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Arctic Ocean deep-sea record of northern Eurasian ice sheet history

Robert F. Spielhagen[a,b,*], Karl-Heinz Baumann[c], Helmut Erlenkeuser[d], Norbert R. Nowaczyk[e], Niels Nørgaard-Pedersen[a,f], Christoph Vogt[c], Dominik Weiel[g]

[a] *Academy of Sciences and Literature Mainz, Geschwister-Scholl-Str. 2, D-55131 Mainz, Germany*
[b] *Leibniz Institute for Marine Sciences, East Coast Building, Wischhofstr. 1–3, D-24148 Kiel, Germany*
[c] *Department of Geosciences, University of Bremen, Klagenfurter Str., D-28359 Bremen, Germany*
[d] *Leibniz-Laboratory for Radiometric Dating and Stable Isotope Research, Christian-Albrechts-Universität, Olshausenstr. 40, D-24118 Kiel, Germany*
[e] *GeoForschungsZentrum Potsdam, Telegrafenberg Building C, D-14473 Potsdam, Germany*
[f] *Dalbyvej 59, DK-4623 Lille Skensved, Denmark*
[g] *Johannes-Meyer-Str. 13, D-49808 Lingen, Germany*

Abstract

The sediment composition of deep-sea cores from the central Arctic Ocean, the Fram Strait, and the Yermak Plateau was analyzed for several parameters to reconstruct the history of marine paleoenvironment and terrestrial glaciation in the last 200,000 years. Layers with high amounts of coarse, terrigenous ice-rafted debris (IRD) and often high contents of smectite were deposited during extensive glaciations in northern Eurasia, when ice sheets reached the northern continental margins of the Barents and Kara seas and discharged icebergs into the Arctic Ocean. Intercalated layers with relatively low IRD and smectite contents, but abundant planktic foraminifers in the coarse fraction were deposited during periods of Atlantic Water inflow to the Arctic Ocean and seasonally open waters (leads) in a sea ice cover with only few icebergs in the Arctic Ocean. High IRD contents in the sediments reflect the presence of ice sheets on the Kara and Barents seas shelves and the hinterland during the entire oxygen isotope stage 6 (ca 190–130 ka), in substage 5b (ca 90–80 ka), at the stage boundary 5/4 (around 75 ka), and in late stage 4/early stage 3 (ca 65–50 ka). These results are in excellent correlation with those from recent field work in northern Scandinavia, European Russia, Siberia, and on the shelves. Relatively low amounts of IRD in central Arctic Ocean sediments from the Late Weichselian glacial maximum (ca 24–18 ka) correlate well with the recent reconstruction of a very limited eastern ice sheet extension during this time.

Oxygen and carbon isotope records of planktic foraminifers from the analyzed sediment cores show a number of prominent excursions which can be interpreted as evidence for freshwater events in the Arctic Ocean. The synchroneity of freshwater events and IRD input suggests a common source. Strongest events were associated with deglaciations of the Barents and Kara seas after the ice sheets had blocked the outflow of large rivers for several millennia. The outflow of freshwater from large ice-dammed lakes occurred at ca 130, 80–75, and 52 ka. Freshwater events in the central Arctic Ocean during the last deglaciation (ca 18 ka) were relatively small compared to the previous events. This indicates that during most of the Late Weichselian glacial maximum a river outflow from northern Siberia to the Arctic Ocean was possible.

Atlantic Water inflow to the Arctic Ocean and seasonally open waters in the ice (leads) occurred during the interglacials of oxygen isotope stage 1 and substage 5e, during several interstadials (stage 3, substages 5a and 5c), and to a lesser degree within stadials and glacials (stages 2, 4, and 6). With the exception of the interglacials, these periods were times of strong ice growth on the continents as revealed by terrestrial data. The coincidence suggests that open waters in the Arctic Ocean and the Nordic Seas were an important moisture source (in addition to more southerly sources) which fostered the growth of ice sheets on northern Eurasia.

1. Introduction

In the last decade, a wealth of new data has come up, which radically changed our view of the history and variable configuration of Weichselian ice sheets in northernmost Europe and Asia. Since the 1970s, Grosswald and colleagues had pursued the view of a huge continental ice sheet during the Last Glacial Maximum (LGM) at ca 20,000 years ago (20 ka), covering entire Arctic Eurasia from Scandinavia in the west to Beringia in the far east (e.g., Hughes et al., 1977; Grosswald, 1980; Grosswald and Hughes, 2002). This ice sheet configuration and reduced versions of it have

*Corresponding author. GEOMAR Research Center for Marine Geosciences, Wischhofstr. 1–3, Kiel D-24148, Germany. Tel.: +49-431-600-2855; fax: +49-431-600-2941.

E-mail address: rspielhagen@geomar.de (R.F. Spielhagen).

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been used widely as boundary conditions for paleoenvironmental models of the LGM (e.g., Manabe and Broccoli, 1985; Kutzbach and Guetter, 1986; Peltier, 1994). Several Russian scientists disagreed with Grosswald and proposed a very limited LGM ice extent in northern Russia and Siberia (e.g., Isayeva, 1984; Biryukov et al., 1988), but they achieved only little attention outside of Russia.

In the last decade, new efforts were started to solve the ice sheet problem. By revisiting key sites during intensive mapping work and through the application of advanced dating techniques it became clear that a large LGM ice sheet had existed neither in northern Siberia nor in Beringia (Svendsen et al., 1999; Brigham-Grette, 2001) and that larger glaciations must have been significantly older (Fig. 1). Recent results from field work carried out within the European Science Foundation program QUEEN suggest that the maximum eastward extension in the Weichselian was reached already at ca 90–80 ka (Svendsen et al., 2004). At such times, northward discharging rivers like Ob and Yenisei were blocked by the ice sheet, resulting in the formation of large freshwater lakes (Mangerud et al., 2001, 2004).

Continental glaciations often leave traces in deep-sea sediments. It had been shown earlier that coarse terrigenous detritus in deep-sea glacial sediments could be traced back to its place of origin in previously glaciated areas (Clark et al., 1980; Wohlfeil, 1983). The most likely transport agents are icebergs breaking off from calving ice sheets at the coastline. Thus, coarse ice-rafted debris (IRD) in deep-sea sediments provides a direct link to the timing and regional extent of continental glaciations. For the circum-Arctic land areas, such a connection was tried first by Clark et al. (1984), who correlated sediments from a western Arctic deep-sea core to glacial deposits on Banks Island in the western Arctic Archipelago. The composition and abundance of marine microfossils, another important component in Arctic sediments, holds clues to address the growth of ice sheets. Hebbeln et al. (1994) used

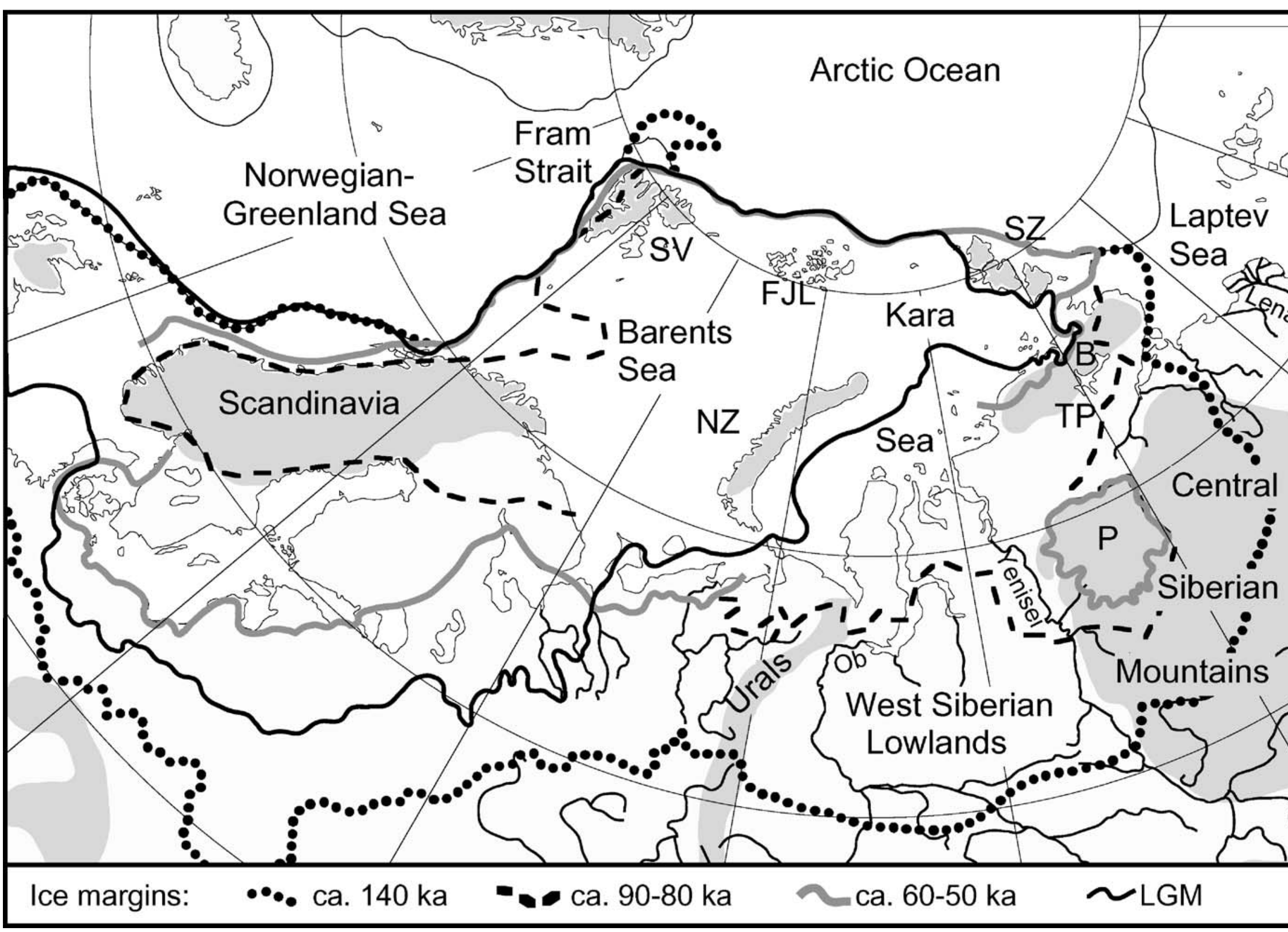

Fig. 1. Map of northern Eurasia with ice sheet boundaries during the Saalian glaciation (at ca 140 ka) and the three maximum glaciation stages of the Weichselian (redrawn from Svendsen et al., 2004). The exact position of the ice margin in northernmost European Russia (90–80 ka) and in the southernmost Kara Sea (60–50 ka) is still under discussion. Mountain ranges are marked by dark shading. Shelf areas are indicated by light shading. B, Byrranga Mountains; FJL, Franz-Josef-Land; NZ, Novaya Zemlya; P, Putorana Plateau; TP, Taymyr Peninsula; SV, Svalbard; SZ, Severnaya Zemlya; LGM, Last Glacial Maximum (ca 20 ka).

planktic foraminifer abundances as an indicator for open water environments. They demonstrated the close relationship of such open waters as a moisture source in the western Fram Strait and phases of build-up of the western Barents Sea ice sheet, just before and during the LGM. Third, the isotopic composition of planktic foraminifers from marine sediment cores is a measure of surface water salinity and its variation in the Arctic Ocean (Spielhagen and Erlenkeuser, 1994). Several excursions in records of oxygen and carbon isotopes of planktic foraminifers from Arctic sediment cores have been interpreted as releases of freshwater from collapsing continental ice sheets during glaciations and especially at glacial terminations (Stein et al., 1994a; Nørgaard-Pedersen et al., 1998).

In this study, we present a synthesis of data from five sediment cores along a transect from the Fram Strait across the Eurasian Basin to the central Arctic Ocean. Several data sets from these cores have been published before. However, we strongly revised older stratigraphic models to accord with the model of Jakobsson et al. (2000). While older models for central Arctic Ocean sediment cores (e.g., Clark et al., 1980; Spielhagen et al., 1997) gave average pelagic sedimentation rates of a few millimeters per 1000 years, Jakobsson et al. (2000) and Backman et al. (2004) could demonstrate that rates were significantly higher (1–5 cm/ka) at least for the last 200 ka. We use this high-resolution age model and several sediment proxies to determine (1) the times of Atlantic Water inflow and occurrences of open waters in the marginal Arctic, which acted as possible moisture sources for ice sheet build-up, (2) the deposition of IRD from floating icebergs, indicating a maximum northernmost extension of ice sheets, and (3) the outflow of large amounts of freshwater during deglacial events, which reflects melting ice sheets and the drainage of dammed lakes in Siberia to the Arctic Ocean. In combination with published data from the Eurasian continental margins, shelves, and land areas, we will use the results to reconstruct the history of glaciation in northernmost Eurasia and environmental changes in the Eurasian Basin of the Arctic Ocean during the last 200 ka.

2. Modern environmental setting

The Arctic Ocean comprises several subbasins of >3000 m water depth and is surrounded by the widest shelf areas of all oceans (Fig. 2). Especially, the Eurasian shelves of the Chukchi, East Siberian, Laptev, and Kara seas are extremely shallow and in large parts reach only 30–50 m maximum water depth. Below the shelf break at approximately 50 m water depth, the steep upper continental slope leads to the deep sea. Today, the deep Arctic Ocean area is perennially covered by sea ice, but the surrounding shelves usually become ice-free in summer. Rare icebergs originate from glaciated regions on Svalbard, Franz-Josef-Land, Severnaya Zemlya, Ellesmere Island, and Greenland. The ice drift pattern is dominated by the clockwise-rotating Beaufort Gyre in the Amerasian Basin and the Transpolar Drift in the Eurasian Basin (Fig. 2), which transports sea ice from the Siberian shelves to the Fram Strait and further southward along East Greenland. Sea ice formation on the shallow Siberian shelves is supported by low-salinity surface waters, originating from the huge freshwater discharge of Siberian rivers (Aagaard and Carmack, 1989). In the deep Arctic basin, the cold low-salinity surface water layer ($T < -1.5\,°C$, $S \approx 30-32$) is approximately 200 m thick and underlain by Atlantic Water of higher salinity ($S > 34.8$) (Anderson et al., 1994). The Atlantic Water core at 200–500 m may reach temperatures up to 2 °C, whereas bottom waters are slightly above $-1\,°C$. This Atlantic Water is entering the Arctic Ocean through the western Fram Strait and across the Barents Sea and submerges under the halocline (Rudels et al., 1994). Its inflow keeps large areas in the Fram Strait and the southern Nansen Basin ice-free in summer and represents an important source of heat and moisture. Otherwise the ice cover effectively prevents the exchange of heat and gas between the ocean and the atmosphere. Nevertheless, especially in the summer open water areas develop as leads between large ice floes and amount to 5–8% of the ice-covered area. Atmospheric heat transport from the south often keeps summer air temperatures over the ice well above the freezing point of $\sim -1.8\,°C$. Leads do not refreeze and provide a place for evaporation and the intrusion of daylight into the water column, which supports planktic productivity.

The Eurasian Arctic hinterland displays a very variable morphology. Ridges in Scandinavia and in the Ural, Byrranga, Putorana, Werchoyansk, and Anadyr mountains change with lowlands in the Pechora region and many coastal areas, and a large basin in western Siberia. The rivers Ob, Yenisei, and Lena draining the Arctic hinterland (among many others) belong to the nine largest rivers in the world in terms of freshwater discharge (Aagaard and Carmack, 1989).

3. Marine proxies for terrestrial ice sheet history

Onshore glacial successions often show major stratigraphic hiatusses resulting from the erosion of older sediments by one or more overriding ice sheet(s) or from subsequent abrasion by precipitation or fluvial erosion/incision. In contrast, sediment cores from submarine morphologic highs often contain continuous archives of the paleoenvironmental history. Furthermore, a number of stratigraphic methods are available for marine sediments to establish high-resolution age models. This

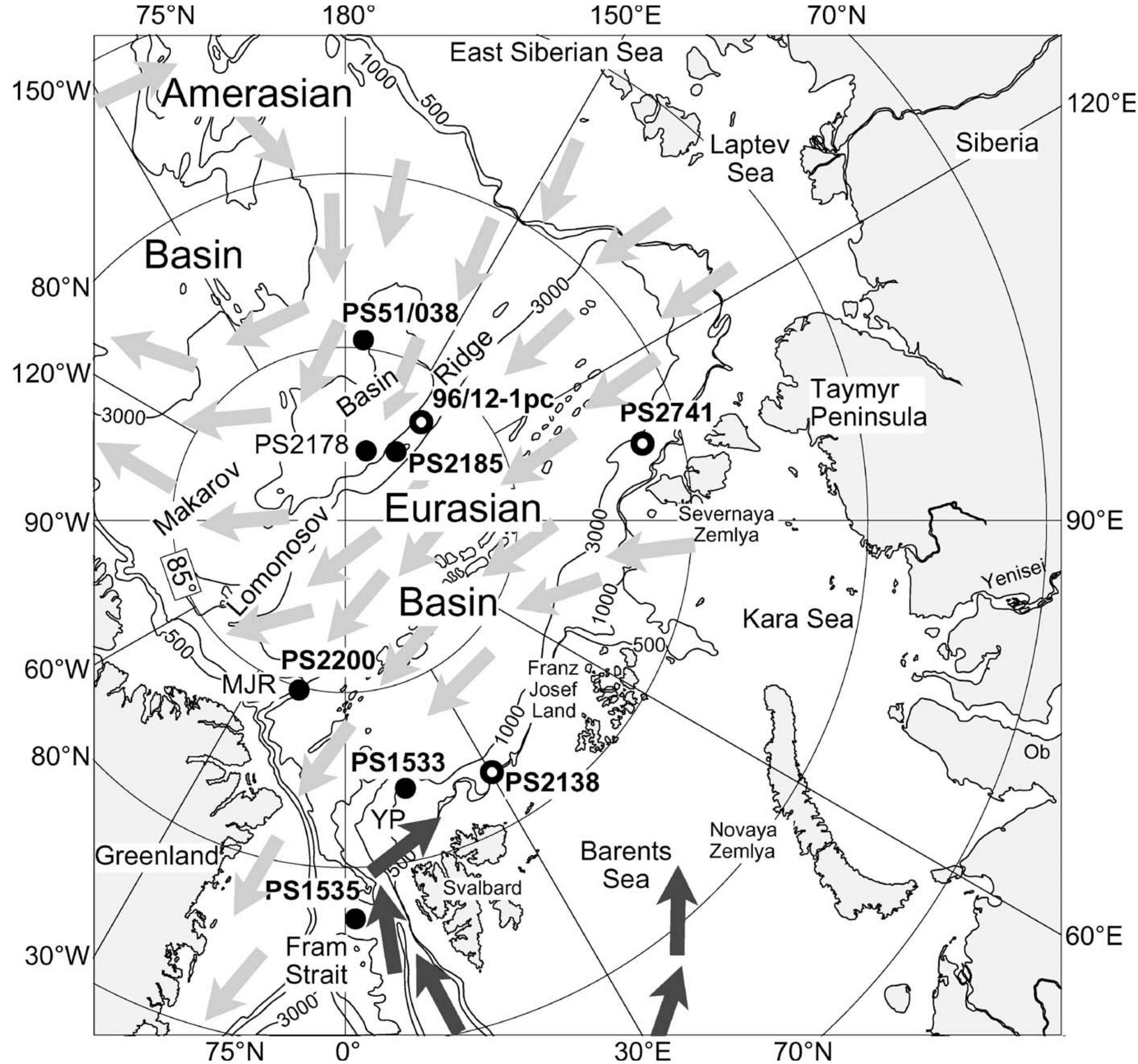

Fig. 2. Bathymetric map and main surface circulation features of the eastern and central Arctic Ocean and its marginal seas. Sites of sediment cores mentioned, in this study, are indicated. Dark arrows mark the inflow of Atlantic Water; light arrows mark the ice drift pattern of the Beaufort Gyre in the Amerasian Basin and the Transpolar Drift in the Eurasian Basin. MJR, Morris Jesup Rise; SAT, St. Anna Trough; YP, Yermak Plateau.

is of particular importance for reconstructions of the ice sheet history because several marine proxies directly reflect the development of ice sheets on land or the oceanic boundary conditions which influence the growth and decay of the ice sheets.

Moisture supply is an important prerequisite for the growth of any ice sheet. Precipitation in northernmost Siberia and on adjacent shelves is extremely low today (<300 mm/a). Paleobotanical data (Tarasov et al., 1999) and evidence for a reduced hydrologic cycle as revealed by ice core data (Alley et al., 1993) suggest dry conditions over large parts of northern Eurasia for the LGM. This underlines the importance of moisture sources adjacent to growing ice sheets. Hebbeln et al. (1994) showed that times with open waters in the western Fram Strait just before and during the LGM coincided with phases of build-up of the western Barents Sea ice sheet. Based on an earlier study of modern sedimentation in ice-covered and (seasonally) ice-free areas of the Fram Strait (Hebbeln and Wefer, 1991), they concluded that the high abundances of polar and subpolar planktic foraminifers in LGM sediments are a proxy of a strong inflow of temperate Atlantic Water into the Arctic Ocean. Thus, high abundances of microfossils in Arctic and sub-Arctic sediment cores have been used as evidence for open water conditions and Atlantic Water inflow events (of variable intensity) during several intervals in the last 130,000 years (e.g., Gard, 1993; Dokken and Hald, 1996; Hald et al., 2001; Matthiessen et al., 2001). Recently, the LGM foraminifer abundance peak could be traced up to 84°N into the Nansen Basin (Nørgaard-Pedersen et al., 2003). Considering the modern situation in the central Arctic Ocean where ca 5–8% open water areas in the summer promote a relatively "strong" bioproductivity in the surface waters (Gosselin et al., 1997), the term "open water conditions" will henceforth be used for similar ice conditions in the interior Arctic Ocean at times of strong Atlantic Water inflow in the past. Independent of any stratigraphic model, it could be shown that concentrations of cosmogenically produced ^{10}Be in core PS2185 from the Lomonosov Ridge are high in sediments with high abundances of planktic foraminifers (Spielhagen et al., 1997). Considering the general dependence of ^{10}Be fluxes on biological productivity in the euphotic zone (Kusakabe et al., 1987), it is suggested that ^{10}Be concentration peaks in foraminifer-free sections indicate (seasonally) open waters and high productivity. The primary indicators for such environments may have been subject to carbonate dissolution on the seafloor. We note that this is in line with conclusions of Eisenhauer et al. (1994) on the variability of Arctic ^{10}Be fluxes between cold and warm climatic phases.

Once an ice sheet or glacier expands to the shoreline, icebergs are released which can drift toward the open ocean, carrying terrestrial debris (IRD) of clay to boulder size adfrozen at the base, on their surfaces and within the ice. In contrast, modern Arctic sea ice usually carries only fine-grained sediments of silt and clay size (Nürnberg et al., 1994). During the Late Quaternary cold phases when the sea level was lowered by more than 50 m (Lambeck and Chappell, 2001), the northern shoreline of Eurasia must have moved to or close to the recent shelf break. Under such circumstances and especially during times of sea-level rise (e.g., glacial terminations), icebergs from extensive shelf glaciations could easily escape to the open Arctic Ocean. On the other hand, a decreasing flow velocity, a regional recession of an ice sheet from the shoreline, or a rapidly falling sea level would have diminished or even stopped the discharge of icebergs and deposition of coarse IRD. Sediments on the Arctic continental slopes and in the deep basins were to a large part deposited by gravity-controlled processes (Goldstein, 1983; Kleiber et al., 2001; Svindland and Vorren, 2002), although they may contain hemipelagic sequences of variable grain-size. Sediment cores from morphological highs in the Eurasian Basin (Morris-Jesup Rise, Yermak Plateau, Gakkel Ridge, and Lomonosov Ridge) are usually free of turbidities and typically show a succession of fine- and coarse-grained layers (Fütterer, 1992; Vogt, 1997). Holocene deep-sea sediments from the Eurasian Arctic Ocean generally have a low coarse-fraction content (<10 wt%, Nørgaard-Pedersen et al., 1998, 2003) which reflects the scarcity of icebergs in the modern Arctic. Older layers are often rich in coarse IRD and attributed to times with abundant icebergs in the Arctic Ocean (Elverhøi et al., 1995; Bischof et al., 1996; Spielhagen et al., 1997; Knies et al., 2000, 2001). A specific mineralogical composition of IRD can give important information about the provenance of the icebergs if potential sources are known (Darby et al., 1989; Bischof et al., 1996; Spielhagen et al., 1997; Knies et al., 2000; Vogt et al., 2001). Because coarse-fraction content and abundances of planktic foraminifers are in strong anticorrelation in most sediment cores from the Arctic Ocean (Pak et al., 1992; Darby et al., 1997; Spielhagen et al., 1997; Nørgaard-Pedersen et al., 1998, 2003; this study), we use the coarse-fraction content as a proxy for IRD deposition from icebergs and as evidence for the existence of glaciers or ice sheets at sea level. Exceptions from this rule are easily identified from the comparison of the >63 μm and foraminifer abundance records. We note a pronounced positive correlation ($r = 0.6$) between the contents of the >63 μm fraction and the undoubtedly iceberg-rafted >250 μm fraction in the coarse-grained, foraminifer-poor layers, which again suggests icebergs as the predominant transport agents for these sediments.

In general, oxygen isotope ($\delta^{18}O^{18}O/^{16}O$) records of foraminifer shells from marine sediment cores reflect the variability of continental ice volume through time (Shackleton, 1967) and show a typical "global" pattern of low interglacial and higher glacial values (Imbrie et al., 1984). This holds true for most benthic foraminifer δ^{18}O records from deep-sea sites because temperature changes (another factor influencing the isotopic composition) in great water depths are supposed to be minimal in the Late Quaternary glacial–interglacial cycles (Shackleton, 1967). Planktic foraminifers in the Arctic Ocean usually live in the uppermost water column (Carstens and Wefer, 1992) and their δ^{18}O values mainly reflect the regional salinity variations caused by the inflow of low saline, ^{18}O-depleted river water (Spielhagen and Erlenkeuser, 1994; Bauch et al., 1997). Planktic δ^{18}O records of Arctic and sub-Arctic sediment cores often deviate from the "global" pattern because additional ^{18}O-depleted meltwater has entered the surface ocean during glaciations and especially in the glacial terminations. Several such meltwater events were found in Weichselian records from the sub-Arctic Norwegian–Greenland Sea and correlated to the history of ice sheets on Scandinavia, the Barents Sea and Greenland (e.g., Jones and Keigwin, 1988; Köhler and Spielhagen, 1990; Sarnthein et al., 1992; Stein et al., 1996; Hald et al., 2001). Well-dated high-resolution records from the Arctic Ocean are mostly confined to the range of radiocarbon datings (i.e., the last ca 45 ky) and show several meltwater spikes (Stein et al., 1994a; Nørgaard-Pedersen et al., 1998, 2003; Poore et al., 1999), with the strongest ones during the glacial termination. Although carbon isotope ($\delta^{13}C = {}^{13}C/{}^{12}C$) data from high-latitude planktic foraminifers are more difficult to interpret than $\delta^{18}O$ records (Bauch et al., 2000; Kohfeld et al., 2000), Arctic records typically show a low-$\delta^{13}C$ signal associated with meltwater spikes in $\delta^{18}O$ records (Stein et al., 1994a; Nørgaard-Pedersen et al., 1998, 2003; Poore et al., 1999). This feature is ascribed to reduced ventilation of surface waters caused by time-limited, extremely strong stratification from the surficial low-salinity freshwater lid. In contrast, the persisting Holocene river-runoff, which also established a strong stratification, results in relatively high $\delta^{13}C$ values in Arctic Ocean surface waters and planktic foraminifers, caused by atmosphere–water exchange of CO_2 in the rivers and on the shelves (Spielhagen and Erlenkeuser, 1994; Bauch et al., 2000).

In this study, we use planktic $\delta^{18}O$ and $\delta^{13}C$ records from long, high-resolution sediment cores to determine the timing of abrupt freshwater discharges to the Arctic Ocean. In accordance with many other studies in (sub-)Arctic seas, we identified such events by both low oxygen isotope and low carbon isotope values in the same or the consecutive sample. Two scenarios can be responsible. First, the disintegration of huge ice masses, caused by a sea-level rise or by rapid ice surges and discharging subglacial meltwater and icebergs to the ocean. Melting of the icebergs will release additional meltwater. Second, the deglacial northward break-through and outburst of ice-dammed lakes which were fed by northbound rivers at times when ice sheets blocked the discharge to the Arctic Ocean. The latest ice sheet reconstructions for the Early to Late Weichselian (Svendsen et al., 1999, 2004; Mangerud et al., 2002) and evidence for the formation of large ice-dammed lakes in northern Russia and Siberia (Houmark-Nielsen et al., 2001; Mangerud et al., 2001, 2004; Maslenikova and Mangerud, 2001) present a new terrestrial framework for the interpretation of freshwater spikes in isotopic records from the eastern and central Arctic Ocean.

4. Materials, methods, and data sources

Sediment cores, in this study, were obtained during Arctic cruises of R.V. *Polarstern* in 1987, 1991, and 1998 (Fig. 2) from the "Arctic Gateway" (Fram Strait site PS1535 and Yermak Plateau site PS1533) and the central Arctic Ocean. Details about cores and coring sites are given in Table 1. Cores PS1535-5 (box core), -8 (square barrel kastenlot core), and -10 (piston core) are from the same site, but of different length and diameter. A detailed core-to-core correlation is given in Nowaczyk and Baumann (1992). The composite of the three cores from site PS1535 is further on referred to as "core PS1535". Accordingly, "core PS2178", "core PS2185", and "core PS2200" denote the composite of the box cores PS2178-3, PS2185-3, or PS2200-2 and the square barrel kastenlot cores PS2178-5, PS2185-6, or PS2200-5 from the Makarov Basin, the Lomonosov Ridge, and Morris Jesup Rise ("central Arctic"), respectively (cf. Spielhagen et al., 1997). All cores were sampled continuously at 1–2 cm intervals (6 cm in PS2178-5) for volumes of 30–200 cm^3 per sample. Samples were freeze-dried, wet-sieved with deionized water through a 63 μm mesh, dried at 40 °C, and split into several fractions. Foraminifer counts were determined on representative sample splits of >500 grains per sample (125–500 μm fraction). For stable isotope analysis, 25 specimens of planktic foraminifers *Neogloboquadrina pachyderma* (sin.) were picked from the 125–250 μm fraction, crushed, and cleaned by ultrasonic agitation. Stable

Table 1
Information about sites, cores, and data sources of analyzed sediment cores used in this study

Site	PS1533	PS1535	PS2178	PS2185	PS2200	PS51/038
Position	82°01.9′N 15°10.7′E	78°45.1′N 01°51.0′E	88°00.2′N 159°14.0′E	87°31.9′N 144°22.9′E	85°19.6′N 14°00.0′W	85°08.1′N 171°26.4′W
Water depth (m)	2030	2557	4009	1051	1074	1473
Max. core length (m)	4.77	8.13	8.30	7.68	7.70	7.19
Gear	GC	BC, KAL, PC	BC, KAL	BC, KAL	BC, KAL	KAL
Grain size	a, ts	b, ts	ts	c	ts	ts
Plankt. for.	a, ts	d, ts		c	ts	ts
AMS ^{14}C	e, f	e, f		c, g	g	
$\delta^{18}O$, $\delta^{13}C$	e, ts	b, ts		ts	ts	ts
IChRM	h	h	i	j	j	ts
^{10}Be	k		l	l	m	
Smectite				c	n	

Note: BC, box corer, KAL, kastenlot corer, PC, piston corer. (a) Pagels (1991), (b) Köhler and Spielhagen (1990), (c) Spielhagen et al. (1997), (d) Spielhagen (1991), (e) Köhler (1992), (f) Nørgaard-Pedersen et al. (2003), (g) Nørgaard-Pedersen et al. (1998), (h) Nowaczyk and Baumann (1992), (i) Nowaczyk et al. (2001), (j) Frederichs (1995), (k) Eisenhauer et al. (1994), (l) Schäper (1994), (m) Molnar (1995), (n) Vogt (1997), and ts, this study.

oxygen and carbon isotopes were measured by standard techniques (Duplessy, 1978; Winn et al., 1991) on the automated Carbo-Kiel device connected to a Finnigan MAT 251 mass spectrometer. Results are expressed in the δ notation referring to the PDB standard (via NBS 20) and are given as $\delta^{18}O$ and $\delta^{13}C$ values. The external analytical reproducibility is 0.08‰ and 0.04‰ for $\delta^{18}O$ and $\delta^{13}C$, respectively. Paleomagnetic sediment parameters were determined according to procedures described by Frederichs (1995) and Nowaczyk et al. (2001).

Nannofossil analyses of cores PS2185 and PS51/038-4 were performed on selected foraminifer- or carbonate-rich samples only, using a grain-size-based correlation to nannofossil-bearing layers in R.V. *Oden* core 96/12-1pc (Jakobsson et al., 2001). We used samples which had been freeze-dried only a few months after retrieval of the cores in 1991 and 1998 so that possible early diagenetic carbonate dissolution should be minimized. For preparation of coccolith samples a combined dilution/filtering technique as described by Andruleit (1996) was used. A sediment sample was brought into suspension and further diluted with a rotary splitter. The suspension was filtered onto polycarbonate membrane filters (Schleicher & Schuell™, 0.4 μm pore size) and dried at 40°C. A small piece of the filter (1 × 1 cm) was cut out and mounted on an aluminum stub. Coccoliths were searched by means of a scanning electron microscope (SEM) at a magnification of 2000 × or 3000 ×. Since coccoliths were extremely rare, only absence and presence data have been determined.

Sources for published data are given in Table 1. Radiocarbon datings on planktic foraminifers have been corrected for a reservoir age of 400 years and were converted to calendar ages (cal. ka) by the CALIB 4.3 calibration program using the "marine" calibration data set (Stuiver and Reimer, 1993; Stuiver et al., 1998) and, beyond 20.3 ^{14}C-ka, by applying the age shift determined by Voelker et al. (1998). All data are stored in the PANGAEA database and can be accessed on the website http://www.pangaea.de.

5. Stratigraphy of Arctic sediment cores

First stratigraphic models for Arctic sediment cores obtained from the Amerasian Basin in the 1950s and 1960s gave extremely low average sedimentation rates of only a few millimeters per 1000 years for the Quaternary (e.g., Clark, 1970; Herman, 1970). Based on a variety of methods including nannofossils, radioisotopes, and stable isotopes, it could be demonstrated that average rates for the Late Quaternary in the Fram Strait and the adjacent Arctic Ocean were one order of magnitude higher (Gard, 1986, 1987; Baumann, 1990; Eisenhauer et al., 1990; Köhler and Spielhagen, 1990). Extension of the "^{10}Be stratigraphy" for Fram Strait and Yermak Plateau cores to the Gakkel Ridge (Eisenhauer et al., 1994) and the Lomonosov Ridge (Spielhagen et al., 1997) suggested average rates of 0.5 cm/ka for the central Arctic Ocean. However, nannofossil data and a new interpretation of the paleomagnetic record from the Lomonosov Ridge (Jakobsson et al., 2000; Backman et al., 2004) leave no doubt that these estimates were far too low. OSL datings of ice-rafted quartz grains recently confirmed the new age model (Jakobsson et al., 2003). Here we base our stratigraphy approach for the eastern and central Arctic Ocean cores on correlation to the Jakobsson et al. (2000) age model for core 96/12-1pc from the Lomonosov Ridge. Based on several parameters we can now present a core-to-core correlation from the central Fram Strait (78°45′N) to the Yermak Plateau and the Morris Jesup Rise and across the Lomonosov Ridge to the Alpha Ridge in the Amerasian Basin (85°08′N, 171°27′W).

Paleomagnetic inclination data form a baseline for correlation (Fig. 3). Cores from the central Arctic (PS2200, PS2185, and PS51/038-4, as well as core 96/12-1pc of Jakobsson et al., 2000) contain a thick upper section of normal polarity. The reversed interval below was identified as the Biwa II event by Jakobsson et al. (2000), resulting in an age estimate of approximately 240 ka. In the inclination record of core PS1535, Nowaczyk and Baumann (1992) identified the Biwa II interval at 705–730 cm, based on comparison to nannofossils occurrences (Baumann, 1990), planktic foraminifer stable isotopes (Köhler and Spielhagen, 1990), as well as ^{230}Th and ^{10}Be data (Eisenhauer et al., 1990). ESR datings of planktic foraminifers from core PS1535 (Hoffmann et al., 2001) gave an age of 243 ± 30 ka for 700 cm core depth and confirm the paleomagnetic interpretation. Beyond the range of radiocarbon datings in this core we refined the age model of Köhler and Spielhagen (1990) by the identification of (sub)stages (OIS) and events in the planktic oxygen isotope record (Fig. 4) according to Martinson et al. (1987). Oxygen isotope event (OIE) 7.1 (193 ka) is assigned to the low value peak at 552 cm and represents the oldest tie point for the stratigraphic model of sediments from the last 200 ka. In a similar manner and largely in correlation to PS1535, oxygen isotope stratigraphy was applied to core PS1533-3. The base of this core is supposed to be of mid-OIS 6 age (165 ka).

Although a calcareous nannofossil biozonation scheme for high northern latitudes already exists (Gard, 1988; Gard and Backman, 1990), its use is often hindered by the scarcity of coccoliths in Arctic sediments. Thus, as done by Jakobsson et al. (2000, 2001), occurrences of *Gephyrocapsa muellerae* and *Emiliania huxleyi* were primarily used to identify OIS 5 and its substages.

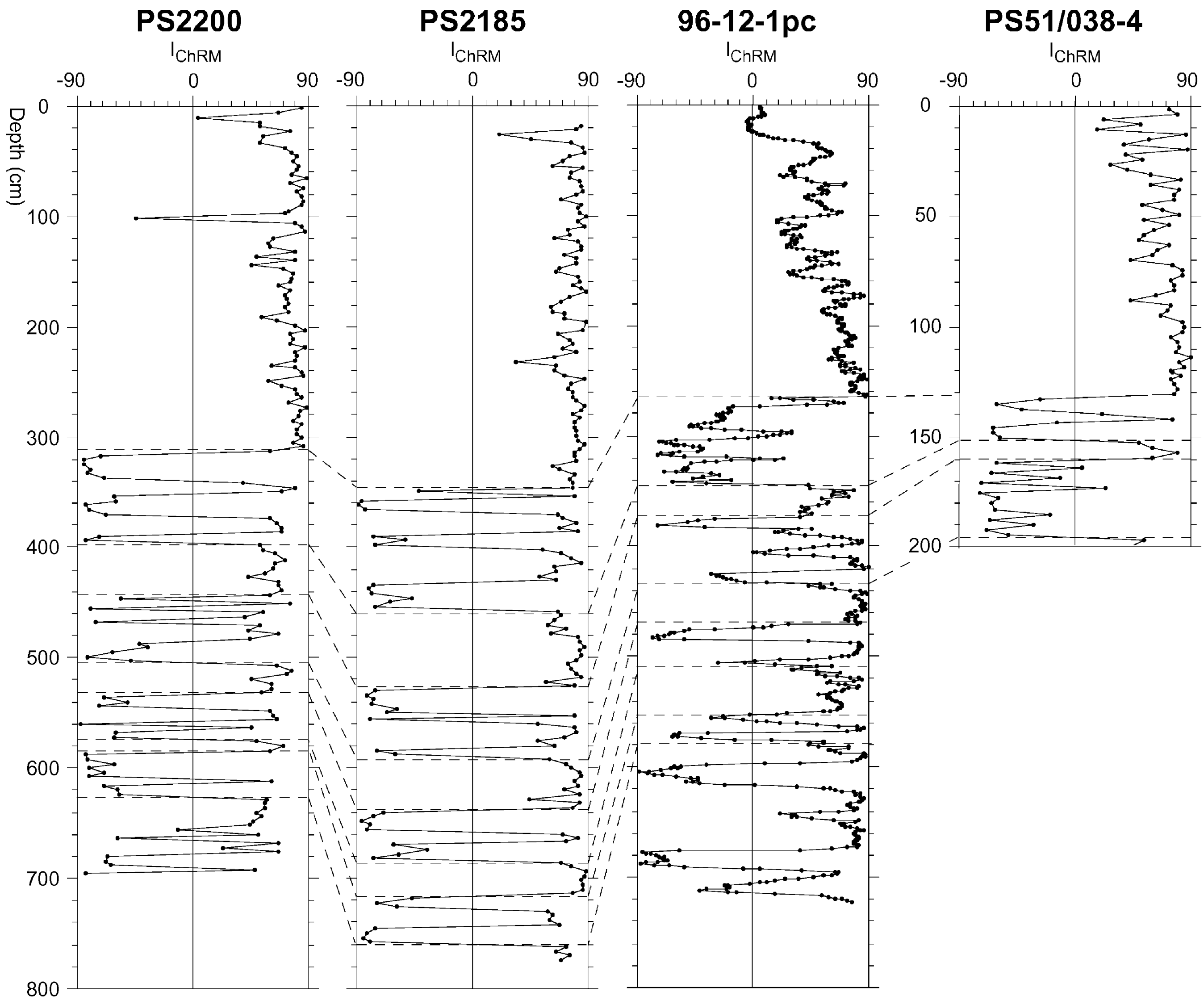

Fig. 3. Paleomagnetic inclination of the characteristic remanent magnetization (IChRM) of sediment cores from the central Arctic Ocean. Correlation of PS2200 and PS2185 is from Frederichs (1995). Correlation of PS2185 and 96/12-1pc is from Jakobsson et al. (2001).

In sediment core PS2185 from the Lomonosov Ridge, nannofossils are present at depths of 181–183, 205–217, and 229–245 cm. The assemblages consist of *G. muellerae*, *E. huxleyi*, and other poorly preserved placoliths (most probably representing heavily corroded specimens of *G. muellerae*). In addition, a few in situ specimens of *Calcidiscus leptoporus*, as well as traces of *Helicosphaera carteri* and *Coccolithus pelagicus* have been observed (Table 2). The occurrence of *G. muellerae* together with *E. huxleyi* unambiguously can be assigned to OIS 5. This finding corresponds well with studies of sediment cores from the Nansen Basin and the Gakkel Ridge (Baumann, 1990; Gard, 1993), where comparable coccolith assemblages were used to define OIS 5. In addition, the single occurrence of *C. pelagicus* at 181 cm may indicate OIS 5a. Thus, identification of the three coccolith-bearing intervals as OIS 5a, 5c, and 5e seems most likely and is supported by the planktic foraminifer distribution (cf. Figs. 4 and 5). Accordingly, we use the coccolith findings at 181, 213, and 241 cm to identify the stratigraphic tie points of OIE 5.1, 5.3, and 5.5, respectively.

In comparison to the Lomonosov Ridge and especially to Nansen Basin and Gakkel Ridge sediments, core PS51/038-4 from the Alpha Ridge contains only traces of nannofossils. This corresponds with earlier findings (Baumann, 1990; Gard, 1993) that the abundance of nannofossils decreased from the Fram Strait toward the northeast, in line with the decreasing influence of the Atlantic Water. Mainly *G. muellerae* and *E. huxleyi* were observed in PS51/038-4. Again this indicates sampling of OIS 5 sediments, which is uncertain at least at 63 cm, but may be extending down to 127 cm. Evidence is not fully conclusive since traces of

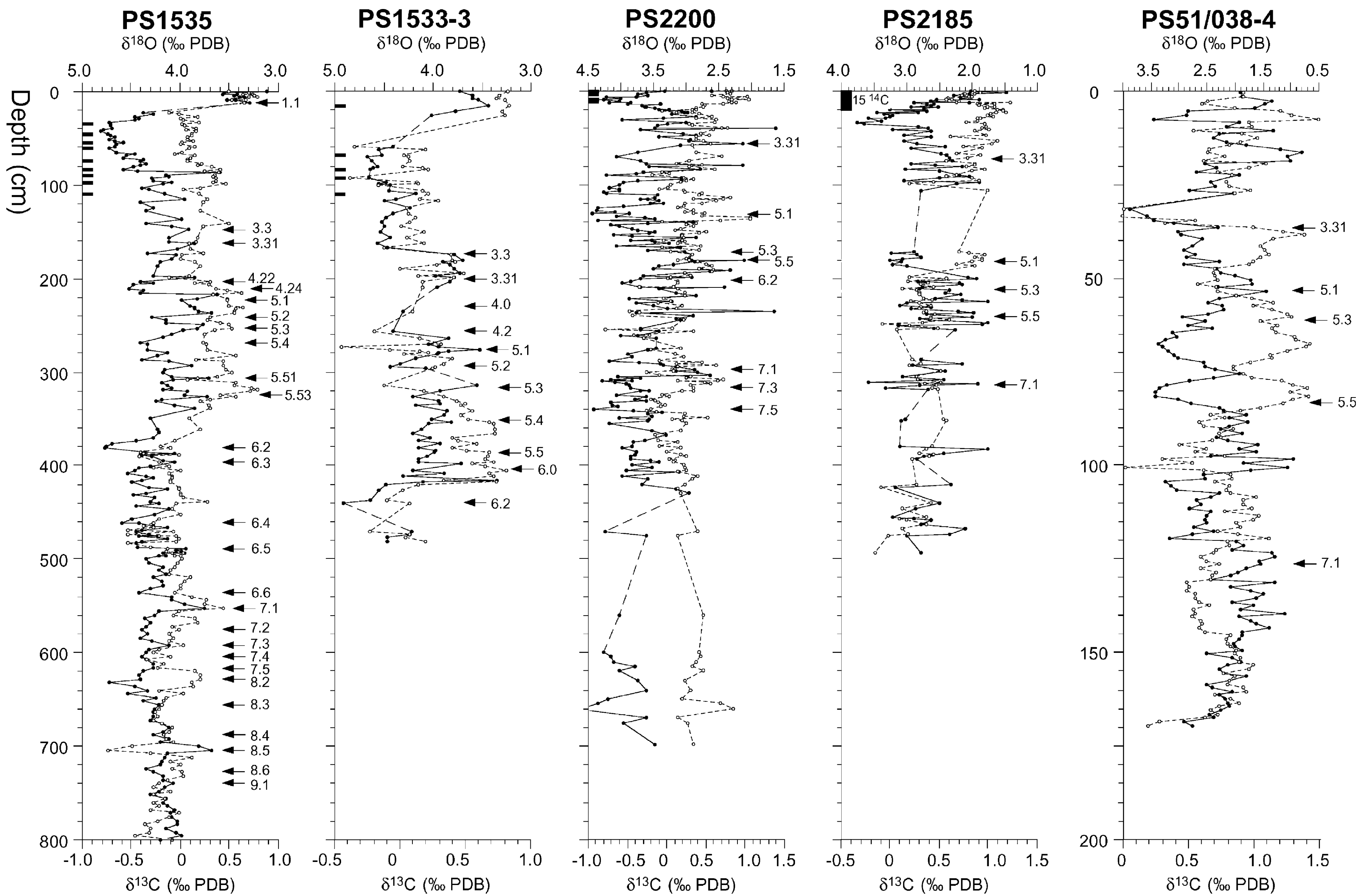

Fig. 4. Oxygen and carbon isotope data of planktic foraminifers *N. pachyderma* (sin.) from sediment cores along a transect from the central Fram Strait to the Alpha Ridge. Identified OIEs (cf. Martinson et al., 1987) are indicated. AMS-^{14}C-dated samples are marked by black bars on the depth scale.

Table 2
Results of nannofossil analyses of selected samples from sediment cores PS2185-6 and PS51/038-4

	Depth (cm)	*C. leptoporus*	*C. pelagicus*	*E. huxleyi*	*G. muellerae*	*Placoliths*	*H. carteri*	Additional findings
PS 2185-6	174							
	181	X	x	X	X	X	x	
	183	x		X	x			
	205			X	x	X		
	211			X	x	x		
	213	X		X	X	x		Silicoflagellates
	217	x		X	x	X		Tintinnids
	229			x		x		
	231			x	x	x		
	235			x	X			
	237			X	X	x		
	241	x		X	X	X		
	243			x	x	x		
	245	x		x	X	x		
PS 51/038-4	63			X	X	x		
	78			x		x		
	127			x	X			
	180			x		x		
	250							
	357			X		x	x	

Note: X = few and x = traces only.

Quaternary coccoliths were also found further downcore, at 180 and 357 cm. Although *E. huxleyi* specimens were recorded, the stratigraphic interpretation of these intervals remains uncertain. At present, we hesitate to use the coccolith findings in PS51/038-4 for stratigraphic purposes.

^{10}Be data were used to correlate cores PS1533-3, PS2178-4, PS2185, and PS2200 (Fig. 6). The correlation of the PS1533-3 record to Fram Strait and Norwegian Sea records was discussed in detail by Eisenhauer et al. (1994), revealing that high and low concentrations are found in sediments from climatically warm and cold phases, respectively. They conclude that, in addition to the effect of variable bioproductivity, deposition of the cosmogenically produced ^{10}Be in the Late Pleistocene Arctic Ocean was closely correlated to the deposition of fine-grained particles which, by scavenging, act as a carrier to the seafloor. Accordingly, cores with thick coarse-grained, IRD-rich sequences (PS2178, PS2185, PS2200, and lowermost part of PS1533-3, cf. Fig. 5) reveal intervals with very low ^{10}Be concentrations, whereas the variability in core PS1533-3, which generally contains less IRD in the upper 400 cm, is lower.

Detailed paleomagnetic and rock magnetic analyses of core PS2178 by Nowaczyk et al. (2001) had revealed an outstanding peak in the record of relative paleointensity at 180 cm which was correlated to a peak centered around 50 ka in a stacked global record of relative paleointensity variations of the geomagnetic field (Guyodo and Valet, 1999). This peak is situated in PS2178 directly above the IRD-rich layer that is found in all other cores from the interior Arctic and is considered to be of OIS 4 or early OIS 3 age on the basis of nannofossil and ^{10}Be data. There is independent paleomagnetic evidence that sedimentation rates were relatively high (several cm/ka or more) at site PS2178 at that time (Nowaczyk et al., 2001), which largely excludes the influence of bioturbation on the stratigraphic position of the paleointensity peak in the sediment column. From the very similar grain-size pattern in the cores we suppose that the change from coarse- to fine-grained sediment deposition was synchronous in the central Arctic Ocean, at least on time scales that can be resolved in our cores (i.e., including uncertainties of ± 2–5 ka, caused by bioturbation). We therefore apply an age of 50 ka to the grain-size change in early OIS 3 in cores PS2185, PS2200, and PS51/038-4. It should be noted that the stratigraphic models for cores PS1535 and PS1533-3, which were established independent of the paleomagnetic tie point at 50 ka, nevertheless give an age of 50 ± 1 ka for the strong decrease in IRD deposition. Stratigraphy within the uppermost part of the interval with normal inclination in our cores is based on ^{14}C datings (cf. Fig. 4) and was discussed in detail by Nørgaard-Pedersen et al. (1998, 2003).

No ^{14}C datings are available yet from core PS51/038-4 from the Alpha Ridge. Using the tie point at 50 ka to date the grain-size change at 7.5 cm, sedimentation rates thereafter must have been extremely low and we hesitate

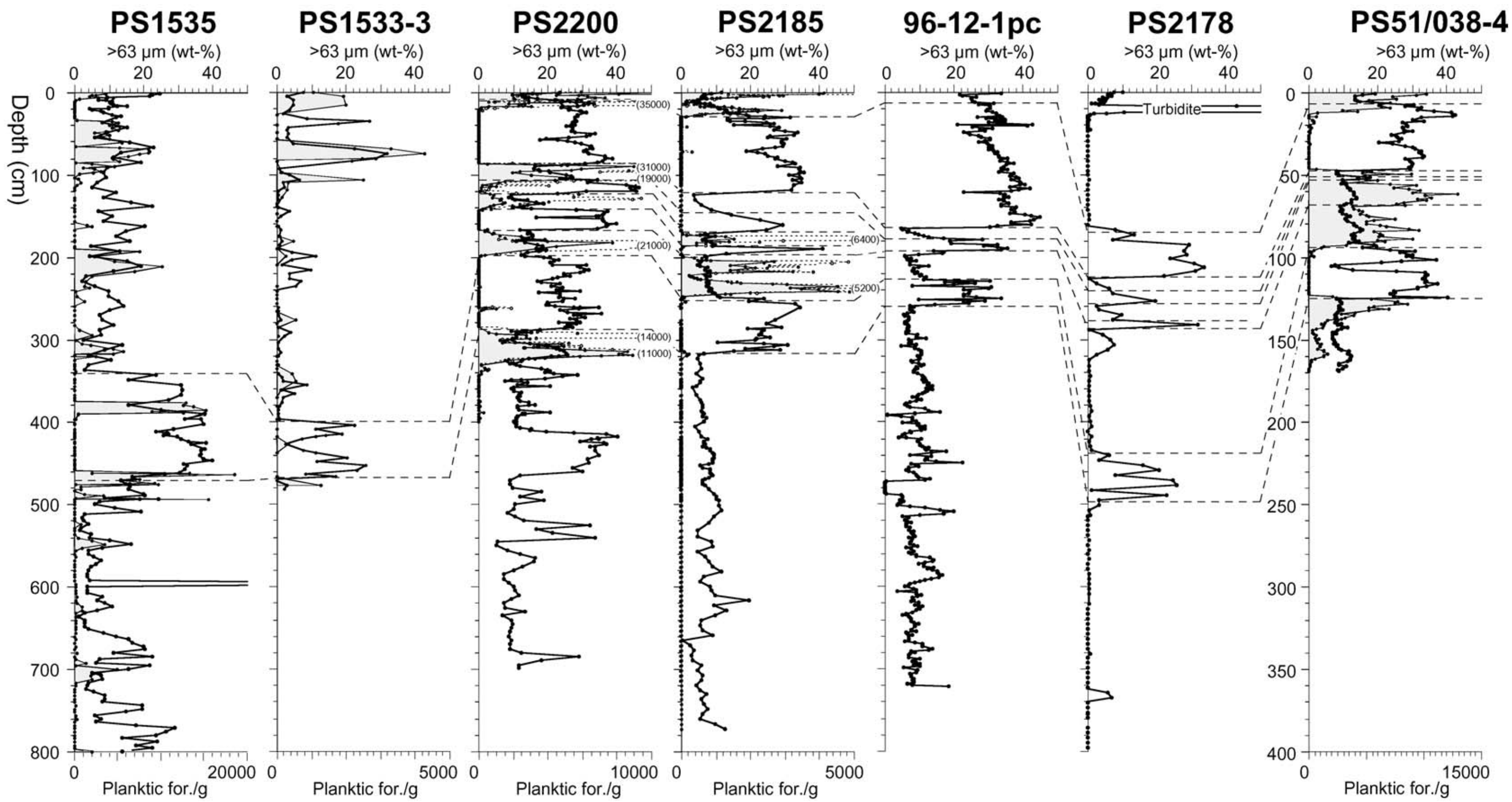

Fig. 5. Coarse-fraction (> 63 µm) content and abundance of planktic foraminifers *N. pachyderma* (sin.) in sediment cores along a transect from the central Fram Strait to the Alpha Ridge. Dotted lines in records of PS2200 and PS2185 give calculated fluxes of planktic foraminifers (specimens/cm^2 ka; same scale as foraminifer abundance). Data of 96/12-1pc are from Jakobsson et al. (2000).

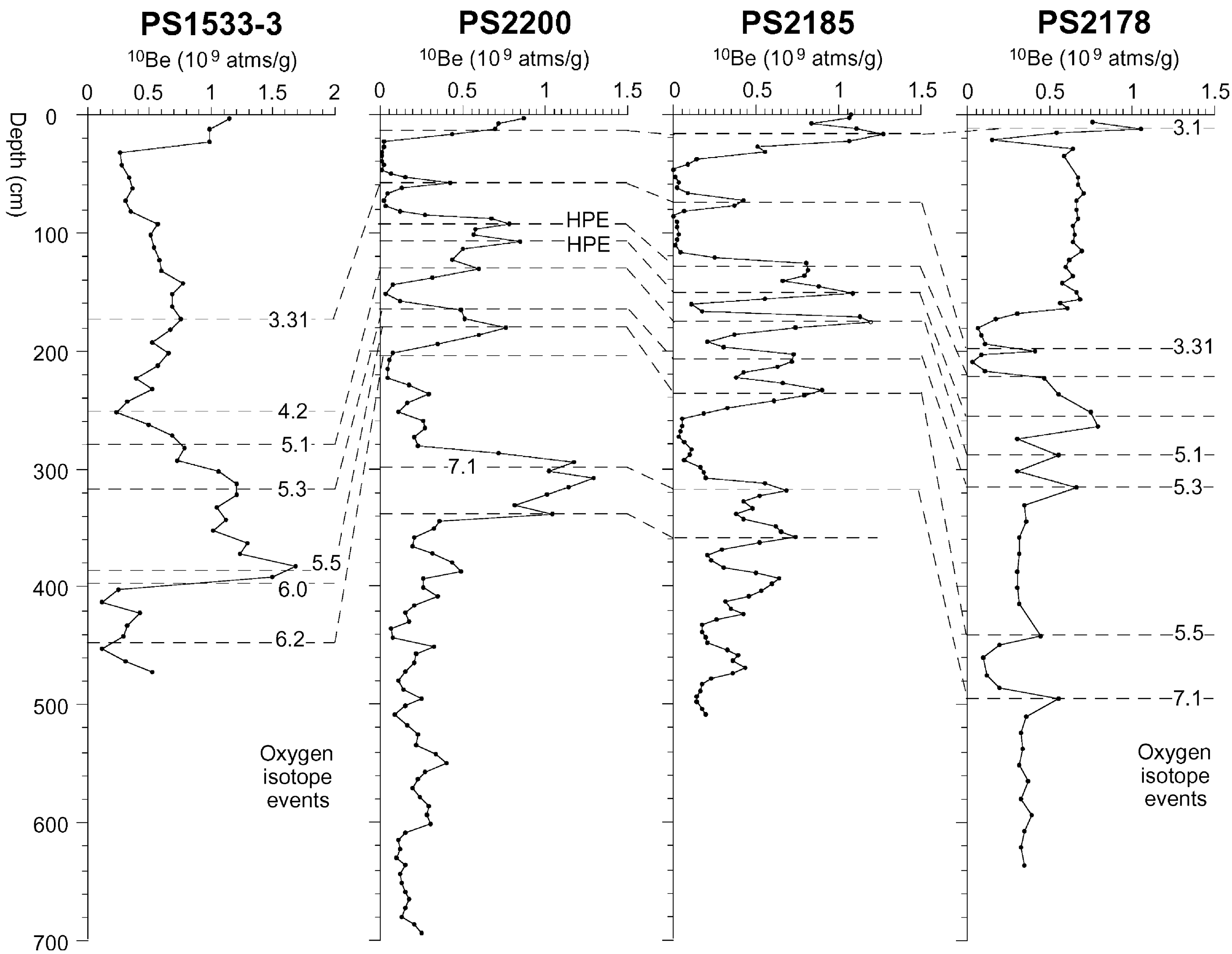

Fig. 6. ^{10}Be data of sediment cores along a transect from the Yermak Plateau to the Morris Jesup Rise and across the Lomonosov Ridge to the Makarov Basin. Correlation to OIEs (cf. Martinson et al., 1987) is indicated. HPE, high production events identified in the planktic foraminifer abundance records of core PS2200.

to draw any paleoenvironmental conclusions from our data of the upper 7 cm in this core. Another tie point is the identification of the younger boundary of the Biwa II event at 133 cm in the paleomagnetic inclination record. Between these tie points, we used a correlation to core PS2185 by grain-size and planktic foraminifer abundance data (Fig. 5) to establish a preliminary age model for core PS51/038-4.

6. Proxy data sets for the last 200 ka from Arctic sediment cores

6.1. Planktic foraminifer abundances

The number of planktic foraminifers in Arctic Ocean sediments from the last 200 ka is strongly variable (Fig. 7). Many samples are barren or contain only minimal amounts which allow at least stable isotope measurements. Relative to these sediments, some layers are rich in planktic foraminifers, although the absolute numbers of specimens per gram is quite low if compared to many sediment cores from lower latitudes. Since no dry bulk density data are available from the Arctic Gateway cores, it is not possible to calculate planktic foraminifer fluxes over the last 200 ka for these sites and we are generally cautious to apply a linear correlation between foraminifer abundance and environmental parameters (volume of Atlantic Water inflow, degree of ice coverage). However, comparison of planktic foraminifer abundances with calculated fluxes from PS2185 and PS2200 (Fig. 5) suggests that the abundance data give a good representation of the planktic foraminifer sedimentation in the Arctic deep-sea.

A peak in planktic foraminifer abundance is found in all sediments from uppermost OIS 7 (OIE 7.1). In Arctic Ocean cores PS2200, PS2185, and PS51/038-4, our stratigraphic models give an age of ca 195 ka for the

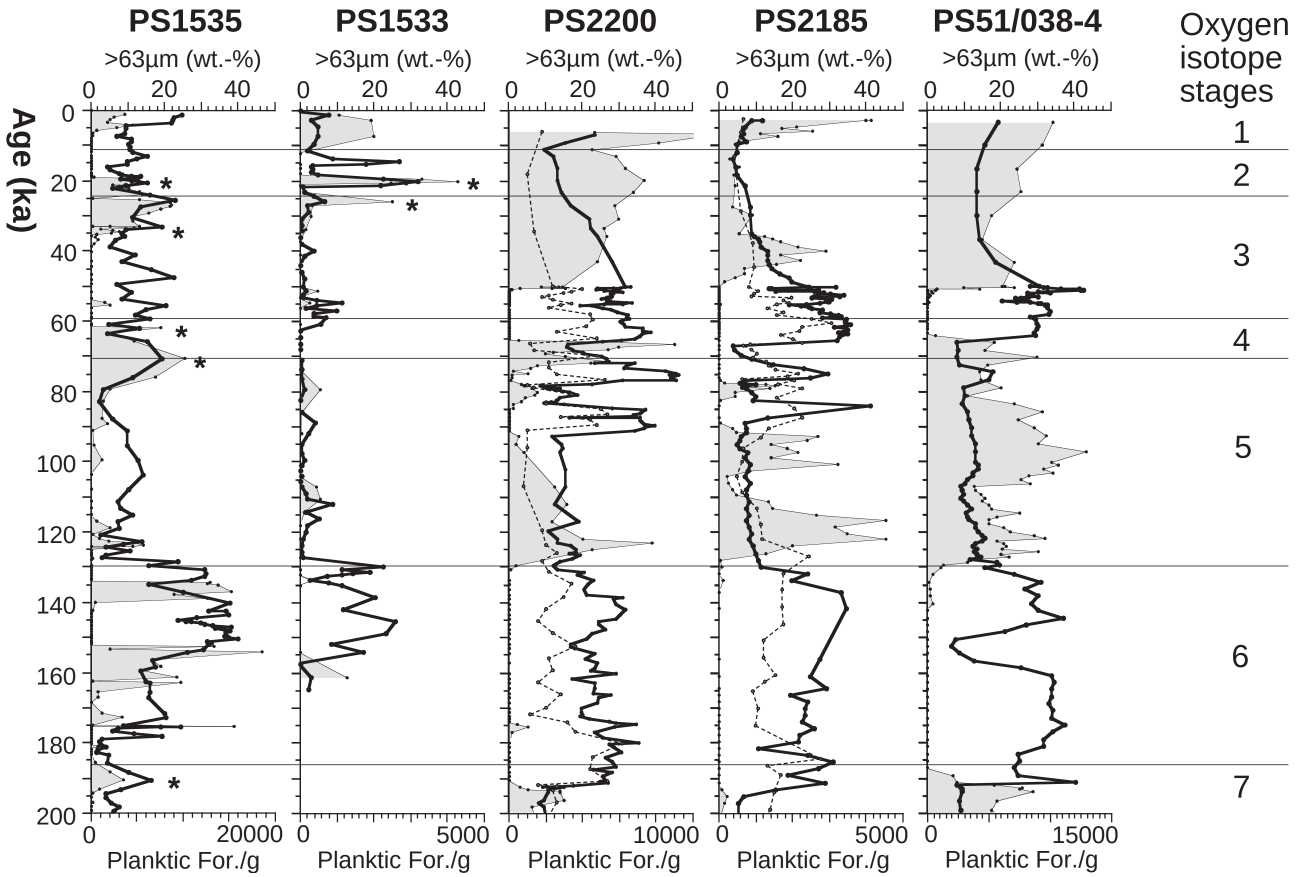

Fig. 7. Abundance of planktic foraminifers and coarse fraction in Arctic sediment cores within the last 200 ka. Asterisks mark layers where high coarse-fraction content results from abundant planktic foraminifers. Dotted lines represent smectite content in PS2185 and PS2200 (% of the <2 µm fraction; same scale as >63 µm).

abundance peak, whereas in Fram Strait core PS1535 it is ca 190 ka. Arctic Ocean sediments from OIS 6 contain extremely low amounts of planktic foraminifers and in several cases the analyzed samples were barren. Fram Strait core PS1535 has two intervals with abundant planktic foraminifers in the coarse fraction (ages 165–152 and 140–134 ka) and a minor abundance peak from ca 173 ka which might correspond to a similar minor peak in PS2200 (ca 171 ka). The abundance pattern in sediments from OIS 5 typically shows three peaks that correlate to substages 5e, 5c, and 5a and have corresponding peaks in the ^{10}Be records (cf. Fig. 6). The pattern is best seen in PS2185 and PS51/038-4. On the Morris Jesup Rise (PS2200) sedimentation rates were extremely low in middle OIS 5 and only the peaks from substages 5e and 5a are obvious. On the Yermak Plateau (site PS1533), sediments from OIS 5e are almost barren, possibly due to carbonate dissolution. However, a high ^{10}Be content at ca 390 cm in this core (Fig. 6) may be indirect evidence of strongly increased planktic productivity during substage 5e. In Fram Strait core PS1535 the planktic foraminifer content in OIS 5 sediments is relatively low except in substage 5a, although peaks corresponding to OIS 5e and 5c can be clearly identified. Such low calcium carbonate values are a typical feature in sediments from lower and middle OIS 5 sediments from the Fram Strait (Hebbeln and Wefer, 1997). At the boundary between OIS 5 and 4, a planktic foraminifer abundance peak is found in cores PS1535, PS2200, and PS51/038-4. For site PS2185, a corresponding enhanced planktic productivity can be inferred from a ^{10}Be double peak at 120–145 cm in this core. Foraminifers in this layer may be missing due to carbonate dissolution. A nearly barren interval is found in all Arctic Ocean cores, representing the time span 64–50 ka. Samples from <50 ka contain abundant foraminifers in cores PS2200 and PS51/038-4, but sedimentation rates were extremely low (<0.5 cm/ka) and impede a paleoenvironmental interpretation (cf. Nørgaard-Pedersen et al., 2003). PS2185 has a higher resolution and contains a well-defined planktic foraminifer peak from 50 to 35 ka and high abundances in Holocene sediments. In the Arctic Gateway cores, sediments from 60 to 35 ka contain very little planktic foraminifers. PS1533-3 has high abundances in sediments from around 20 and 26 ka, while PS1535 additionally contains a peak at around 33 ka.

All analyzed cores contain abundant planktic foraminifers in Holocene sediments.

6.2. Ice-rafted debris

Comparison of the grain-size records (wt% >63 μm) with the planktic foraminifer abundance records (Fig. 7) shows for most samples from the central Arctic an anticorrelation between both parameters. A "background" coarse-fraction content of ca 10 wt% is typical for these cores, probably resulting from winnowing effects on the fines which are evident on the Lomonosov Ridge and the Morris Jesup Rise (Bergmann, 1996). Coarse-grained layers in sediment cores from these sites (PS2185, PS2200) show an elevated amount of the clay mineral smectite in the fine fraction (Fig. 7). In a few samples (marked by asterisks in Fig. 7) from the Arctic Gateway cores the coarse fraction consists to a large amount (up to 90%) of biogenic particles. In the following, we discuss the grain-size records of our cores only for the variability of the non-biogenic (i.e., probably iceberg-rafted) coarse fraction.

A high content of coarse IRD in sediments from OIS 6 is characteristic for all analyzed cores. According to our stratigraphic models, strong input of IRD in the central Arctic started already in late OIS 7, directly following OIE 7.1. Here, IRD contents remain on the same level of 20–35 wt% throughout the sediment column from OIS 6, with the exception of a finer-grained layer from ca 155±10 ka. The lack of well-constrained stratigraphic tie points within OIS 6 makes it difficult to define the exact age of this interval. It is possible that it corresponds to the change from medium to high IRD contents in the Arctic Gateway cores at 160–155 ka. A distinct change from coarse- to finer-grained sediments is noted at the stratigraphic boundary of OIS 6 and 5 in all analyzed cores. Remarkably little change in grain sizes is found in the central Arctic Ocean cores for OIS 5e to 5c. In PS2200 and PS2185, three coarse layers are found, which were deposited at ca 90–85, 77–73, and 65–50 ka. Because core PS2178 from the deep Makarov Basin, where winnowing effects are less likely, has a very similar succession of sandy layers at 300–170 cm (Fig. 5), there is every indication for an ice-rafted origin of the coarse sediments. The youngest of these layers is an outstanding feature also in many other cores from the central Arctic (cf. Pak et al., 1992; Nørgaard-Pedersen et al., 1998), including 96/12-1pc (Jakobsson et al., 2000) and PS51/038-4. In the latter core, the coarse layer from ca 75 ka is only weakly developed and sediments from ca 85 ka show no distinct grain-size change in both these cores. Central Arctic Ocean sediments from <50 ka show an upward decrease of coarse-fraction content and a minimum in the deposits from the LGM around 20 ka (Nørgaard-Pedersen et al., 1998, 2003).

With the exception of one sample from the last deglaciation in Yermak Plateau core PS1533-3, coarse-fraction contents in Weichselian sediments in the Arctic Gateway cores never reach the OIS 6 values (Fig. 7). Elevated values are found in sediments from OIS 5d, late OIS4/early OIS 3 and the last deglaciation (ca 15–12 ka). In Fram Strait core PS1535, the Late Holocene sediments also have a relatively high IRD content.

6.3. Evidence for freshwater events from stable oxygen and carbon isotopes of planktic foraminifers

Planktic stable oxygen and carbon isotope records from long Arctic Ocean sediment cores usually show much more scatter than deep-sea records from low- and mid-latitude oceans. This fact can be explained by the proximity to Late Quaternary ice sheets. Abrupt releases of freshwater with a strongly different isotopic signature compared to ocean water and the establishment of a low-salinity freshwater lid in the ocean, which subsequently inhibited vertical water mass exchange, are reflected in the isotopic records as excursions of variable amplitude, duration, and regional extent. Our records show many such freshwater peaks, which are identified by both low oxygen isotope and low carbon isotope values in the same or the consecutive sample (Fig. 8). Amplitudes can reach 2‰ in δ^{18}O, but usually are 0.5–1.0‰. Small amplitudes of <0.3‰ are not considered here.

It is obvious from our records that the freshwater events did not occur randomly at the various sites within the last 200 ka. Many of the isotopic excursions were recorded at several, if not all sites during defined time intervals at 160–155, 140–125, 90–75, 65–60 ka, and at ca 50 ka. All analyzed cores contain sediment sequences which are totally barren of planktic foraminifers, but which coincide with intervals where freshwater spikes were found in other cores. One possibility for the lack in foraminifers is carbonate dissolution at the seafloor. However, it cannot be excluded that freshwater discharge from the continents decreased salinity in the upper Arctic Ocean below the tolerance limit of planktic foraminifers. The exact limit for *N. pachyderma* (sin.) is not known, but culturing results on *N. dutertrei* (Bijma et al., 1990) showed that shell growth stopped below an average salinity of 28. Since *N. pachyderma* (sin.) is from the same genus, a similarly low-salinity limit can be expected for this species.

A number of other freshwater events can be detected in our records outside the above-mentioned intervals. Especially, the record from Fram Strait core PS1535 shows a number of spikes between 50 and 30 ka. Both Arctic Gateway cores have a foraminifer–barren interval between the LGM and the Holocene. Resolution in the central Arctic Ocean cores again is too low for the last

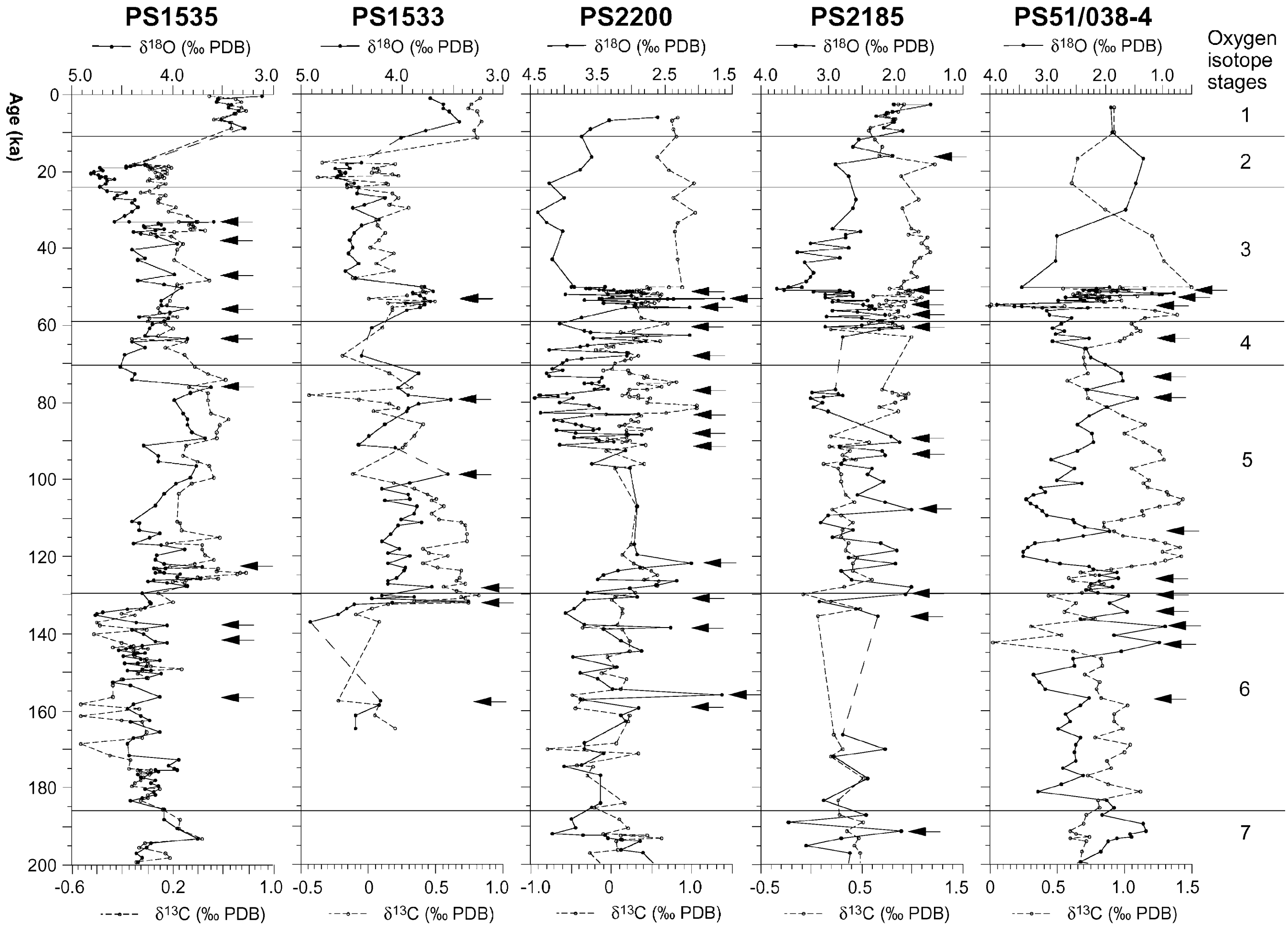

Fig. 8. Planktic oxygen and carbon isotope records of Arctic sediment cores for the last 200 ka. Arrows mark identified freshwater events characterized by low values of both parameters.

50 ka to allow the identification of clearly outstanding freshwater spikes.

7. Arctic Ocean paleoenvironments and the Late Quaternary glaciation history of Northern Eurasia

7.1. General paleoenvironmental scenarios

The records of IRD input, planktic foraminifer abundances, and planktic isotope values from the last 200 ka clearly demonstrate that sedimentary environments in the eastern and central Arctic Ocean were strongly variable. In particular, anticorrelation of IRD and foraminifer abundances in the central Arctic cores suggest that there were two major scenarios.

The first one is represented in the cores by the foraminifer-rich intervals which reflect times with an inflow of Atlantic Water (Hebbeln et al., 1994; Dokken and Hald, 1996; Nørgaard-Pedersen et al., 2003) of variable strength, temperature, and regional extension. Today, the North Atlantic and the Nordic Seas are a major moisture source for Northern Eurasia (Kuznetsova, 1998; Rinke et al., 1999). Using this modern situation as an analog, it can be inferred that the Atlantic Water inflow in the past in a similar way occurred along with an atmospheric circulation pattern which supplied atmospheric heat and moisture to the Arctic. Moreover, the resulting open water areas must have been an additional source for evaporation over much of the eastern and central Arctic Ocean.

The second major scenario is represented in the cores by the coarse-grained layers which contain very few or no foraminifers. The sedimentary processes must have been dominated by iceberg transport of terrigenous material. Elevated smectite concentrations in the clay fraction of the coarse-grained layers (Fig. 7) serve as a tracer for the provenance of the IRD. The dominant sources of smectite in the Arctic Ocean area are the southeastern and eastern Kara Sea and the western Laptev Sea where smectite concentrations in the clay fraction of the surface sediments reach >40%

(Nürnberg et al., 1994; Stein et al., 1994c; Wahsner et al., 1999). It should be noted that another clay mineral (kaolinite) has a major source in the Barents Sea and that smectite-rich layers in sediment cores from the eastern and central Arctic Ocean are always enriched in kaolinite (Spielhagen et al., 1997; Vogt, 1997). This indicates the coincidence of erosional processes in both source areas. The (probably iceberg-rafted) heavy mineral assemblage in the coarse-grained sediments of PS2185 also suggests a provenance from the Kara Sea and partly from the Laptev Sea (Behrends, 1999). There is ample evidence for Late Quaternary glaciations on the Barents and Kara seas shelves and in the partly mountainous hinterland (Svendsen et al., 1999, 2004). We will use the occurrence of smectite-rich IRD layers in the Arctic deep-sea sediment cores as evidence for a paleogeographic situation when large ice sheets covered these shelves. Eustatic sea-level fall of more than 50 m during these glaciations (Lambeck and Chappell, 2001) placed the coastline near the present shelf break where icebergs could break off and easily escape into the Arctic Ocean. Using this framework of general paleoenvironmental scenarios in the eastern and central Arctic Ocean, we will discuss the evidence from Arctic deep-sea sediments for the variable paleoenvironments in the eastern Arctic Ocean and the history of ice sheets on Northern Eurasia during the past 200 ka (late OIS 7 to present).

7.2. *Saalian glaciation in oxygen isotope stage 6 (186–130 ka)*

In our cores, the high IRD content is the most conspicuous feature of sediments from OIS 6 (Fig. 7), which is suggested to be roughly equivalent to the Warthe stage of the Saalian glaciation in Northwest Europe (Ehlers et al., 1991). According to our stratigraphic models for the central Arctic cores, the strong IRD input started already in latest OIS 7 between 195 and 190 ka and was maintained for the entire stage (with the exception of 160–150 ka at site PS51/038). The evidence lies in the fact that the IRD-rich sediment package rests directly on a foraminifer-rich layer, which is identified as representing OIE 7.1. Regardless of the exact age of the lower boundary of the IRD-rich package, the data suggest that the strong IRD input immediately followed a time of Atlantic Water inflow and seasonally open waters in the central Arctic Ocean during late OIS 7. This Atlantic Water inflow was probably accompanied by an atmospheric moisture supply to Northern Eurasia which supported the build-up of an ice sheet.

Compared to the early onset of IRD deposition in the central Arctic, central Fram Strait sedimentary records document a delay of the enhanced IRD input. As shown earlier by Hebbeln and Wefer (1997), major iceberg rafting in the Fram Strait in OIS 6 started only after 180 ka. It is difficult to explain the diachronous onset in the central Fram Strait (PS1535) and central Arctic records, because the Fram Strait was the only outlet for large icebergs from the Arctic Ocean in glacial times. Possibly, the southward export of first IRD-laden icebergs in earliest OIS 6 occurred mainly through the western Fram Strait, close to the East Greenland shelf, and less icebergs drifted through the central Fram Strait. This hypothesis is corroborated by the grain-size distribution in Fram Strait cores. Similar to the central Arctic Ocean cores, where the maximum IRD content level (30–40 wt% $>63\,\mu$m) is reached already in sediments from early OIS 6 and is maintained up to the OIS 6/5 boundary, cores from the western Fram Strait show a stable, but high coarse-fraction content (cf. Hebbeln and Wefer, 1997). On the other hand, in cores from the central and eastern Fram Strait as well as in PS1533-3 from the Yermak Plateau, a distinct upward coarsening can be noted in sediments from ca 160 to 155 ka above a layer with high abundances of planktic foraminifers and coccoliths (Hebbeln and Wefer, 1997; this study). Our data from PS1535 and PS1533-3 support the conclusions of Hebbeln and Wefer (1997) of an interval of Atlantic Water inflow to the Fram Strait around 160 ka. This inflow and the induced supply of moisture to the Arctic could have facilitated a further growth and extension of ice sheets over Northern Eurasia. A final inflow event in OIS 6 is recorded in the Arctic Gateway cores in sediments from the penultimate glacial maximum (OIE 6.2, 135 ka). This section, which is unequivocally identified by highest oxygen isotope values of >4.7‰, is marked by a minimum in the coarse-fraction record caused by extremely little coarse IRD but plenty of planktic foraminifers. A minimum in IRD contents and an associated (small) peak in planktic foraminifer abundances can also be found in late-OIS 6 sediments from the Yermak Plateau (PS1533-3) and the Lomonosov Ridge (PS2185) and are correlated to OIE 6.2. The same combination of sedimentological features was found also in other Fram Strait cores (Hebbeln and Wefer, 1997), but the lack of sufficient age control within OIS 6 led to the determination of an older age (145 ka) for the corresponding inflow event of Atlantic Water. Considering the extension of the Svalbard ice sheet to the western shelf edge during late OIS 6 (Mangerud et al., 1998), we interpret the low IRD contents in OIE 6.2 sediments as evidence for extreme environmental gradients in the western and central Fram Strait at that time. If any, only very few sediment-laden icebergs may have reached the open water area caused by the Atlantic Water inflow. Again, the open waters in the Fram Strait and further south may have served as moisture sources for a (final) growth and extension of ice sheets over Northern Eurasia.

The elevated smectite contents in the IRD-rich sediments from OIS 6 in central Arctic cores PS2185 and PS2200 point to the eastern Kara Sea/western Laptev Sea as a main source area for icebergs. This source must have been active continuously for approximately 60 ka, which implies a huge ice sheet in the main source area, because it seems unlikely that a regional ice cap could supply such large amounts of IRD for a long time interval. Recent reconstructions by Astakhov (2004) and Svendsen et al. (2004) showed that the OIS 6 ice sheet over Northern Eurasia was the largest within the last 200 ka (Fig. 1). The authors propose an ice sheet extent over the entire Kara Sea to the shelf break and across the Severnaya Zemlya archipelago. High IRD abundances in late OIS 6 sediments in core PS2138-1 from the northern Barents Sea continental margin (995 m water depth) support the hypothesis of a large-scale glaciation on the shelf (Knies et al., 1999, 2000, 2001). On the other hand, core PS2741-1 from the continental margin NE of Severnaya Zemlya (2530 m water depth) contains very little coarse IRD in OIS 6 sediments (Knies et al., 2000). The age of the core base is assumed to lie in OIS 6 (Knies et al., 2000; Matthiessen et al., 2001) and the latest estimate puts it at ca 180 ka (Knies et al. (2001). Because the OIS 6 sediments are fine-grained and contain coarse IRD only in the very oldest and youngest sections, Knies et al. (2001) speculate that ice sheet fluctuations were either small, or glacimarine conditions were stable at times when ice sheets had a small extent. From onshore or shelf sites very little field information is available about the extent of the Saalian ice sheet over Severnaya Zemlya (cf. Arkhipov et al., 1986; Astakhov, 2004). Shallow seismic data suggest an ice sheet progradation to the shelf break prior to OIS 4 (Weiel, unpublished data). To reconcile the continental margin data (Knies et al., 2000, 2001) with our results from the central Arctic Ocean, which imply large amounts of icebergs from a relatively stable, productive ice margin, we speculate that the ice sheet over Severnaya Zemlya did not reach to the shelf break NE of the archipelago during most of OIS 6. The fine-grained nature of the sediments in PS2741-1 may result from subglacial meltwater outflow and the deposition of suspended material on the slope. The IRD layer from the deglaciation probably represents a deposit from a final ice advance to the shoreline or the time of early deglacial sea-level rise.

The wide eastward extent of the OIS 6 ice sheet over northern Siberia blocked the northward drainage of the Ob and Yenisei rivers and led to the development of a huge lake system south of the ice sheet (Arkhipov et al., 1986, 1995). Lakes in the north of European Russia were probably smaller than in the Weichselian because in most areas the ice advanced to the north–south watershed (Arkhipov et al., 1995; Astakhov, 2004). The planktic isotope records from our sediment cores (Fig. 8) show clear evidence for one or more strong freshwater events in the Arctic Ocean during the deglaciation between 130 and 140 ka. Amplitudes of $\delta^{18}O$ changes reach 1.5‰ and are associated with equivalent changes in $\delta^{13}C$, indicating the expansion of an unusually thick freshwater layer over the entire eastern and central Arctic Ocean and a strongly reduced ventilation of the upper water mass. We propose that the freshwater was released from a dammed lake south of the ice sheet when a northern pathway for the water opened in the late glacial and/or the deglaciation. From our stratigraphic database it is either difficult to estimate the exact number and duration of the freshwater event(s), because they apparently vary from site to site. Fram Strait core PS1535 reveals an event at ca 138 ka, before the glacial ($\delta^{18}O$) maximum (OIE 6.2, 135 ka). Planktic isotope records from PS1533-3 and PS2185 indicate one freshwater event before OIE 6.2 and one thereafter (132–130 ka). Based on these records we propose two strong freshwater events in late OIS 6, separated by a time of Atlantic Water inflow to the Fram Strait around 135 ka, which had some minor influence also in the central Arctic Ocean. It can be speculated that the moisture supply associated with the Atlantic Water inflow may have introduced a final short-term expansion of the ice sheet on the Kara Sea and further to the south, thereby closing the northern outlet of the lake.

Evidence for an earlier freshwater event at ca 155 ka comes from the planktic isotope records of the Arctic Gateway cores and PS2200. The record from PS51/038-4 shows a minor spike for this time, while PS2185 has no planktic foraminifers in sediments from middle OIS 6. At the Gateway sites, the event fell into a time of Atlantic Water inflow (indicated by abundant planktic foraminifers in the sediments, cf. Fig. 7), which preceded the interval of enhanced coarse-grained IRD deposition discussed above. The origin of the freshwater, which produced the spike in the isotope records, remains enigmatic. It may result from a regional glacial surge and meltwater event in the Arctic Gateway area introduced by a mid-OIS 6 ice sheet growth. An alternative could be a closure and reopening of the glacial outlet for the rivers in northern Siberia. Our present data do not allow a closer determination of the freshwater source.

Recently, erosional features on top of the Lomonosov Ridge (above ca 1000 m water depth) have been detected (Jakobsson, 1999) which were interpreted as evidence for ice grounding and the development of a floating ice shelf over part of the eastern Arctic Ocean in the Late Quaternary (Polyak et al., 2001). Analysis of several sediment cores from below and above the erosion level on the ridge suggests the later part of OIS 6 as the timing of this event (Jakobsson et al., 2001). All our sediment cores are from water depths below the erosional unconformity. They are not overconsolidated by a grounding ice sheet and should thus, following Jakobsson et al. (2001), represent undisturbed records of sedimentation in OIS 6. The data sets, however, let us neither support nor reject the ice shelf hypothesis of Polyak et al. (2001). There is no modern analog for a ca 1000 m thick, floating ice shelf which crossed a ca 500 km wide deep-sea basin and we can only speculate about the sediment type deposited under the floating ice. Such sediments must be free of planktic organisms because those live in the uppermost euphotic zone of the water column. The high IRD contents in OIS 6 sediments are a feature which may result from sediment transport at the underside of the ice. Core PS51/038-4 from the Alpha Ridge contains low amounts of planktic foraminifers throughout the OIS 6 sediment sequence, which indicates low, but continuous bioproductivity at this site. Thus, the Alpha Ridge may not have been reached by the ice shelf. Cores PS2185 and PS2200 contain foraminifer-free intervals in the OIS 6 section, but none of these layers shows any special features or peaks in the grain-size distribution or clay mineral composition. Erosional features on the crest of the outer Morris Jesup Rise (Fig. 9) have been interpreted earlier as current erosion channels (Bergmann, 1996). Their shape resembles those from the Lomonosov Ridge (Jakobsson, 1999) and the Yermak Plateau (Vogt et al., 1994) and some of them are up to 10 m deep. Like on the Lomonosov Ridge, the occurrence of the erosional surface features is limited to water depths <1000 m (i.e., above site PS2200). We interpret these features as the result of ice grounding and tentatively correlate their age to the ice grounding event on the Lomonosov Ridge in OIS 6. The available PARASOUND records do not allow a closer determination of the processes which formed the erosion channels. It remains to future research in the area to find out whether an ice shelf was present over the Morris Jesup Rise or if the channels developed from huge, deep-keeled drifting icebergs or ice islands. The latter may have been fragments of the proposed ice shelf in the eastern Arctic Ocean during OIS 6.

7.3. Early and middle oxygen isotope stage 5 (130–95 ka)

Compared to the underlying Saalian section, central Arctic Ocean sediments from oxygen isotope substages 5e to 5c contain only minor amounts of IRD. Smectite contents in PS2185 and PS2200 are also on a

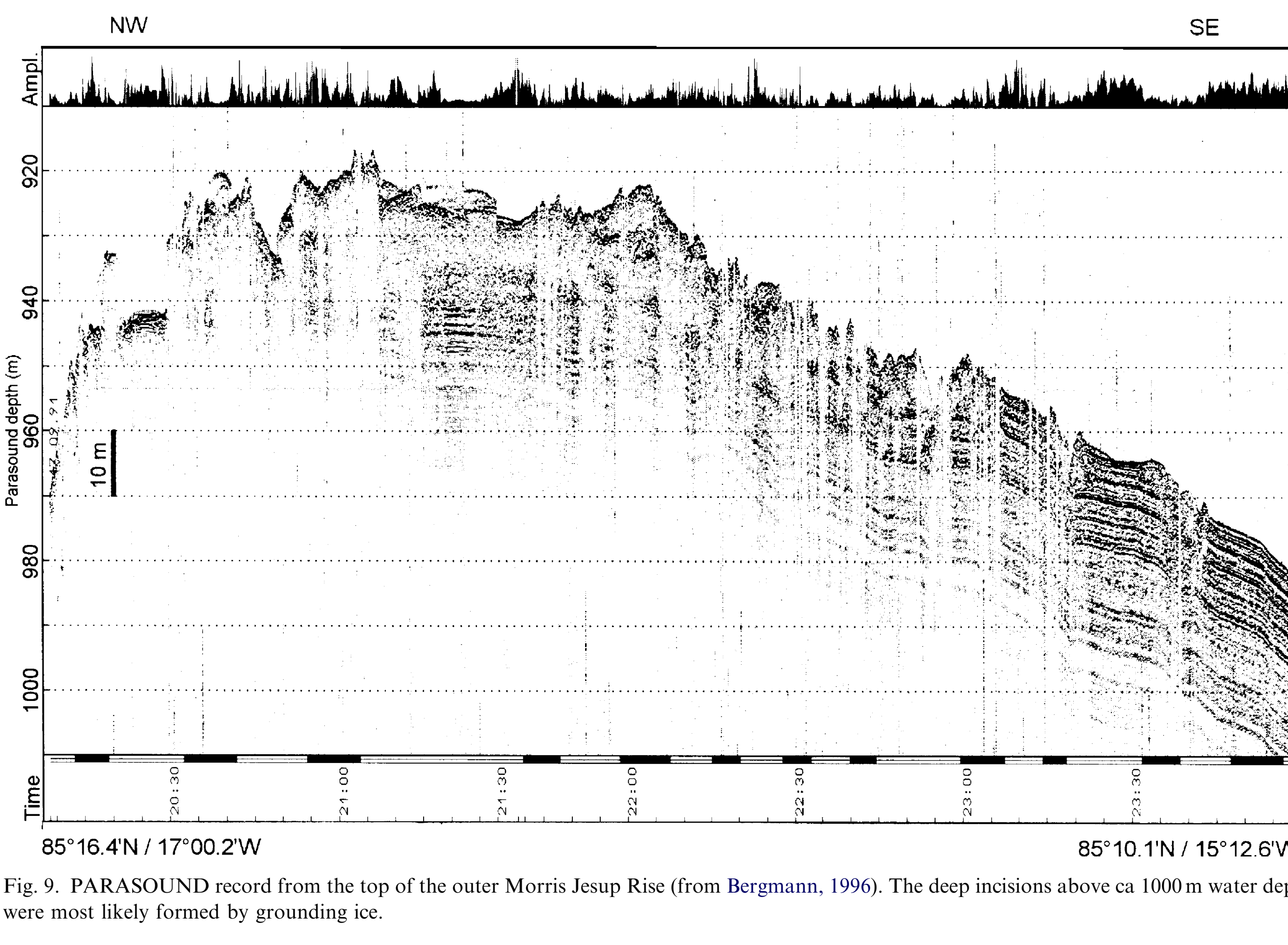

Fig. 9. PARASOUND record from the top of the outer Morris Jesup Rise (from Bergmann, 1996). The deep incisions above ca 1000 m water depth were most likely formed by grounding ice.

background level. Coarse-fraction contents are lower than for the surface samples which reflect the average environmental situation of the last ca 3 ka. Considering the signs of winnowing on the ridges (Bergmann, 1996) which may explain the background level of 5–10% coarse fraction there, we see neither evidence for a large number of icebergs in the eastern and central Arctic Ocean nor for a large-scale glaciation in the potential source areas on the Eurasian shelves. The high planktic foraminifer abundances in Eemian sediments (OIS 5e, ca 130–115 ka) in all cores except PS1533-3 reveal a strong inflow of Atlantic Water to the Arctic Ocean. This is corroborated by findings of coccoliths (Baumann, 1990; Gard and Backman, 1990), dinocysts (Matthiessen and Knies, 2001; Matthiessen et al., 2001), and benthic foraminifers (Wollenburg et al., 2001) in sediment cores from the Arctic Gateway and the northern Barents Sea margin. Onshore marine sediments from many sites south of the present Arctic coast, deposited during the post-glacial period of strong isostatic depression, contain boreal molluscs presently associated with Atlantic Waters and indicative of temperatures in the early Eemian higher than at present (cf. Svendsen et al., 2004, for a review of relevant references). Taken together, we consider the Eemian environments in the eastern and central Arctic Ocean as a typical interglacial situation.

High planktic foraminifer abundances in PS2185, PS51/038-4, and 96/12-1pc (cf. Backman et al., 2004), as well as a corresponding minor peak in the PS1535 record, are evidence for a second inflow event in OIS 5c (around 100 ka). Coccolith, dinocyst and benthic foraminifer abundances are lower than in OIS 5e sediments, indicating that the Atlantic Water inflow was either weaker or the ice cover was somewhat denser. Between the two foraminifer-rich layers, a foraminifer minimum in PS2185 and PS51/038-4 reflects a time of climatic deterioration corresponding to OIS 5d in the central Arctic Ocean. Incoherent single freshwater spikes are found in the isotope records from these cores at ca 110 ka (Fig. 8), but from the available data we cannot determine the source of the freshwater. Knies et al. (2001) report an IRD layer in sediment core PS2741-1 from NE of Severnaya Zemlya, which they suppose to reflect an ice advance from the islands onto the shelf in OIS 5d. Some meltwater from this source may have reached the central Arctic Ocean, but there is no other evidence for the proposed ice advance in our cores.

The Arctic Gateway cores show minor coarse-fraction peaks in post-Eemian sediments, which indicate somewhat more deposition from icebergs. Other cores from the Fram Strait and the northern Barents Sea margin have similar features (Hebbeln and Wefer, 1997; Knies et al., 2001). A connection between the first Weichselian ice sheet advance onto the shelf off Svalbard in OIS 5d (Mangerud et al., 1998) and time-equivalent IRD occurrences on the northern Barents Sea margin was suggested by Knies et al. (2001). It probably explains also the IRD peaks in other cores from the Arctic Gateway. The freshwater spike at 100 ka in the isotope record of PS1533-3 may result from the subsequent deglaciation. Obviously, central Arctic sedimentation was not affected by the OIS 5d ice advance in the Svalbard/Barents Sea area.

7.4. Early Weichselian glaciation in oxygen isotope substage 5b (95–80 ka)

Several cores from the central Arctic contain an IRD-rich layer in sediments from OIS 5b (ca 90–80 ka, Fig. 7). Its high smectite content again indicates an origin of the icebergs from the Kara and Laptev seas region and, accordingly, a large-scale glaciation on the shelf and in the hinterland. Although the data from the underlying OIS 5c sediments do not offer fully consistent evidence for open water conditions in the Arctic Ocean, a strong Atlantic Water inflow, and associated atmospheric moisture supply, we favor such a scenario because of the high planktic foraminifer abundances in several cores.

The thickness of the IRD-rich OIS 5b layer is variable from core to core. At site PS51/038 on the Alpha Ridge, which is most remote from the margin, no peak in the grain-size record from OIS 5b is found. Its presence in 96-12/1pc from the Lomonosov Ridge is doubtful and other cores from this area (PS2178, PS2185) contain a thin, coarse layer only. In PS2200 from the Morris Jesup Rise, however, the IRD layer is >20 cm thick. From the regional differences we conclude that most of the icebergs probably drifted through the Eurasian Basin, closer to the continental margin than in OIS 6. The IRD-rich layers from 85 to 80 ka in cores from NE of Severnaya Zemlya and the northern Barents Sea margin (Knies et al., 2001), as well as the southern Yermak Plateau (PS2122-1; Vogt et al., 2001) may support this hypothesis, although IRD contents are not exceptionally high, if compared to older (OIS 6) and younger (OIS 4 and 3) sediments. In contrast to the earlier glaciation in OIS 6, our Arctic Gateway cores show only little evidence for intense iceberg rafting. PS1533-3 from the Yermak Plateau has a minor IRD peak in sediments from ca 90 ka, and in PS1535 from the central Fram Strait the middle OIS 5 coarse-fraction record has its maximum already in OIS 5c (ca 105–100 ka). Other cores from the Fram Strait have medium high IRD contents, but no significant peak in OIS 5b sediments (Hebbeln and Wefer, 1997).

Terrestrial mapping and OSL dating of tills within the QUEEN program made clear that a large-scale glaciation had covered the northernmost parts of European Russia and Siberia north of approximately 67°N in OIS 5b, reaching eastward to 100°E and covering a large part of Taymyr Peninsula (Svendsen et al., 2004; Fig. 1). Little is known from the various archipelagos along the Eurasian continental margin about the northern extent of the ice sheet. Data from Svalbard suggest a short-term (<10 ka) ice advance to the west coast and in places to the shelf edge (Mangerud et al., 1998). The deep-sea results from central Arctic Ocean and Eurasian continental margin cores generally support a wide ice sheet extension over the Barents and Kara seas shelves. The limited thickness and areal distribution of the corresponding IRD layer may have resulted from relatively low iceberg calving rates. Possibly, the coastline along the continental margin was reached in many areas only during a lesser part of the glaciation. A northward flow of Siberian ice to the Arctic Ocean mainly through the St. Anna Trough in the NW Kara Sea and a westward ice drift in the Eurasian Basin may explain the observed lower IRD contents on parts of the Lomonosov Ridge area and on the Alpha Ridge.

The relatively low IRD contents in OIS 5b sediments in Arctic Gateway cores (Hebbeln and Wefer, 1997; own data) are hard to reconcile with a drift of many icebergs out of the Arctic Ocean. The simplest explanation would be melting of the icebergs already inside the Arctic Ocean because of a temperate inflow of Atlantic water. Available microfossil data from Fram Strait sediment cores to detect such an Atlantic Water inflow give non-uniform results. Our Arctic Gateway cores have minor (PS1535) or no (PS1533-3) planktic foraminifer contents in OIS 5b sediments, whereas coccolith abundances in these cores are high (Baumann, 1990; Nowaczyk and Baumann, 1992). Abundant coccoliths in OIS 5b sediments were also reported from other Fram Strait sites (Hebbeln and Wefer, 1997). On the other hand, Gard and Backman (1990), on the basis of coccolith data from 68 sediment cores from the Nordic Seas and the Arctic Gateway, proposed a restriction of Atlantic Water inflow to the central Norwegian Sea. At present, it seems impossible to resolve this contradiction which may be caused in part by stratigraphic uncertainties for many of the cores.

Like during the northern Eurasian glaciation in OIS 6, an ice sheet over the Barents and Kara seas must have blocked the outflow of rivers. Terrestrial lake deposits in many places prove the formation of large lakes (e.g., Astakhov et al., 1999; Houmark-Nielsen et al., 2001; Mangerud et al., 2001, 2004) in northernmost Russia west of 90°E. The planktic foraminifer isotope records from our cores display several peaks in the interval 90–75 ka (Fig. 8), which are supposed to reflect freshwater events caused by the northward drainage of these lakes. The spikes are best seen in the records from PS1535, PS1533-3, and PS2200. Lomonosov Ridge core PS2185 has no foraminifers in the thin IRD-rich layer from OIS 5b, possibly caused by extremely low salinities (<28) in the central Arctic Ocean. This hypothesis may be corroborated by a time-equivalent foraminifer–barren zone in core PS2138-1 (Knies et al., 2000) from the northern Barents Sea margin. On the other hand, we cannot exclude that the spread of freshwater from lake drainage was restricted to the Eurasian Basin, similar to the proposed iceberg drift path within OIS 5b.

7.5. Late oxygen isotope stage 5 and early stage 4 (80–65 ka)

An interval of Atlantic Water inflow to the Arctic Ocean in OIS 5a and early OIS 4 is documented in very many cores from the Arctic Gateway and the central Arctic. As discussed above, some authors propose that enhanced coccolith production started already in OIS 5b. Gard (1986, 1987, 1988) found very high abundances of coccoliths in OIS 5a–4 sediments in a large number of analyzed cores from the Fram Strait and the adjacent Nordic Seas and Arctic Ocean. It was concluded that the inflow of Atlantic Water in OIS 5a reached north of 80°N and that there was relatively little dilution by terrestrial sediments (Gard and Backman, 1990). Several further microfossil studies gave similar results (Baumann, 1990; Nowaczyk and Baumann, 1992; Matthiessen and Knies, 2001; Matthiessen et al., 2001; Wollenburg et al., 2001). Peaks in the planktic foraminifer abundance records of our central Arctic cores show that part of the Atlantic Water reached the eastern and central Arctic Ocean, where seasonally open water conditions were established. Possibly, in the early part of OIS 5a the associated atmospheric transport of relatively warm air supported the deglaciation of the Barents and Kara seas shelves at ca 85–80 ka, at a time when northern high-latitude insolation was at a maximum (cf. Berger and Loutre, 1991). With the exception of one sample, measured smectite contents in PS2185 remain on a high level in sediments from OIS 5b–4. This indicates that the (north)eastern Kara Sea/western Laptev Sea area remained a dominant source for central Arctic Ocean sediments, although iceberg rafting of coarse particles was not a major transport process in most of OIS 5a. We explain the high smectite contents by sea ice transport from the isostatically depressed Kara Sea, which was partly flooded subsequent to the decay of the OIS 5b ice sheet (Möller et al., 1999; Svendsen et al., 2004). A similar post-glacial peak in smectite and kaolinite content can be found in the lowermost Eemian sediments of PS2185 at 246 cm (cf. Spielhagen et al., 1997).

A consistent feature in all analyzed central Arctic Ocean sediment cores (including PS2178 and 96-12/1pc) is an IRD-rich layer deposited around the OIS 5/4 boundary (Fig. 7). With the exception of PS51/038-4, where this layer is less well developed, the IRD-rich layer always has a fining-upward character and a sharp boundary to underlying more fine-grained sediments.

This indicates a rapid onset and gradual decrease of iceberg rafting in the central Arctic Ocean. High smectite contents in the IRD layer in PS2185 and PS2200 point to the eastern Kara Sea as the main source area for the icebergs. Core PS2741-1 from NE of Severnaya Zemlya has no significant IRD contents in sediments from 80 to 70 ka (Knies et al., 2001), which proves an extended glaciation in westernmost Laptev Sea unlikely. The IRD record of core PS2138-1 from the northern Barents Sea margin shows medium high values in a layer from 75 to 70 ka, interpreted as evidence for the build-up of a Middle Weichselian glaciation on Svalbard (Knies et al., 2000, 2001). Considering the ample evidence for strong Atlantic Water inflow and open waters in the Fram Strait in OIS 5a which were an important moisture source, we suppose that an associated atmospheric moisture transport with westerly winds gave way for an early mid-Weichselian formation of ice domes. Ice sheet growth on the Barents and Kara seas may have been supported by the rapid decrease of northern high-latitude insolation which reached a minimum around 72 ka (Berger and Loutre, 1991). Since IRD contents in Fram Strait and Yermak Plateau cores are low (cf. Hebbeln and Wefer, 1997), melting of the icebergs derived from northern Eurasia probably occurred already within the Arctic Ocean.

In contrast to the intervals of strong IRD input in OIS 6 and 5b, no terrestrial evidence for an extensive glaciation in northern Eurasia between Scandinavia and the Taymyr Peninsula at the OIS 5/4 boundary has been found yet. Latest reconstructions propose that a large ice sheet disappeared at ca 80 ka and grew again around 65 ka (Svendsen et al., 2004). Our stratigraphy of the deep-sea cores seems exceptionally well-constrained for the OIS 5a interval, based on findings of coccoliths that can be correlated to many other cores in the Nordic Seas. Therefore, we strongly believe that our arguments for some glaciation in the potential source area of smectite and kaolinite at the OIS 5/4 boundary are well substantiated. However, we can only speculate about details of the paleogeographical situation in that area. Possibly, a number of smaller ice domes existed which were confined to the various archipelagos on the Barents and Kara seas shelves and supplied icebergs to the Arctic Ocean. The planktic foraminifer isotope records do not show compelling evidence for strong freshwater events and we suppose that any ice sheet that might have existed on the Kara Sea shelf around 75 ka did not block the outflow of river water.

Above the IRD-rich layer from the boundary of OIS 5 and 4, we find again signs for a strong inflow of Atlantic Water to the interior Arctic Ocean. Cores PS1535, PS2200, PS51/038-4, and 96-12/1pc contain a layer with abundant planktic foraminifers, but only little IRD (Fig. 7). As discussed above, foraminifers are missing in this particular layer in PS2185 probably due to carbonate dissolution. Using the high abundances of coccoliths in Fram Strait and Nordic Seas sediments as additional evidence for open water conditions (cf. Gard and Backman, 1990; Hebbeln and Wefer, 1997), it can be inferred that surface waters in the Arctic Gateway and (probably seasonally) in the Arctic Ocean represented an important moisture source for eastward directed atmospheric transport, precipitation, and a regrowth of ice sheets in Northern Eurasia.

7.6. Middle Weichselian glaciation in oxygen isotope stages 4 and 3 (65–50 ka)

A thick IRD-rich layer deposited in the younger part of OIS 4 and oldest OIS 3 is one of the most conspicuous features in sediment cores from the eastern and central Arctic Ocean (Fig. 7). It is not only found in all sediment cores from this area discussed in this study, but also in many other cores analyzed so far (Pagels, 1991; Nørgaard-Pedersen et al., 1998; Jakobsson et al., 2000, 2001; Vogt et al., 2001). The high smectite and kaolinite contents (cf. Vogt, 1997) again suggest an origin from the shelves of the Barents and Kara seas and the western Laptev Sea and an extended glaciation there. In many cases, the layer has a very sharp lower boundary. It indicates that the iceberg rafting started abruptly in the study area of the central Arctic at ca 65 ka, according to our stratigraphy, and we suppose that the onset of ice-rafting was synchronous at all our sites in the area. Data from the Barents and Kara seas continental margins allow us to constrain the regional extent of the ice sheet. Cores PS2138-1 from the northern Barents Sea margin and PS2741-1 from NE of Severnaya Zemlya contain a distinct IRD-rich layer deposited around the OIS boundary 4/3 which was interpreted as evidence for an ice sheet advance onto the outer shelf (Knies et al., 2000, 2001). West of Svalbard, the Svalbard–Barents Sea ice sheet also reached the shelf break during OIS 4 (Mangerud et al., 1998). The iceberg output from this ice sheet to the Arctic Gateway is reflected in the relatively high IRD contents in cores PS1533-3, PS1535 and several others in sediments from 65 to 50 ka (Hebbeln, 1993; Hebbeln and Wefer, 1997, this study). From high-resolution seismic profiles of the western Laptev Sea continental margin, Kleiber et al. (2001) mapped extended stacked debris flow deposits. They concluded that the ice sheet in this area reached to the shelf break in OIS 4 and that this eastern ice sheet extension was larger than in the Late Weichselian (OIS 2). The southern glacial limits in Russia, mapped during field work within the QUEEN program, lie ca 50–100 km south of the present southern Barents Sea coast and on the southernmost Kara Sea shelf (Fig. 1, Svendsen et al., 2004). The resulting ice sheet configuration for 65–50 ka was similar to the ice sheet in OIS 5b (when there was no ice in southern Scandinavia, but ice on the northern part of the West Siberian lowland and on the Taymyr peninsula; cf. Fig. 1). However, the thickness of the IRD-rich layer from the Middle Weichselian glaciation, for which calculated bulk accumulation rates were 2–4 times higher than for the OIS 6 and OIS 5b deposits at the same sites in the central Arctic, indicates a very rapid sedimentation. The ice sheet that supplied the icebergs must have been exceptionally productive in terms of iceberg calving rates and/or sediment load. The relatively high abundances of coccoliths (Baumann, 1990) and planktic foraminifers in PS1535 from the central Fram Strait as well as the high coccolith and $CaCO_3$ contents in cores from the eastern Fram Strait (Hebbeln and Wefer, 1997; Hald et al., 2001) in sediments from entire OIS 4 may indicate that seasonally open water conditions still existed in the Arctic Gateway when the ice sheet over northern Eurasia had already reached much of its full size. Hebbeln and Wefer (1997) propose that in OIS 4 cool temperate Atlantic waters penetrated northward into the Fram Strait. The synchroneity of ice sheet build-up and open water conditions in the Arctic Gateway as a source for atmospheric moisture in OIS 4 again underlines the close connection between oceanic and terrestrial developments. During this glacial phase the moisture supply to Northern Eurasia must have been even stronger than during earlier glaciations in OIS 6 and 5. This is indicated by rapid IRD deposition in the central Arctic, which implies more icebergs and probably a faster ice flow over the glaciated shelves than before. From this study, we cannot determine the mechanism giving rise to the stronger precipitation apparently needed. However, it seems possible that there is an "optimum combination" of Atlantic Water inflow to the Nordic Seas (moisture supply) and terrestrial climatic conditions (temperature and atmospheric pressure gradients), which gives way for exceptionally high precipitation, rapid ice build-up, and fast ice movement on northernmost Eurasia (Spielhagen, 2001). Conditions in OIS 4 may have been close to this "optimum" situation.

The planktic isotope records from all analyzed sediment cores show strong freshwater spikes in sediments deposited during the mid-Weichselian glaciation (Fig. 8). Highest amplitudes in $\delta^{13}O$ are found in foraminifers from the uppermost part of the IRD-rich layer. In the central Arctic cores PS2185 and PS2200, they are associated with a thin finer-grained layer (coarse fraction <20 wt%), which is overlain by the youngest, thin IRD-rich layer (Fig. 7). Both these layers have a significantly lower smectite content than the underlying thick IRD package. The lower coarse-fraction content in the layer with the freshwater spikes may result from dilution of coarse IRD by finer-grained sediments which were flushed off the shelf during rapid freshwater outflow. Because the configuration of the mid-Weichselian ice sheet over the Barents and Kara seas (cf. Svendsen et al., 2004) implies a blocking of river drainage and, again, the formation of large lakes, we correlate the freshwater event(s) with the deglacial northward breakthrough of rivers and lakes. Our stratigraphic models give an age of ca 52 ka for this major event in the Arctic. In sediment cores from the Mendeleyev Ridge (80.3–82.5°N, 175–179°W), Poore et al. (1999) found isotopic evidence for a major freshwater event shortly before 49 ^{14}C-ka (ca 50 cal-ka). They proposed that meltwater originated from an early OIS 3 deglaciation of the Innuitian ice sheet over NW North America. Considering the terrestrial evidence for the Middle Weichselian freshwater lakes and for the deglaciation of the northern Eurasia ice sheet in early OIS 3, it is tempting to correlate the freshwater event on the Mendeleyev Ridge with the developments on the Siberian shelves. The fact that the freshwater spike in the cores analyzed by Poore et al. (1999) is strongest at the northernmost site, which is closest to the glaciated Eurasian margin, may support our hypothesis. On the other hand, minimum $\delta^{18}O$ values in the Mendeleyev Ridge freshwater spikes are even lower than those in our cores, which suggests a closer proximity to the freshwater source. At present it remains open whether there were several (at least two) synchronous freshwater discharges from different sources into the Arctic Ocean at ca 52 ka or if freshwater spread across much of the Arctic Ocean from a northern Eurasian source.

At the same time as in the Arctic, a major freshwater event occurred in the Nordic Seas (Vogelsang, 1990; Baumann et al., 1995; Sarnthein and Altenbach, 1995) and we may speculate about a common origin for the events in both regions. While the origin of other meltwater events in the Nordic Seas (e.g., in the last deglaciation) can be tied unambiguously to the decay of nearby ice sheets (cf., Jones and Keigwin, 1988; Sarnthein et al., 1992), the exact origin of the freshwater event in early OIS 3 in the Nordic Seas remains uncertain. In addition to possible other sources like the decaying Scandinavian and Barents Sea ice sheets, the outflow of freshwater from the Arctic Ocean through the Fram Strait may have contributed significantly to the low-salinity event in the Nordic Seas around 52 ka.

7.7. Late Weichselian glaciation and Holocene (50 ka to present)

Following the major deglacial event in the Arctic around 50 ka, deposition at the deep-sea sites in the central Arctic Ocean was dominated by fine-grained sediments with relatively low contents of coarse IRD (Fig. 7). Late Weichselian sedimentation rates are often <1 cm/ka (Stein et al., 1994a, b; Darby et al., 1997;

Spielhagen et al., 1997; Nørgaard-Pedersen et al., 1998, 2003) and the apparent slow decrease of IRD contents in the cores from 50 to 30 ka may in part result from bioturbational mixing of older, coarse grains into the younger, finer-grained sediments. Contents and calculated fluxes of IRD on the Lomonosov Ridge are 5–10 times lower in sediments from the LGM than in sediments from the Middle Weichselian glaciation (Nørgaard-Pedersen et al., 1998). Since also the smectite content is on the background level, there is no evidence for many icebergs from a possible northern Siberian glaciation in central Arctic Ocean sediments younger than ca 50 ka. In fact, sediments from the LGM on the Lomonosov Ridge have the least coarse-fraction contents within this time span and it has been proposed that icebergs were very rare (Nørgaard-Pedersen et al., 1998, 2003). Sediment core PS2741-1 from the NW Laptev Sea continental margin contains a layer with significant amounts of IRD from 40 ka, but otherwise the deposits are fine-grained and do not bear evidence for an extended Late Weichselian glaciation on Severnaya Zemlya (Knies et al., 2000, 2001). Analyses of high-resolution seismic profiles from the western Laptev Sea and Vilkitsky Strait support this conclusion (Kleiber et al., 2001). Terrestrial work and findings of mammoth remains indicate that glaciers on Severnaya Zemlya were even smaller during the LGM than at present (Raab et al., 2003, and references therein). On the other hand, there is evidence for a Late Weichselian ice advance from the shelf onto the NW Taymyr peninsula (Alexanderson et al., 2001, 2002). The most recent consensus of the majority of scientists working in the area is that the Late Weichselian features on NW Taymyr were formed during a short-term eastward advance of an ice sheet otherwise confined to the Barents Sea and northernmost Kara Sea (Svendsen et al., 2004). Since the freshwater spikes in planktic isotope records from the eastern and central Arctic Ocean (Stein et al., 1994a; Nørgaard-Pedersen et al., 1998, 2003) are much smaller in amplitude than those from the deglaciation around 52 ka (cf. Fig. 8), we conclude that any blocking of Siberian river drainage to the Arctic Ocean in the Late Weichselian was just a brief episode and no major lake formation occurred south of the ice sheet.

The history and areal extent of the Late Weichselian Barents Sea ice sheet have been studied in detail by several authors in the last two decades (for reviews see Landvik et al., 1998; Mangerud et al., 1998; Svendsen et al., 1999, 2004). During its maximum in the LGM it covered the entire Barents Sea (except the coastal area east of ca 44°E). The eastern boundary in the southern Kara Sea was located ca 150 km east of Novaya Zemlya (Svendsen et al., 1999; Polyak et al., 2002). During much of the LGM, glacial ice probably covered the northernmost Kara Sea and the St. Anna Trough (Fig. 1), but left a corridor for river drainage between the ice and Severnaya Zemlya (Polyak et al., 2002; Stein et al., 2002; Svendsen et al., 2004). Deglaciation of the St. Anna Trough started before 13.3 ^{14}C-ka and was completed by ca 10 ^{14}C-ka (Polyak et al., 1997). Sediment cores with IRD layers and seismic data from the northern Barents Sea margin indicate that the Late Weichselian Barents Sea ice sheet reached the continental margin between Svalbard and Franz-Josef-Land already at ca 27 ka (23 ^{14}C-ka, Kleiber et al., 2000; Knies et al., 2000, 2001), whereas this approach west of Svalbard was delayed to ca 23.3 ka (19.6 ^{14}C-ka, Andersen et al., 1996; Landvik et al., 1998). High amounts of IRD with typical Mesozoic rocks from the Barents Sea in sediments from 50 to 35 ka in core PS1535 indicate that a minor glaciation may have existed in the NW Barents Sea area prior to the main ice sheet build-up in latest OIS 3 (Spielhagen, 1991). Since core PS1533-3 from the Yermak Plateau contains little to no IRD in sediments from 50 ka to the deglaciation, we assume that only very few sediment-laden icebergs drifted far off the continental margin north of Svalbard.

Calculated planktic carbonate fluxes were extremely low during much of the Late Weichselian in the central Arctic Ocean (Nørgaard-Pedersen et al., 1998) and the apparent high planktic foraminifer contents in some cores after 50 ka are caused by low sedimentation rates and low terrigenic dilution. Only in the middle of OIS 3 (around 40 ka), a peak in the flux rates indicates seasonally open water conditions in the central Arctic Ocean. The associated inflow of Atlantic Water probably correlates to the "high productivity" event HP3 in the Fram Strait and Nordic Seas at ca 38–41 ka (35–37.5 ^{14}C-ka), as determined by Dokken and Hald (1996) and Hald et al. (2001). Since this time interval was characterized in Scandinavia and Svalbard by a retreat of continental ice (Baumann et al., 1995; Mangerud et al., 1998), the Atlantic Water inflow around 40 ka was probably quite strong and caused a negative glacier ice mass balance and no ice sheet growth, as indicated also by low or decreasing IRD contents in sediments from 40 to 35 ka.

The younger Atlantic Water inflow and high productivity events HP2 (32–26 ka) and HP1 (around 20 ka) were associated with a two-step build-up of the Svalbard–Barents Sea ice sheet (Hebbeln et al., 1994; Elverhøi et al., 1995; Hald et al., 2001). Both events are reflected in PS1535 and PS1533-3 by high amounts of planktic foraminifers. In a recent study, the distribution of Atlantic Water in the Fram Strait and adjacent Arctic Ocean up to 84°N during HP1 (equivalent to the LGM) could be mapped (Nørgaard-Pedersen et al., 2003). It was concluded that the central Arctic Ocean remained largely unaffected by the surface-near Atlantic Water inflow and comprised a thick sea ice massif with very little open water areas. None of the cores from this study contained evidence for a significant glaciation in northernmost Siberia during the LGM (Nørgaard-Pedersen et al., 2003).

The deglaciation history of the Barents Sea and NW Kara Sea is reflected in sediment cores from the continental margins, the eastern Nordic Seas, the Fram Strait, and the adjacent Arctic Ocean by strong IRD input and evidence for meltwater events (Jones and Keigwin, 1988; Sarnthein et al., 1992, 1995; Bischof, 1994; Hebbeln et al., 1994; Stein et al., 1994a; Elverhøi et al., 1995; Andersen et al., 1996; Dokken and Hald, 1996; Lubinski et al., 1996, 2001; Hebbeln and Wefer, 1997; Polyak et al., 1997; Nørgaard-Pedersen et al., 1998, 2003; Knies et al., 1999, 2000, 2001; Kleiber et al., 2000; Vogt et al., 2001). It started at ca 15 ka (Landvik et al., 1998; Nørgaard-Pedersen et al., 1998), by 13 ka ice had disappeared except on the islands, and by 10 ka also the fjords were ice-free (Landvik et al., 1998). Of all sediment cores in the present study, only the cores from the Arctic Gateway contain an IRD-rich layer deposited during the deglaciation. Both PS1533-3 and PS1535 have a carbonate-free section (5 and 10 cm thick) from this time interval. These sediments are barren of planktic foraminifers and cause a gap in the isotope records which covers the change from high, glacial to low, interglacial δ^{18}O values. Other Fram Strait cores show a similar feature (cf. Hebbeln and Wefer, 1997). Since several records from the Fram Strait show a strong meltwater spike from ca 14.5 ^{14}C-ka (Jones and Keigwin, 1988; Sarnthein et al., 1992; Dokken and Hald, 1996), we speculate that in some regions of the Arctic Gateway the strong meltwater inflow caused a salinity decrease below the tolerance limit of the foraminifers ($S \approx 28$). Deglacial meltwater events are also recorded in the Arctic Ocean, but amplitudes of freshwater spikes in the isotope records decrease toward the pole (Stein et al., 1994a; Nørgaard-Pedersen et al., 1998, 2003). Although low sedimentation rates at the deep-sea sites limit the resolution of the sediment cores, there is evidence from isotope records and analysis of IRD for two separate events in the interior Arctic, which reflect the diachronous deglaciation of the northern Barents Sea and northernmost North America at 15–13.5 ^{14}C-ka (18–16 cal-ka) and 14–12 ^{14}C-ka (16.5–14 cal-ka), respectively (Nørgaard-Pedersen, 1996; Nørgaard-Pedersen et al., 1998, 2003).

In the Holocene, deep-sea sedimentation in the central Arctic Ocean and the Arctic Gateway was dominated by input of fine-grained terrigenous material and biogenic carbonate, which is typical also for the modern situation (Hebbeln and Wefer, 1991; Stein et al., 1994b, c; Nørgaard-Pedersen et al., 1998; Hebbeln, 2000). The deposits contain abundant planktic foraminifers and little IRD. Modern sources for icebergs around the Arctic Ocean are Greenland, Svalbard, Franz-Josef-Land, Ellesmere Island, and Severnaya Zemlya, but only from the latter two sources icebergs reach the central and eastern Arctic Ocean and supply some IRD to the deep-sea areas. Available records show no systematic changes of IRD contents within the Holocene (Nørgaard-Pedersen et al., 1998, 2003). Today, sea ice is supposed to be the main transport agent for terrigenic particles in hemipelagic Arctic sediments (Nürnberg et al., 1994; Eicken et al., 2000). Planktic foraminifer abundances in central Arctic Ocean sediment cores reach a level of high values shortly after Termination I (Nørgaard-Pedersen et al., 1998, 2003) at the time when post-glacial sea-level rise led to flooding of the wide Siberian shelves (cf. Bauch et al., 2001). These data indicate that the establishment of full Atlantic Water inflow to the Arctic Ocean and interglacial conditions with frequent open water areas (leads) in the sea ice cover occurred at ca 8–9 ka.

8. Arctic Ocean paleoenvironments and northern Eurasia ice sheets during the past 200 ka—a complex relationship

When comparing reconstructions of the northern Eurasian ice sheet history based on terrestrial field work with the interpretations of our sediment records of the last 200 ka from the Arctic Ocean, a number of regularities are revealed. The Arctic Ocean foraminifer abundance record provides evidence for the inflow of Atlantic Water to the Arctic basin before each of the large-scale glaciations in northern Eurasia, i.e. before OIS 6 (ca 190–130 ka), in OIS 5b (ca 90–80 ka), and in late OIS 4/early OIS 3 (ca 65–50 ka), which are reflected by strong IRD input to the deep-sea (Fig. 10). At times, when considerable open water areas (leads) were present in the Arctic Ocean, the adjacent areas of the Fram Strait and the Nordic Seas must have had even warmer, partly ice-free surface waters, which could provide moisture from evaporation for an eastward atmospheric transport and the build-up of ice sheets on northern Eurasia from enhanced precipitation. It is often difficult to determine the initial age of a glaciation phase from terrestrial evidence, because usually the oldest glacial features were later eroded by the growing ice sheet (cf. Svendsen et al., 2004). Also, a long-lasting strong input of IRD to the ocean will occur only when the ice sheet source has reached already a considerable size, and the IRD-rich layers probably reflect times of (almost) full ice sheet development and the glacial terminations. Thus, we hypothesize that the apparent pause between Atlantic Water inflow (indicated by high foraminifer abundances) and the presence of large ice sheets (indicated by strong IRD input) as shown in Fig. 10 did not exist, because in most cases the phase of ice sheet build-up in areas far off the continental margin is not recorded in deep-sea sediments by IRD input. The excellent correlation, however, between the ice sheet

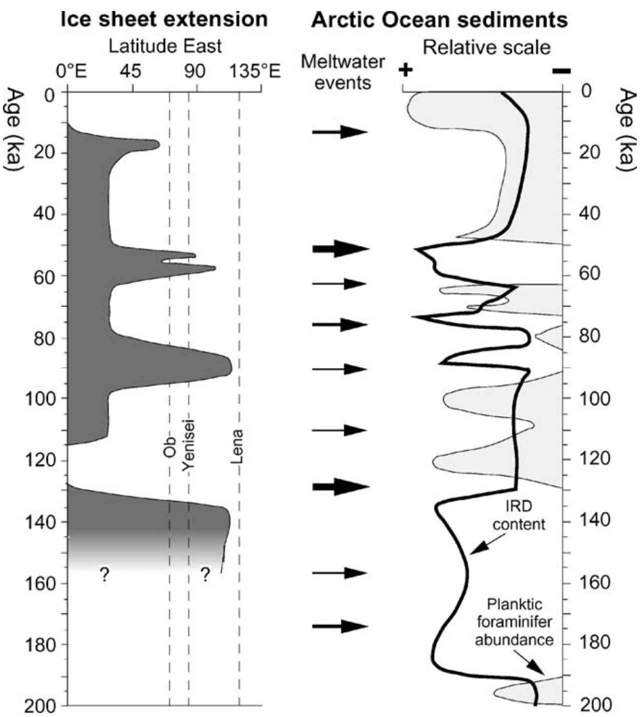

Fig. 10. Comparison of reconstructions of the eastward extension of ice sheets during the past 150 ka (from Svendsen et al., 2004) with results from central Arctic deep-sea sediment cores presented in this study. The geographical longitudes of present river mouths of Ob, Yenisei, and Lena are indicated by vertical lines. A residual ice sheet over Scandinavia (0–30° E) is assumed for most of the last 200 ka.

history as reconstructed from terrestrial field work, and the history of IRD input is striking and suggests that the applied stratigraphic models are largely correct. The only exception is the period of strong IRD input around 75 ka, for which no counterpart in terms of a large-scale ice sheet on northern Eurasia could be revealed yet.

During glacials, and especially during the glacial terminations, the ice sheet developments on the northern Eurasian continent and the Barents and Kara seas shelves significantly influenced the freshwater budget of the Arctic Ocean. Besides the release of icebergs, the damming and diversion of northward flowing rivers (cf. Mangerud et al., 2001, 2004) and the subsequent discharge of freshwater from ice-dammed lakes during ice sheet decay must have changed the surface water environments drastically. Using the terrestrial ice sheet reconstructions of Svendsen et al. (2004) as boundary conditions, the major Siberian rivers of Ob and Yenisei were dammed at least three times during the past 200 ka (190–130, 90–80, 65–50 ka; Fig. 10). These periods of reduced river influx to the Arctic Ocean were terminated in the deglaciations by the probably catastrophic discharge of huge amounts of freshwater from the ice-dammed lakes. Accordingly, the strongest freshwater events in the Arctic Ocean were recorded at ca 130 and 52 ka. The freshwater spikes in records from the early Weichselian glaciation seem less focussed on a certain short interval and it is possible that the deglaciation occurred less rapidly than the others. An important support for the terrestrial reconstruction of a limited eastward extension of the ice sheet during the LGM is the observation of little IRD in central Arctic Ocean sediments and only small amplitudes of freshwater spikes in the planktic isotope records. The large amounts of freshwater from the largest events at 130 and 52 ka may even have reached the Nordic Seas and thus contributed to the contemporaneous low-salinity events in the northernmost North Atlantic.

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