantly into the bedrock underlying the glacial deposits. Based on the internal reflectors, the glacial deposits can be divided into several subunits, indicating repeated glacier advances to the outer shelf (Fig. 2A).

The morphologic character of the uppermost glaciogenic seismic unit suggests four moraine ridges along the section (I to IV on Fig. 2A). The prominent moraine ridge (I) at the mouth of Kongsfjorden has an elevation of approximately 200 ms TWT and can be correlated to an ice-marginal position mapped on land by Lehman & Forman (1992). On the middle and outer shelf, the moraine ridges are characterized by more gentle proximal and steeper distal slopes (Fig. 2A), and exhibit elevations of less than 100 ms TWT. Based on the speed of sound in water of 1500 m/s, the moraine ridges reach elevations between 75 and 150 m (100–200 ms TWT) above the surrounding bathymetry.

### Table 3. AMS radiocarbon dates from the sediment cores; see text for details.

<table>
<thead>
<tr>
<th>Core no.</th>
<th>Depth (cm)</th>
<th>Sample</th>
<th>Laboratory ID</th>
<th>AMS result no.</th>
<th>Species</th>
<th>Radiocarbon age</th>
<th>Reservoir corrected age</th>
</tr>
</thead>
<tbody>
<tr>
<td>JM99-583</td>
<td>27–29 B</td>
<td>NSRL-11174</td>
<td>CURL-4229</td>
<td>Abraded fragm. Mya truncata or Hiattella Nonion labradoricum</td>
<td>&gt;48 600</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>49–51 B</td>
<td>NSRL-11175</td>
<td>CURL-4230</td>
<td>Paired mollusc</td>
<td>12 640 ± 50</td>
<td>12 200 ± 50</td>
<td></td>
</tr>
<tr>
<td></td>
<td>56 B</td>
<td>NSRL-12339</td>
<td>CURL-5606</td>
<td>Paired mollusc</td>
<td>12 270 ± 65</td>
<td>11 830 ± 65</td>
<td></td>
</tr>
<tr>
<td></td>
<td>90 B</td>
<td>NSRL-11176</td>
<td>CURL-4231</td>
<td>Fragment</td>
<td>&gt;48 600</td>
<td></td>
<td></td>
</tr>
<tr>
<td>JM99-591</td>
<td>48–50 B</td>
<td>NSRL-11177</td>
<td>CURL-4232</td>
<td>Paired juvenile mollusc</td>
<td>11 000 ± 40</td>
<td>10 560 ± 40</td>
<td></td>
</tr>
<tr>
<td></td>
<td>71 B</td>
<td>NSRL-12340</td>
<td>CURL-5607</td>
<td>Shell fragment</td>
<td>&gt;44 300</td>
<td></td>
<td></td>
</tr>
<tr>
<td>JM99-592</td>
<td>18 B</td>
<td>NSRL-11178</td>
<td>CURL-4233</td>
<td>Single valve of Astarte sp.</td>
<td>810 ± 30</td>
<td>370 ± 30</td>
<td></td>
</tr>
<tr>
<td></td>
<td>55–59 B</td>
<td>NSRL-11179</td>
<td>CURL-4234</td>
<td>Paired Mya truncata</td>
<td>7 030 ± 25</td>
<td>6 590 ± 25</td>
<td></td>
</tr>
<tr>
<td></td>
<td>81–84 B</td>
<td>NSRL-11180</td>
<td>CURL-4235</td>
<td>Single Mya or Hiattella Nonion labradoricum</td>
<td>9 430 ± 50</td>
<td>8 990 ± 50</td>
<td></td>
</tr>
<tr>
<td></td>
<td>81–84 C</td>
<td>NSRL-11181</td>
<td>CURL-4236</td>
<td>Paired juvenile taxodont(?)</td>
<td>9 500 ± 60</td>
<td>9 060 ± 60</td>
<td></td>
</tr>
<tr>
<td></td>
<td>228 B</td>
<td>NSRL-12341</td>
<td>CURL-5608</td>
<td>Paired Portlandia</td>
<td>12 000 ± 60</td>
<td>11 560 ± 60</td>
<td></td>
</tr>
<tr>
<td></td>
<td>266 B</td>
<td>NSRL-12342</td>
<td>CURL-5609</td>
<td>Paired Portlandia</td>
<td>12 530 ± 65</td>
<td>12 090 ± 65</td>
<td></td>
</tr>
<tr>
<td></td>
<td>280 B</td>
<td>NSRL-12343</td>
<td>CURL-5610</td>
<td>Paired Maoma, juvenile</td>
<td>12 700 ± 65</td>
<td>12 260 ± 65</td>
<td></td>
</tr>
<tr>
<td>NP90-9-PC3</td>
<td>47–49</td>
<td>WHG-928</td>
<td>AA6888</td>
<td>Nonion labradoricum</td>
<td>8 825 ± 80</td>
<td>8 385 ± 80</td>
<td></td>
</tr>
<tr>
<td></td>
<td>103–105 B</td>
<td>WHG-931</td>
<td>AA6890</td>
<td>Nonion labradoricum</td>
<td>10 865 ± 100</td>
<td>10 425 ± 100</td>
<td></td>
</tr>
<tr>
<td></td>
<td>175–177 B</td>
<td>WHG-929</td>
<td>AA6899</td>
<td>Nonion labradoricum</td>
<td>12 260 ± 135</td>
<td>11 820 ± 135</td>
<td></td>
</tr>
<tr>
<td></td>
<td>279–281 B</td>
<td>WHG-932</td>
<td>AA6891</td>
<td>Nonion labradoricum</td>
<td>12 850 ± 100</td>
<td>12 410 ± 100</td>
<td></td>
</tr>
<tr>
<td></td>
<td>390–392</td>
<td>WHG-933</td>
<td>AA6893</td>
<td>Nonion labradoricum</td>
<td>13 190 ± 135</td>
<td>12 750 ± 135</td>
<td></td>
</tr>
<tr>
<td></td>
<td>443–450</td>
<td>WHG-941</td>
<td>AA6895</td>
<td>Mixed benthic forams</td>
<td>13 960 ± 120</td>
<td>13 520 ± 120</td>
<td></td>
</tr>
<tr>
<td>NP90-11-PC3</td>
<td>15–17</td>
<td>WHG-934</td>
<td>AA6894</td>
<td>Nonion labradoricum</td>
<td>8 190 ± 75</td>
<td>7 750 ± 75</td>
<td></td>
</tr>
<tr>
<td></td>
<td>39–41</td>
<td>WHG-943</td>
<td>AA6896</td>
<td>Nonion labradoricum</td>
<td>10 670 ± 85</td>
<td>10 230 ± 85</td>
<td></td>
</tr>
<tr>
<td></td>
<td>63–65</td>
<td>WHG-944</td>
<td>AA6938</td>
<td>Nonion labradoricum</td>
<td>11 850 ± 105</td>
<td>11 445 ± 105</td>
<td></td>
</tr>
<tr>
<td></td>
<td>87–89</td>
<td>WHG-945</td>
<td>AA6939</td>
<td>Nonion labradoricum</td>
<td>12 765 ± 105</td>
<td>12 325 ± 105</td>
<td></td>
</tr>
<tr>
<td></td>
<td>119–121</td>
<td>WHG-946</td>
<td>AA6940</td>
<td>Nonion labradoricum</td>
<td>12 880 ± 115</td>
<td>12 440 ± 115</td>
<td></td>
</tr>
<tr>
<td>NP87-136-GC2</td>
<td>47–49</td>
<td>WHG-345</td>
<td>AA4010</td>
<td>Cibicides lobatus</td>
<td>1 570 ± 50</td>
<td>1 130 ± 50</td>
<td></td>
</tr>
<tr>
<td></td>
<td>147–149</td>
<td>WHG-342</td>
<td>AA4189</td>
<td>Cibicides lobatus</td>
<td>4 500 ± 60</td>
<td>4 060 ± 60</td>
<td></td>
</tr>
<tr>
<td></td>
<td>178–180</td>
<td>WHG-574</td>
<td>AA4750</td>
<td>Nonion labradoricum</td>
<td>5 320 ± 60</td>
<td>4 880 ± 60</td>
<td></td>
</tr>
<tr>
<td></td>
<td>187–189</td>
<td>WHG-343</td>
<td>AA4187</td>
<td>Cibicides lobatus</td>
<td>7 380 ± 120</td>
<td>6 940 ± 120</td>
<td></td>
</tr>
<tr>
<td></td>
<td>202–204</td>
<td>WHG-515</td>
<td>AA4635</td>
<td>Nonion labradoricum</td>
<td>9 440 ± 100</td>
<td>9 000 ± 100</td>
<td></td>
</tr>
<tr>
<td></td>
<td>235–237</td>
<td>WHG-298</td>
<td>AA7298</td>
<td>Nonion labradoricum</td>
<td>10 510 ± 80</td>
<td>10 070 ± 80</td>
<td></td>
</tr>
<tr>
<td></td>
<td>315–316</td>
<td>WHG-344</td>
<td>AA4009</td>
<td>Nonion labradoricum</td>
<td>10 985 ± 120</td>
<td>10 545 ± 120</td>
<td></td>
</tr>
<tr>
<td></td>
<td>379–381</td>
<td>WHG-299</td>
<td>AA7299</td>
<td>Nonion labradoricum</td>
<td>12 190 ± 90</td>
<td>11 750 ± 90</td>
<td></td>
</tr>
<tr>
<td></td>
<td>444–446</td>
<td>WHG-630</td>
<td>AA4970</td>
<td>Mixed benthic forams</td>
<td>12 640 ± 150</td>
<td>12 200 ± 150</td>
<td></td>
</tr>
</tbody>
</table>
Fig. 2. All east–west running seismic lines. For location, see Fig. 1. A. Seismic section along the Kongsfjorden trough with interpretation of the outer moraine complex. The studied sediment cores (Fig. 5) are located, and the main moraine stages are given roman numerals I–IV. B. Seismic section west of Prins Karls Forland showing a possible slide of a glacigenic sediment unit at the outer shelf. Sediment core JM99-583 was recovered from the top of the slide block. Fmc = The Forlandet moraine complex. C. Seismic section across the shelf west off Prins Karls Forland showing various iceberg plough marks cutting into dipping seismic reflectors of a prograding shelf sequence. The outer shelf morphology may point towards a moraine ridge.
The shelf between Kongsfjorden and Isfjorden

The sea floor west of Prins Karls Forland (Fig. 3) exhibits an irregular bathymetry. A distinct coast-parallel ridge complex occurs on the outer shelf, 20–30 km offshore. This c. 1 km wide complex can be mapped for more than 50 km, and the main ridge reaches 50 m above the adjacent sea floor. Smaller subparallel ridges are mapped both west and east of the main complex. Large closed depressions, 6–10 km in size and up to 100 m deep, are found east of the main ridge complex. Smaller depressions (<20 m deep) are found immediately east of some of the smaller ridges.

We interpret the ridge complex to be moraine ridges formed at the margin of a glacier covering the shelf west of Prins Karls Forland. The large depressions on the proximal side of the moraines are interpreted to be troughs formed by glacial erosion or glaciotectonic detachment of larger blocks of sediments. The seismic lines cross several of the moraines and associated troughs (Fig. 3). We informally refer to the set of ridges as ‘the Forlandet moraine complex’.

The seismic section JM99-2 (Fig. 2B) transects the northern part of the shelf west of Prins Karls Forland (Fig. 1). The glacigenic unit as a whole appears to be much thinner than in outer Kongsfjordrenna (Fig. 2A). The section crosses the Forlandet moraine complex (Fig. 2B) with its characteristic rugged surface. A pronounced ridge occurs at a water depth of c. 300 m in the western part of the section, where sediment core JM99-583 (Fig. 5) retrieved a firm, overconsolidated diamicton at only 0.7 m sediment depth. The seismic character of the ridge is similar to the distal flank of the moraine ridge (IV) at the mouth of the Kongsfjorden trough (Fig. 2A), and we interpret the ridge as a terminal moraine. Similar moraine ridges also exist at the

**Fig. 3.** High-resolution bathymetry of parts of the shelf west of Prins Karls Forland. The data were collected with an EM1002 multibeam echo-sounder by the Norwegian Hydrographic Survey. A. The data were gridded with 50 m cell size and presented as colour-shaded contour maps (2 m depth contours). Note the distinct Forlandet moraine complex associated with closed depressions on its proximal side. The curved contours at the western part of line 99-2 may indicate downslope sliding of a large part of the outer shelf, including the moraine ridge (Fig. 2B), in postglacial time. B. Shaded relief map of the same area with the main glacial geological features outlined.
mouth of the Isfjorden trough (Svendsen et al. 1996; Solheim et al. 1998) and at the shelf break south of Isfjorden.

Seismic section JM99-1 (Fig. 2C) shows bedrock with westward dipping clinoforms covered by a thin (<10 m) veneer of glacial and marine sediments above a major unconformity. The section crosses the Forlandet moraine complex (Fig. 3), and, as seen from the bathymetry, it shows a continuous series of trough and ridge morphology. The north–south trending troughs are more than 10 m deep and up to 2000 m wide. This outer shelf morphology is suggestive of either eroded bedrock surfaces or moraine ridges, but a specific interpretation remains inconclusive. However, 60 km further south, on the outer shelf off Isfjorden, there is evidence of a Late Weichselian moraine ridge (Svendsen et al. 1996).

The adjacent seismic sections JM99-16 and JM99-17 (Fig. 4) permit mapping of the extent of the uppermost glacigenic unit from the Kongsfjorden cross-shelf trough, where it was cored and dated (Fig. 5), onto the shelf west of Prins Karls Forland, and further south to the Isfjorden cross-shelf trough (Fig. 1). The unit shows a clear north–south asymmetry. It varies in thickness from a few ms TWT in the central and northern part of the Kongsfjorden trough to about 80 ms TWT in the southern part (Fig. 4). A maximum thickness of approximately 100 ms TWT is reached on the shallow bank northwest of Prins Karls Forland, south of the trough margin (Fig. 4A). Further south, the unit becomes thinner and finally wedges out. However, the seismic line JM99-17 shows one faint sub-seafloor reflector that may represent the base of the upper glacigenic unit.

This unit most likely correlates to the thick wedge of glacigenic sediments off Isfjorden (Svendsen et al. 1996) (Fig. 1). Its thickness exceeds more than 250 m near the shelf edge. At the shelf edge, Svendsen et al. identified a Late Weichselian moraine ridge. Radiocarbon dates at the base of glaciomarine sediments in a

---

**Fig. 4.** Two north–south adjacent seismic lines along the shelf west of Prins Karls Forland from the Isfjorden trough to the Kongsfjorden trough.
Fig. 5. Lithological logs of cores located in Fig. 1. Unit numbers related to each core, black lines indicate correlations. For details on radiocarbon dates, see Table 3.
distinct depression about 10 km from the shelf edge indicate a deglaciation immediately prior to 14.5–14 ka (Svendsen et al. 1996).

Lithostratigraphy

Five sediment cores from the Kongsfjorden trough and one from the shelf west of Prins Karls Forland were investigated in this study (Figs 1, 5). The sediments can be divided into five lithofacies that occur in a consistent stratigraphic order in the different cores. Facies thicknesses can be seen from Fig. 5. From the base, these facies are:

Facies A: Matrix-supported diamicton

A well-consolidated diamicton was recovered in four cores (Fig. 5). It has a sandy to silty matrix and a high content of large clasts (Fig. 6A, C, D). It also contains abraded shell fragments that have an infinite radiocarbon age. This lithofacies exhibits a sharp contact to overlying crudely stratified or laminated sediments (Fig. 6D). The shear strength is high in core

Fig. 6. Selected X-radiographs of main sediment facies recovered in the cores. Unit numbers refer to Fig. 5 and the text. A. Subglacial till (core JM99-583 Unit 1). B. Glaciomarine sediment (Unit 2) overlain by bioturbated marine sand (Unit 3). C. Subglacial till (Unit 1a) overlain by glaciomarine diamicton (Unit 1b, crudely stratified glaciomarine sediment (Units 2 and 3). D. Contact between subglacial till (Unit 1) and reddish grey laminated mud (Unit 2).
JM 99-583 from the shelf off Prins Karls Forland, and lower in the cores from the Kongsfjordrenna trough (Fig. 5). In all the cores, the diamicton is interpreted as a subglacial till deposited by a grounded glacier.

Facies B: Crudely stratified diamicton
This facies exhibits a lithological composition similar to facies A, but the matrix is crudely stratified (Fig. 6C) and the diamicton shows lower shear strength (Fig. 5). This sediment facies is interpreted to be an ice-proximal glaciomarine deposit or sub-aqueous slump deposit derived from failure of older deposits of facies A.

Facies C: Laminated clay
Brownish to reddish, well to crudely laminated silty clay, void of gravel-sized clasts, occurs in four of the cores. The laminae thickness is 2–10 mm. The facies is interpreted as a glaciomarine sediment deposited during periods with high sediment supply and limited ice-rafting. The lack of IRD, which contrasts the other marine facies, may indicate periods of sea ice cover.

Facies D: Laminated and massive clayey silt
This is a laminated to massive clayey silt facies containing frequent gravel-sized clasts, interpreted as dropstones, as well as paired molluscs. The lamination varies from well developed (Fig. 6D) to crude (Fig. 6A), and is sometimes only visible on X-radiographs (Fig. 6C). Numerous radiocarbon dates on foraminifera and molluscs yield late-glacial and early Holocene ages (Fig. 5, Table 3). The sediment is inferred to have been deposited in a seasonally sea-ice-free glaciomarine environment influenced by relatively high sediment supply from suspension and from ice-rafting.

Facies E: Bioturbated sandy silt
This is massive sandy silt with numerous gravel-sized clasts interpreted as dropstones, paired molluscs and frequent polychaeta tubes. Facies E constitutes the uppermost unit of the cores. Sharp lower boundaries to underlying facies and distinctly younger radiocarbon ages (Fig. 5, Table 3) suggest that the facies formed underlying facies and distinctly younger radiocarbon ages (Fig. 5, Table 3) suggest that the facies formed.

Unit 1. – The lowermost 28 cm of the core constitutes a massive, very compact matrix-supported diamicton (facies A) with undrained shear strength exceeding 370 kPa (Fig. 5). The matrix comprises very dark-greyish brown silt, with a content of <10% angular to subrounded gravel-sized clasts (Fig. 6). An abraded shell fragment yielded a radiocarbon age of >48.6 ka BP (Table 3). The boundary to Unit 2 is sharp. From the massive unsorted character and high shear strength values, the diamicton is interpreted as a subglacial till.

Unit 2. – The compact diamicton is overlain by a 45-cm-thick less consolidated and crudely stratified dark-grey matrix-supported diamicton (facies B). It contains scattered subangular to subrounded and occasionally fractured and rounded clasts. The upper part of the unit, between c. 40 and 20 cm core depth, starts with an erosional lower boundary and fines upwards from silty sand with pebbles to sandy silt. Undrained shear strength averages 35 kPa (Fig. 5), assumed to be the normal consolidation for the sediment.

Paired molluscs were found at 56 and 50 cm b.s.f. yielding radiocarbon ages of 11.8 and 12.2 ka BP, respectively (Fig. 5, Table 3). The inverse age relationship suggests turbation of the sediment after deposition. Redeposition of older material is also indicated by an abraded shell fragment at 30 cm b.s.f. which yielded an age of >48.6 ka BP (Table 3). We suggest that this sediment unit was deposited in a glaciomarine environment influenced by subaqueous sediment flows.

Unit 3. – The uppermost 20 cm of the core comprises a sharply based crudely stratified sandy silt with numerous pebbles (facies E). The matrix colour is very dark greyish brown and compares to Unit 1. The sediment is bioturbated with abundant polychaeta tubes. From the lithological composition, we interpret the sediment to have been deposited by sediment flows, probably originating from the Unit 1 till exposed at shallower water depth and subsequently exposed to winnowing and bioturbation.

Sediment stratigraphy on the outer shelf

Core JM99-583
Core JM99-583 (Fig. 5) was retrieved from the moraine ridge on seismic line 99-2 (Fig. 2B) south of the Kongsfjorden trough (Figs 1, 2B). The corer penetrated to a depth of 94 cm b.s.f. (below sea floor) before it stopped in a very firm diamicton. Other coring attempts failed due to the firm sediments at the surface.

Sediment stratigraphy in the Kongsfjordrenna cross-shelf trough

The sediment cores in the Kongsfjordrenna cross-shelf trough were located along the seismic transect with the objective of penetrating the postglacial sediments and of sampling the diamicton associated with the moraines and glacial deposits interpreted from the seismic profiles (Figs 2A, 4A).

Core JM90-591
Core JM90-591 was retrieved from the proximal slope of the cross-trough ridge interpreted as a marginal moraine (III) on the seismic profile (Fig. 2A).
Penetration stopped in a firm diamicton; total core recovery was 100 cm.

**Unit 1.** – At this site, Unit 1 is subdivided into two subunits. The lowermost Unit 1a is a massive diamicton (our facies A). It is clast-supported with a matrix of very dark greyish brown sandy silt (Fig. 5). The clasts are angular to subrounded and up to 3 cm in diameter. The diamicton shows undrained shear strength between 60 and 70 kPa. At 79 cm depth there is a sharp boundary to Unit 1b, a dark grey crudely stratified matrix-supported diamicton with more silt and less clast (facies B). Unit 1b is moderately firm, with a shear strength of 40–50 kPa. A thick abraded shell fragment was dated to >44.3 ka BP (Fig. 5, Table 3).

The massive appearance, lack of structures, and shear strength >60 kPa suggest that the lower part of the diamicton (Unit 1a) is a subglacial till. The upper crudely stratified part (Unit 1b) is interpreted as either a till deposit or diamicton redeposited from Unit 1a sediments during the deglaciation.

**Unit 2.** – A 15-cm-thick, faintly laminated clayey silt bed (facies D) has a sharp contact to the underlying Unit 1 diamicton. Several sand laminae and a 12-mm-thick lens of fine-grained sand occur in the upper part of the unit. The sediment has a dark grey to greyish brown colour and is void of gravel-sized clasts. Shear strength values are low (Fig. 5). The same sediment facies can be recognized in a similar stratigraphic position in all cores from the Kongsfjordrenna trough.

**Unit 3.** – Faintly laminated silty sand of facies D occurs over a sharp lower contact at 50 cm b.s.f. (Fig. 5). The sediment fines upwards to sandy silt. There are scattered gravel particles in the unit, more frequent in the lowermost 10 cm. A radiocarbon date on a paired mollusc shell close to the lower boundary yielded an age of 10.6 ka BP (Table 3). This unit is interpreted to be a glaciomarine sediment with the sand fraction as well as the gravel clasts being ice-rafted detritus.

**Unit 4.** – With a transitional boundary to the underlying sediment, the uppermost 9 cm of the core comprises bioturbated sandy silt with abundant gravel particles and polychaeta tubes (facies E).

**Core NP90-11-PC3**

The core recovered 460 cm of sediment from the upper western slope of the Kongsfjordjupet basin at 264 m water depth (Fig. 2A).

**Unit 1.** – The lowermost 280 cm comprises a homogeneous pebbly diamicton with a matrix of sandy silt (facies B). The clasts are interpreted to be dropstones in a glaciomarine sediment.

**Unit 2.** – The glaciomarine facies is overlain by a 40-cm-thick compact diamicton (facies A) interpreted as a subglacial till.

**Unit 3.** – The till is overlain by a 20-cm-thick reddish grey laminated mud with sharp lower and upper boundaries (facies C). A radiocarbon date indicates that deposition occurred around 12.4 ka BP (Table 3).

**Unit 4.** – Overlying the laminated mud is a more greyish, partly sulphide-stained, silt showing a distinct fining upward (facies D). An important marker horizon is a red silt bed (facies C) found between 50 and 60 cm depth that can be used to correlate to the cores further east (Fig. 5). A series of three radiocarbon dates indicates deposition between 12.3 and 10.2 ka BP (Table 3) at the top of the unit (Fig. 5).

**Unit 5.** – The top sediment is olive grey sandy silt (facies E). The olive colour indicates elevated carbon content suggestive of seasonally open waters. A single radiocarbon date of 7.7 ka BP (Table 3) confirms deposition during the Holocene.

**Core JM99-592**

Core JM99-592 is also located at the western slope of the Kongsfjordjupet basin (Fig. 2) at 266 m water depth. The core recovery was 285 cm, but the corer did not reach the underlying firm diamicton that was sampled in both cores JM99-591 and NP90-11. However, the basal radiocarbon date of c. 12.3 ka BP (Table 3) suggests that the core contains most of the sediment succession above the till.

**Unit 1.** – The lower part of the core (285–40 cm depth) comprises faintly laminated dark grey clayey silt (facies D). Scattered pebble clasts and poorly defined zones with a slightly higher pebble concentration occur. Paired mollusc shells of *Portlandia arctica* and juvenile *Macoma calcarea* occur randomly in the unit. Between 180 and 173 cm the silt is more fine-grained than in the rest of the unit, massive to crudely stratified, with a brownish grey colour (facies C). This interval also contains distinctly fewer pebbles compared to the rest of the unit. A series of six radiocarbon dates brackets the sedimentation between c. 12.3 and 6.6 ka BP (Fig. 5, Table 3), and records a distinct decrease in sedimentation rate after c. 10 ka BP.

We interpret the sediment to have been deposited in a glaciomarine environment on the lower ice-proximal slope of the moraine ridges. The core did not penetrate any underlying till, but the fine-grained sediments at the base, the arctic molluscs and the radiocarbon age of 12.3 ka BP suggest that deposition started shortly after the deglaciation.

**Unit 2.** – The uppermost 40 cm comprises fine-grained silt with some sand laminae and a high concentration of gravel particles at the top (facies E). The lower contact is sharp. The uppermost 16 cm exhibits frequent polychaeta tubes. A radiocarbon measurement on a single valve of *Astarte* sp. at 18 cm depth yielded an age of 370 ± 30 BP (Table 3) and suggests that the lower
boundary represents a hiatus spanning almost 6000 years.

Core NP90-9-PC3
Two cores from the station show the same stratigraphy. A total of 520 cm sediment was recovered from 276 m water depth.

Unit 1. – The cores stopped in a >80-cm-thick firm diamicton (facies A). The unit is homogeneous in character, and is interpreted as a subglacial till.

Unit 2. – A 50-cm-thick reddish grey laminated mud (facies C) overlies the till. A radiocarbon date on mixed foraminifera at the base of the unit yielded an age of 13.5 ka BP (Table 3). The relatively high basal age is supported by the series of dates in the overlying Unit 3 (Fig. 5).

Unit 3. – A 340-cm-thick succession of silt exhibits the same development as Unit 3 in core NP90-11. The sediment is a greyish mud with an increasing content of sulphide-stained silt towards the top (facies D). Between 125 and 150 cm there is a distinct bed of red silt (facies C) correlated to the other cores (Fig. 5). The series of four dates indicates that the whole unit was deposited before c. 10 ka BP (Fig. 5, Table 3).

Unit 4. – There is a sharp lower boundary to the uppermost sediments, which comprise more sandy silt (facies E). A radiocarbon date on Nonion labradoricum of 8.4 ka BP shows deposition during the Holocene.

Core NP87-136-GC2
Core NP87-136-GC2 was recovered from the same location as core NP90-9, and lithological correlation between the two cores provides a more complete sequence (Fig. 5).

Unit 1. – The lowermost 50 cm comprises reddish brownish silt (facies D) that can be differentiated from the overlying Unit 2 only by its colour. Both the lithology and the radiocarbon date of 12.2 ka suggest a correlation to Unit 3 in core NP 90-9 (Fig. 5).

Unit 2. – This unit comprises olive grey silt with occasional sulphide stains (facies D). At 340–360 cm depth it is interrupted by the characteristic reddish facies C bed that is correlated to the cores to the west (Fig. 5). The radiocarbon dates suggest deposition between 12 and 9 ka.

Unit 3. – The lower part (Unit 3a) is sandy silt with a sharp lower boundary (facies E). The sand content decreases upwards and grades into a greyish silt (facies D). A correlation to core NP90-9 Unit 4 is suggested by both the sediment lithology and the radiocarbon ages. We conclude that the unit was deposited under marine conditions similar to the present.

Correlations
The firm diamicton (facies A), which is interpreted as a subglacial till, is found in cores JM99-583, JM99-591, NP90-11 and NP90-9. The seismic profiles show that it represents a laterally continuous unit deposited in the central and inner parts of the Kongsfjordrenna trough. The overlying suite of glacimarine to marine sediments thickens from the moraine ridge III, close to core JM99-591 towards the Kongsfjorddjupet basin to the east, as shown by both the seismic profile (Fig. 2A) and the sediment cores (Fig. 5). Seismic line JM99-16 (see above) shows that this succession can be mapped southwards, and correlates with the sediment stratigraphy in core JM99-583 west of northern Prins Karls Forland. Most of the sediments above the diamicton were deposited between the deglaciation at 13 ka BP and the mid-Holocene. Records from the later part of the Holocene have generally been lost in a hiatus close to the top of the cores.

A 10 to 30-cm-thick brownish to reddish grey clayey silt void of gravel clasts (facies C) has been recognized in all cores in the trough, at increasing sediment depth towards the east (Fig. 5). The bed is easily recognizable, and constitutes a marker horizon in the Kongsfjordrenna area. Radiocarbon dates from several cores bracket its deposition between 11.5 and 10.5 ka BP, probably closer to the younger age (Fig. 5). The unit is interpreted to represent a sudden sedimentary event. It may correlate to a reddish silt reported from raised sections on the south coast of Kongsfjorden, where it was deposited prior to 10.4 ka BP and interpreted as deposited from tidewater glacier-fed sediment plumes containing sediments from local Devonian red sandstones (Lehman & Forman 1992).

Age of the last deglaciation
The oldest radiocarbon ages from sediments above the glacial diamicton were obtained from core NP90-9 (Fig. 5) retrieved from the eastern flank of the Kongsfjorddjupet basin (Fig. 2A). The dates of 13.5 and 12.75 ka BP are older than any ages obtained from the cores further west, and show that the deglaciation of the whole Kongsfjordrenna trough was completed by c. 13 ka BP. This is in agreement with a deglaciation c. 13 ka BP of the Kongsfjorden basin east of the moraine ridge I, as Lehman & Forman (1992) proposed from their studies of raised marine sediments.

Discussion
In the following, we conclude that grounded glacial ice reached the shelf break at the mouth of the Kongsfjordrenna trough, and, significantly, the adjacent shelf west of Prins Karls Forland. As discussed above, this conclusion needs to be reconciled with terrestrial
studies from the Kongsfjorden and Forlandsundet region, suggesting that land areas adjacent to Kongsfjorden were ice-free during the Late Weichselian (see review above). Generally, these studies propose limited ice-sheet extent based on morphologically preserved raised beaches of pre Late Weichselian age, and lack of evidence for glacial deposition or erosion during the Late Weichselian. Thus, the results obtained from the land and marine records have been regarded as contradictory (Mangerud & Svendsen 1992; Mangerud et al. 1992; Svendsen et al. 1996; Landvik et al. 1998; Andersson et al. 1999, 2000; Houmark-Nielsen & Funder 1999). The following discussion focuses on understanding the dynamics of the ice sheet from the marine record presented in this paper and the available terrestrial studies.

The Late Weichselian ice extent

Pre-Late Weichselian raised beach sediments at Kvadehuskletta (Fig. 1) were studied by Miller (1982) and Forman & Miller (1984). Based on the morphological expression of the sediment surface of the raised beaches, pedogenesis of soil profiles, age determinations, and the lack of overlying glacial deposits, they concluded that the sediments had not been overridden by Late Weichselian glaciers. This view was supported by Lehman & Forman (1992), who proposed a minimum reconstruction where the Late Weichselian glaciers at least reached the prominent submarine moraine ridge at the mouth of Kongsfjorden (our moraine ridge no. 1, Fig. 2A) and its lateral moraine on the northern shore of the fjord. Radiocarbon dates from a sediment section proximal to the moraine show that ice retreat was underway by 12.3 ka BP. Based on a review of all available data from the west coast of Spitsbergen, including the unpublished results from cores NP90-9 and NP90-11 presented in this study, Landvik et al. (1998) suggested that the Late Weichselian glacier filled the Kongsfjordrenna trough. A similar westward ice extent was supported by Houmark-Nielsen & Funder (1999). However, they suggested that an outlet glacier was confined to the deeper parts of the trough, a reconstruction that was based on the magnitude of the isostatic rebound along Kongsfjorden (Lehman & Forman 1992) and the adjacent ice-free areas concluded from other studies (Forman & Miller 1984; Lehman & Forman 1992; Andersson et al. 1999). In a recent analysis of large-scale bedforms, Ottesen et al. (2005) proposed that the entire Norwegian–Svalbard margin was glaciated, and that ice streams followed the major cross-shelf troughs.

The seismic data and seafloor morphology presented in this study suggest that glacier ice reached the shelf break off the Kongsfjordrenna trough several times during the Quaternary. The deposition of over-consolidated diamictons and the subsequent onset of marine sedimentation at 12–13 ka BP show that the trough was filled by ice during the Late Weichselian. As shown by seismic line 99-16 (Fig. 4) and core JM99-583, the glacier was not only confined to the cross-shelf trough, but probably also extended to the outer continental shelf west of Prins Karls Forland, where subglacial till was deposited. We interpret the deposition of this till to have occurred during the same glacial event that caused the formation of the westermost coast-parallel moraine ridges. Thus, the whole shelf was glaciated prior to a brief retreat to the large Forlandet moraine complex (Fig. 3). The orientation of the coast-parallel moraines indicates that ice movement across the shelf from Prins Karls Forland, and not fed by ice in the Kongsfjorden or Isfjorden cross-shelf troughs.

The Late Weichselian ice extent reconstructed here can be related to the extent and retreat of grounded ice in the Isfjorden cross-shelf trough and the adjacent banks (Fig. 1). Svendsen et al. (1992, 1996) showed ice extent to the shelf break. A correlation of this advance to the ridge mapped along the shelf break south of the Isfjorden trough has been proposed (Solheim et al. 1996; Landvik et al. 1998). The ice retreat in the Isfjorden trough started at c. 15 ka, and a minor readvance at the transition between the shelf and the fjord occurred c. 12.4 ka (Svendsen et al. 1996).

Coastal glaciation dynamics

We see two possible scenarios for the configuration of the Late Weichselian ice margin along the west coast of Spitsbergen:

1. Glacial ice filled the Isfjorden and Kongsfjordrenna cross-shelf troughs as low gradient outlet glaciers, while parts of the present coastal areas of Prins Karls Forland and east coast of Forlandsundet were ice-free (cf. Andersson et al. 1999, 2000; Houmark-Nielsen & Funder 1999).

2. Ice streams drained the fjords and cross-shelf troughs off Kongsfjorden and Isfjorden, and glaciers with less active basal movement covered the intermediate areas. Glacier ice extended to the shelf edge west of Prins Karls Forland (cf. Landvik et al. 1998), but we suggest that there were large lateral differences in ice thicknesses and glacier dynamics.

Our study shows that glacier ice filled the Kongsfjorden trough and the northern part of the bank west of Prins Karls Forland during the Late Weichselian. As shown by seismic sections JM99-16 and JM99-17 (Fig. 4), grounded ice extended southwards to the Isfjorden trough, as suggested by eroded, undulating bedrock underlying the glacial deposits, and thus correlate to the units shown by Svendsen et al. (1996). The ice was not therefore confined to the Isfjordrenna and Kongsfjordrenna troughs only, but also inundated
the shallow banks between the two fjord systems. Even if this glacier was fed from the mountains of Prins Karls Forland, the scenario leaves little room for ice-free conditions along the coasts of Forlandsundet (Forman 1989; Andersson et al. 1999, 2000) and Brøggerhalvoya (Forman & Miller 1984; Miller et al. 1989), as advocated by Houmark-Nielsen & Funder (1999) and Andersson et al. (2000).

The possibility of ice-free areas on land may be evaluated by reconstructing ice surface profiles along the Kongsfjorden–Kongsfjordrenna trough (Fig. 7). The moraine ridge on the north shore of Kongsfjorden (Lehman & Forman 1992; Houmark-Nielsen & Funder 1999) reaches an elevation of more than 150 m a.s.l., whereas a veneer of erratics can be found up to 200 m a.s.l. (Houmark-Nielsen & Funder 1999). Based on its trend, Lehman & Forman (1992) correlated the moraine system to the submarine moraine at the mouth of Kongsfjorden, which they dated to \( \sim 13 \) ka. As shown by our data, this is a recessional stage during deglaciation, and hence the highest point of the moraine represents the lowest possible elevation of an ice surface during the Late Weichselian. In our calculation, we have compensated for a glacial isostatic uplift of 45 m (Lehman & Forman 1992) at Kongsfjorden. Based on the gradient of the onshore moraine (Lehman & Forman 1992), we assume the glacial isostatic effect to have been insignificant at moraine IV. With a grounded glacier at 250 m water depth at moraine IV (Fig. 2A), a minimum height of the glacier terminus at LGM is obtained by correcting for a sea level lowering of 135 m (Lambeck et al. 2000; Yokoyama et al. 2000) and adding 10% of the glacier thickness to keep it grounded. However, if the glacier rested at moraine IV until 15 ka, it must have had a thickness which could withstand a sea level rise to \(-110\) m (Yokoyama et al. 2000), resulting in a minimum average surface gradient of \(2.9\) m/km (Fig. 7). Even such a low gradient glacier would cover the suggested non-glaciated sites at both Kvadehuksletta and northern Prins Karls Forland (Fig. 7).

Modern ice streams and some tidewater glaciers are characterized by low surface gradients. Storstrommen in northeast Greenland is grounded in a fjord basin similar to the Kongsfjorden trough, and exhibits a present surface gradient of \(4.5\) m/km (Funder et al. 1998). The surface gradients of the grounded parts of modern Antarctic ice streams vary considerably, depending primarily on subglacial topography and basal conditions. Surface gradients of \(\geq 5\) m/km are common, but may be as low as 2 m/km (Bentley 1987).

Recent reconstructions of palaeo-ice streams from the geological record (see Stokes & Clark 2001; Clark et al. 2003) have revealed an even larger variety of settings than shown by the modern examples. From the east coast of Baffin Island, Kaplan et al. (2001) calculated from geomorphological evidence that the Cumberland sound ice stream had a surface gradient as low as 0.5 m/km, and calculated that it existed with maximum driving stresses as low as 2–7 kPa. Socha et al. (1999) reconstructed an average ice surface slope of about 2 m/km for the Lateglacial (13–11 ka BP), partly land-based Green Bay Lobe in eastern Wisconsin.
A surface gradient of only a few metres per km was also mapped for the Late Weichselian outlet glacier in Scoresby Sund, east Greenland (Funder et al. 1998). In Svalbard, Landvik et al. (2003) found that outlet glaciers with surface gradients 20–40 m/km left the higher parts of Amsterdamoya ice-free during the Late Weichselian.

We conclude that the Kongsfjordrenna trough was occupied by an ice stream in the sense of ‘a region in a grounded ice sheet in which the ice flows much faster than in regions on either side’ (Paterson 1994). As discussed above, our data suggest that glacier ice covered the shelf off Prins Karls Forland between the Kongsfjorden and Isfjorden cross-shelf troughs. The ice stream in Kongsfjordrenna was fed by ice drainage along the major fjord systems of Kongsfjorden and Krossfjorden, which must have drained a large portion of the ice cap/ice culmination over northwest Spitsbergen. A high ice discharge along this pathway trough time is also suggested by the large fan-shaped deposit off the Kongsfjordrenna trough, mainly containing debris flow deposits released by glaciers reaching the shelf break (see, e.g., Vorren et al. 1998).

The geological record from the Kongsfjordrenna trough reflects an actively eroding and depositing fast-flowing glacier. The subglacial environment of such a glacier contrasts with the observations from Brøggerhalvøya (Forman & Miller 1984; Tolgensbakk & Sollid 1987; Miller et al. 1989) and northern Prins Karls Forland (Andersson et al. 1999, 2000), where pre-Late Weichselian sediments were left undisturbed by the last glaciation. The survival of sediments below ice sheets has often been attributed to frozen bed conditions. However, in a coastal environment such as western Svalbard, maintenance of frozen-bed conditions throughout both the growth and decay portions of a complete glacial cycle seem unlikely. From studies at the mouth of Isfjorden (Fig. 1), we know that duration of the last glacier advance beyond the coastline could have been as little as 3000 years if the ice was warm based (Mangerud et al. 1992). Given the significant differences in flow rates, the preservation of older sediments in the inter ice-stream areas along the coast may be attributed to short duration of active ice flow, rather than wide variations in basal thermal regime. Such a mechanism has been suggested to explain preservation of landforms close to the fast-flowing M’Clintoc Channel ice stream (Clark & Stokes 2001) and the Dubawnt Lake ice stream (Stokes & Clark 2003) in the Canadian Arctic.

Based on these considerations, we suggest that the LGM ice configuration on the shelf and in outer fjords between Kongsfjorden and Isfjorden was as shown in Fig. 8. Fast-flowing ice streams drained the fjords and the prominent cross-shelf troughs, as indicated by the deposition of glacial sediments in these areas. Pre-LGM stratigraphies, both with and without evidence of a Late Weichselian glaciation, may serve as lateral constraints on the zone of fast-flowing ice. Land areas in between the ice streams, proposed by a number of workers to have been ice-free during the LGM, were probably covered by glaciers and ice caps which were dynamically less active or existed for a shorter time than ice in the fjord-trough systems.

It is known that about 90% of ice drainage from the Greenland and Antarctic ice sheets is via fast-flowing outlet glaciers and ice streams, although their outlets represent only 13% of the Antarctic coastline and an even smaller proportion of the Greenland Ice Sheet (Paterson 1994; Knight 1999). We suggest the same was true for the Late Weichselian Barents Sea/Svalbard Ice Sheet complex, and that it primarily
drained through a series of ice streams in the major fjords and cross-shelf troughs. This has implications for our understanding of the ice sheet dimensions and dynamics: rather than being a concentric ice sheet flowing from an ice-divide over the central Barents Sea, as suggested by recent reconstructions, most of the ice discharge was channeled through ice streams, resulting in an ice-sheet configuration with a series of domes and local ice caps, with large local differences in both basal temperatures and ice flux.

Conclusions

- The study suggests a more complicated and topographically controlled configuration of Late Weichselian glacier ice along the west coast of Svalbard than suggested by previous reconstructions.
- The Kongsfjordrenna cross-shelf trough and the continental shelf off Prins Karls Forland were glaciated during the Late Weichselian.
- Fast-flowing ice streams filled the Kongsfjorden and Isfjorden troughs, whereas glacier ice in the area between was dynamically less active.
- Deglaciation of the ice stream was rapid, and the ice margin reached the coast of Kongsfjorden by 13 ka.
- The geomorphology and sediment records both offshore and onshore in this region provide excellent potential for the understanding of the dynamics of coastal glaciations and sedimentation through the glacial periods.
- We tentatively suggest that the Late Weichselian Barents Sea/Svalbard Ice Sheet was predominantly drained by a series of ice streams, and that the marginal areas between the major discharge troughs were characterized by thinner and dynamically less active glacier ice.
- The proposed concept implies the possibility of ice-free nunataks to have existed outside the areas of high ice discharge; a scenario compatible with, for example, the Antarctic Peninsula today.

Acknowledgements. – The cores were collected during Norwegian Polar Institute cruises with R/V Lance (1987), R/V Håkon Mosby (1990) and a University of Tromsø/UNIS cruise with R/V Jan Mayen (1999). In this last-mentioned cruise, students from the University of Tromsø and UNIS participated; Steinar Iversen supervised the seismic equipment and Louise Hansen contributed to onboard core descriptions. The bathymetric data are published with permission from the Norwegian Hydrographic Service (permission no. 421/02). The University Centre on Svalbard (UNIS) supported the study in numerous ways. Steven L. Forman, John T. Andrews and Michael Houmark-Nielsen reviewed the manuscript. We extend our thanks to all these persons and institutions.

References
