History of ice-rafting and water mass evolution at the northern Siberian continental margin (Laptev Sea) during Late Glacial and Holocene times

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1. Introduction

The Arctic is a key region in the global climate system through its influence on ocean–atmosphere heat balance and circulation, making it particularly vulnerable to climate warming since the amplitude of temperature rise is expected to be higher here than elsewhere (ACIA, 2005; IPCC, 2007; Bauch and Kassens, 2005; Polyak et al., 2010). The ice-albedo effect and changes in the rate of inflow and temperature of the warm Atlantic-derived water combined with the outflow of Polar water are especially important for the Arctic climate system. During the recent decade a considerable reduction in the sea-ice cover extent and thickness was recorded with an extreme shrinkage occurring in 2007 (Serreze et al., 2007; Comiso et al., 2008). Also, the volume and temperature of the inflowing Atlantic-derived water have increased (Schauer et al., 2004; Polyakov et al., 2005). Besides the overall

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climatic importance of the air temperature rise due to increasing concentration of atmospheric greenhouse gases, the variability of sea-ice concentration and drift pattern, as well as the intensity and temperature of the inflowing Atlantic waters are shown to be associated with the short- and long-term changes in the phase of the atmospheric North Atlantic and Arctic Oscillations (NAO/AO) (Wadhams, 2000; Dickson et al., 2000; Mysak, 2001).

Understanding the ongoing changes and assessing the possible future variability demands improved reconstructions of sea ice and ice sheet history as well as the natural variability of water masses and circulation patterns, which, in our case, is primarily the Late Glacial to Holocene history of Atlantic-derived water inflow into the Arctic Ocean along the northern Eurasian margin. Important proxies to reconstruct these parameters are concentrations of coarse terrigenous ice- and iceberg-rafted material (IRD) and warm-water microfossils in marine sediments, especially from continental margins that mostly comprise sediment records with increased temporal resolution (Darby et al., 2006; Stein, 2008; Polyak et al., 2010). During terminal glacial times, icebergs were delivered to the eastern Arctic from the shelf-based Barents–Kara ice sheet (Fig. 1a; Svendsen et al., 2004). Disintegration of this ice sheet was accompanied by discharge of meltwater, especially in the case of ice-dammed lakes on the shelf (Spiehlagen et al., 2004; Mangerud et al., 2004). Although ice sheet boundaries during the Last Glacial Maximum (LGM) have been fairly well constrained for this region, there are still debates about its eastward extent and the possibility of a short-time blockage of river runoff on the NE Kara shelf (Alexanderson et al., 2001; Knies et al., 2001; Svendsen et al., 2004; Polyak et al., 2008). Similarly, although there is evidence for the absence of a contiguous Kara Sea ice sheet on the eastern Eurasian archipelago, i.e., Severnaya Zemlya, during the LGM (Raab et al., 2003; Hubberten et al., 2004; Möller et al., 2007), the climatically controlled size of the local ice caps and iceberg production during the LGM and postglacial times are not well constrained by data due to the scarcity of marine cores with sufficient age control (Knies et al., 2001).

It is commonly believed that enhanced inflow of Atlantic-derived subsurface waters occurs into the Arctic Ocean during interglacial periods resulting in generally milder climate conditions and reduced sea-ice cover with more open leads along the northern Eurasian periphery (Knies et al., 2000; Spiehlagen et al., 2004). There is evidence that the subsurface inflows of these waters could be quite intense also during deglacial and even glacial epochs (Knies et al., 1999; Nørgaard-Pedersen et al., 2003; Kristoffersen et al., 2004). In the northern Nordic seas Atlantic water was continuously present since the LGM, and neither extensive sea ice cover nor large inputs of meltwater stopped this inflow (Bauch et al., 2001a; Rasmussen et al., 2007). The history of the inflows of Atlantic-derived waters to the Arctic Eurasian margin during the deglacial and Holocene times has been recently reconstructed for the Svalbard region as well as the northern Barents and Kara seas margins (Lubinski et al., 2001; Duplessy et al., 2001, 2005; Köc et al., 2002; Słobowska et al., 2005; Słobowska-Woldenberg et al., 2007, 2008; Ivanova, 2006). In contrast, the eastward propagation (i.e., east of the Kara Sea) of the Atlantic water is so far poorly known due to the lack of sediment cores with high time resolution, sufficient fossil contents, and thus good dating possibilities.

In this study we present continuous records of IRD and Atlantic water indicative microfossils from the western Laptev Sea continental margin (Fig. 1a). The two cores studied are located directly in a key position to investigate the temporal variability in iceberg production and meltwater discharge (Fig. 1a). As previously shown both cores have a sound age frame based on numerous AMS$^{14}$C datings and, thus, allow for a high-resolution paleoenvironmental reconstruction since Late Glacial times (Bauch et al., 2001b; Taldenkova et al., 2008). The further intention of this study is to uncover the complex interaction between climate changes, regional ice sheets/sea-ice cover and Atlantic-derived waters during the transition from the LGM to the present for a region which is located farthest from Atlantic and Pacific water mass influences. Aside from using continuous, high-resolution records of IRD and Atlantic water indicative microfossils (planktic and benthic foraminifers) to obtain a larger-scale regional picture, we will then also discuss the occurrence of authigenic concretions in relation to local changes in bottom water conditions and surface ocean stratification.

2. Regional setting of the Laptev Sea

2.1. Hydrography

The inflow of the warm and saline Atlantic-derived water masses to the Arctic Ocean and its marginal seas is driven by thermohaline and estuarine forcing as well as atmospheric pressure (ACIA, 2005). Atlantic waters enter the Arctic Ocean via two main branches (Rudels et al., 2004; Saloranta and Haugan, 2004; Carmack et al., 2006) (Fig. 1a). The Fram Strait Branch Water (FSBW) flows north of Svalbard with the West Spitsbergen current. The Barents Sea Branch Water (BSBW) is formed from interactions between the Atlantic Water flowing into the southwestern Barents Sea and local fresher and colder Barents and Kara shelf waters. Besides the difference in temperature and salinity characteristics, these Atlantic-derived water masses have specific phytoplankton deposition and nutrient composition that are essential for benthic species occurring within or underneath these water masses (Lubinski et al., 2001; Wollenburg et al., 2004). At the Laptev Sea continental slope, relatively warm (~1 °C) and salty (34.7–34.9) Atlantic-derived waters occur within the depth range of 200–600 m (Timokhov, 1994), i.e., about 100 m below the shelf break.

The Laptev Sea is a shallow shelf sea with the shelf break following the 80–100 m contour lines. The shelf is cut by several palaeoriver valleys formed in the Pleistocene during times of sea-level lowstands (Bauch et al., 1999, 2001b; Bauch and Kassens, 2005; Darby et al., 2006). The hydrography of the Laptev Sea is characterized by the interaction of freshwater input from large Siberian rivers, primarily the Lena River, cold brines produced from sea ice formation, and meltwater input during summer sea-ice melting. Furthermore, Atlantic-derived waters which flow along the continental slope may penetrate the outer shelf, especially along the paleovalleys, due to wind-induced reversed currents (Dmitrenko et al., 2001). Because considerably less riverine waters drain into the western than into the eastern Laptev Sea benthic and planktic communities in the western Laptev Sea are dominated by marine species compared to more euryhaline and brackishwater species occurring in the east (Petryashov et al., 1999; Stepanova et al., 2003, 2007).

2.2. Sea ice: distribution and sediment entrainment

Both, its position in the high Arctic and decreased surface water salinity caused by high river runoff support enhanced sea ice production in the Laptev Sea. Sea ice is supplied to the Transpolar Drift, which is the main ice drift system exporting ice from the Arctic through Fram Strait and into the Nordic seas (Rigor and Colony, 1997) (Fig. 1b). Although the Laptev Sea becomes largely ice-free during the short summer season changeable wind patterns can make a major difference on an interannual basis. For instance, in the western Laptev Sea a sea ice massif may exist perennially, often blocking Vilkitskii Strait thereby occupying between 25
and 75% of the western Laptev Sea in September (Karklin and Karelin, 2009).

Due to suspension freezing processes the shallow Laptev Sea shelf also produces sea ice which is enriched in sediment inclusions of different grain size (Reimnitz et al., 1992, 1994; Dethleff et al., 1993; Nürnberg et al., 1994; Eicken et al., 1997). Although silt and clay particles predominate, coarser material may be entrained due to anchor ice formation or direct adfreezing to the bottom of landfast ice in the nearshore zone (Reimnitz et al., 1987, 1994; Eicken et al., 1997; Reimnitz et al., 1998). The ice freeze-up in the Laptev Sea starts in September, when the drift ice margin is located in the northern part of the sea, whereas its shallow southern region freshened by summer runoff is ice-free and open to storms. Autumn storms cool down the water column by turbid mixing and may re-suspend bottom sediment particles which then become entrained into the newly formed ice (Reimnitz et al., 1987, 1994;
This sediment-laden ice is brought offshore and added to the drift ice cover. Further cooling results in formation of landfast ice in areas with water depths of 20–25 m, where it becomes separated from the drift ice by a polynya formed in winter due to offshore winds. In summer, fast ice mainly melts in place, being destroyed by warmer riverine floodwater in the south and enhanced melting of its northern edge due to heat accumulation by polynya.

The release of coarse-grained terrestrial debris from sea ice to bottom sediments largely occurs in the two regions. Firstly, in the inner shelf zone where fast ice and also river ice melts (Dethleff et al., 1993; Eicken et al., 1997). Secondly, beyond the average multiannual southern drift ice limit where coarse debris from the sediment-laden ice formed in autumn is released along the driftway to the Fram Strait. This implies that the colder the summer is, the farther southward is the seasonal drift ice limit, and more sediment is released next summer within the outer Laptev Sea. However, not only surface water temperature, but also the wind and current regimes determine the configuration of the summer drift ice edge. In the western Laptev Sea due to higher salinity and steeper shelf along the Taimyr and Severnaya Zemlya coasts sea ice production and coarse sediment entrainment are less active than in the southeastern regions. But here also sediment material from the pack ice of Taimyr massif and ice imported from the Kara Sea should be released in summer (Pavlov and Pfirman, 1995; Karklin and Karelin, 2009).
2.3. Severnaya Zemlya: geology, ice caps and icebergs

At present the high Arctic Severnaya Zemlya Archipelago is the closest iceberg production site to our study area (Fig. 1a). It consists of numerous islands with the 3 biggest islands being Bol’shevik, October Revolution and Komsomolots located to the north from Vilkitskii Strait (Figs. 1 and 2). Today ice caps cover major part of the islands and reach a thickness of 1000 m (Govorukha, 1988).

Several outlet glaciers from Akademii Nauk, Rusanov, Karpinskii, Universitet ice caps terminate at sea-level and produce icebergs (Fig. 2). The average size of icebergs released by Severnaya Zemlya glaciers is 35–45 m with the length of 50–420 m (Govorukha, 1988). However, big icebergs (ice islands) occur at the flat eastern margin of Akademii Nauk ice cap (Govorukha, 1988). An ice shelf in Matuschewich fjord fed by Rusanov and Karpinski ice caps produces tabular icebergs up to several kilometers in length (Williams and Dowdeswell, 2001). At present these big icebergs are often grounded in the fjord or trapped by fast ice cover, and rather smaller icebergs deliver sediment material to the western Laptev Sea (Williams and Dowdeswell, 2001).

Sediment material delivered by icebergs from Severnaya Zemlya should be rather diverse as rocks of different composition and age crop out on the islands (Fig. 2; Geologiya SSSR, 1970; Lorenz et al., 2008). Neoproterozoic formations of significant thickness dominated by turbidites with greenish-gray to almost black metamorphic schists with high mica content (phylites) occur along northern Taimyr, Bol’shevik and eastern October Revolution islands, – the areas facing the western Laptev Sea (Fig. 2). Fragments of such rock types are common in our studied cores (Fig. 2) and we regard them as a possible indication of material transported with icebergs from Severnaya Zemlya (see Sections 4.1 and 4.2).

3. Material and methods

The cores studied were obtained in 1998 during TRANSDrift V expedition of RV Polarstern (Fig. 1a). Kastencore PS51/154-11 originates from the upper continental slope. Kastencore PS51/159-10 was recovered from the Khatanga paleoriver valley on the outer Laptev Sea shelf. Hereafter, for simplicity, they will be named cores PS51/154 and PS51/159. Characteristics of the studied cores are summarized in Table 1.

The chronology of the studied cores is based on AMS14C dates of marine biogenic carbonate of bivalves and microfossils obtained from samples of core PS51/154 lithic grains was determined in the samples containing more than 300 grains per sample (Gottschalk, 2008).

Both cores together with their core catchers were sampled continuously in 2-cm thick slices. Samples were freeze-dried and subsequently washed over 63 μm mesh size sieve. For each sample the weight percentage of the sand fraction (>63 μm) was determined. Dry samples were sieved over 500 μm mesh size and all lithic grains were picked and counted under a binocular microscope. From samples of core PS51/154 lithic grains were additionally picked and counted from >1000 μm fraction. Although sometimes

<table>
<thead>
<tr>
<th>#</th>
<th>Latitude °N</th>
<th>Longitude °E</th>
<th>Water depth, m</th>
<th>Core sediment length, cm</th>
<th>Core catcher sediment length, cm</th>
<th>Total sediment length, cm</th>
</tr>
</thead>
<tbody>
<tr>
<td>PS51/154-11</td>
<td>77.276</td>
<td>120.610</td>
<td>270</td>
<td>677</td>
<td>23</td>
<td>700</td>
</tr>
<tr>
<td>PS51/159-10</td>
<td>76.768</td>
<td>116.032</td>
<td>60</td>
<td>419</td>
<td>28</td>
<td>447</td>
</tr>
</tbody>
</table>

Table 2

<table>
<thead>
<tr>
<th>Lab #</th>
<th>Depth, cm</th>
<th>Material</th>
<th>Radiocarbon (AMS) dates from the studied cores and the calibrated calendar years.</th>
</tr>
</thead>
<tbody>
<tr>
<td>PS51/154-11</td>
<td>35.2</td>
<td>3279</td>
<td>342.5 ± 30</td>
</tr>
<tr>
<td>KIA-6919</td>
<td>31 Yoldiella intermedia</td>
<td>1540 ± 45</td>
<td>1079</td>
</tr>
<tr>
<td>KIA-32810</td>
<td>39 foraminifers</td>
<td>5040 ± 50</td>
<td>5398</td>
</tr>
<tr>
<td>KIA-32811</td>
<td>39 bivalves/gastropod</td>
<td>1800 ± 35</td>
<td>1322</td>
</tr>
<tr>
<td>KIA-27683</td>
<td>51 Yoldiella sp.</td>
<td>9570 ± 60</td>
<td>10347</td>
</tr>
<tr>
<td>KIA-32812</td>
<td>73 foraminifers</td>
<td>9410 ± 70</td>
<td>10286</td>
</tr>
<tr>
<td>KIA-32813</td>
<td>73 Yoldiella lenticulata sp.</td>
<td>9665 ± 45</td>
<td>10401</td>
</tr>
<tr>
<td>KIA-27684</td>
<td>85 foraminifers/Portlandia arctica</td>
<td>9505 ± 50</td>
<td>10265</td>
</tr>
<tr>
<td>KIA-32814</td>
<td>115 Yoldiella lenticulata</td>
<td>9630 ± 50</td>
<td>10442</td>
</tr>
</tbody>
</table>
Here we present only the data on the abundance of Atlantic water indicative microfossils, that is benthic foraminiferal species *Cassidulina neoteretis* and diverse subpolar planktic foraminifers (Figs. 6 and 7). Relative percentage of *C. neoteretis* (Fig. 7) was calculated only for samples containing more than 100 tests. The data on the distribution of these microfossils in core PS51/154 were partly presented in Taldenkova et al. (2008). Since that time we obtained new radiocarbon dates, which slightly improved the age model of the upper core section, and also examined the lower 1 m of the core in more detail. These lowermost sediments, which include the core catcher, were not studied before, and extended the time frame of our record.

4. Results

4.1. Core PS51/154: upper continental slope

4.1.1. Chronology and sedimentation rates

The core chronology is constrained by eighteen AMS$^{14}$C dates (Table 2, Fig. 3). Nine of them were published previously (Bauch et al., 2001b). The recently obtained nine datings improved the previous age model, especially the age estimations of the uppermost core section with rather low sedimentation rates. For age model calculations we excluded the two reversal dates from 31 to 39 cm depth obtained on bivalves (Table 2). This seems reasonable, as these reversals are most likely the result of burrowing of the bivalves into sediments accumulated under low sedimentation rates (Fig. 3). The sediment section between 51 and 115 cm shows enhanced bioturbation and the radiocarbon dates yielded practically similar ages (Table 2). Therefore, only the youngest and the oldest dates were considered for age model calculations, and the ages obtained on foraminifers from 73 cm depth was assigned to 51 cm core depth. Samples ages between the 13 used date fixpoints were estimated by linear interpolation and assuming a modern age for the sediment surface. The extrapolated ages of the core base and the bottom of the core catcher are 17.2 and 17.6 ka, respectively (Fig. 3).

Linear sedimentation rates calculated between neighbouring calendar age points reach highest values between 13.5 and 14, and 10 and 12 ka (Fig. 3). In general, the calculated average sedimentation rates were high (89 cm/kyr) during accumulation of the major part of the section, but sharply decreased to about 5 cm/kyr...
at approximately 10 ka, i.e. after the onset of shelf flooding (Bauch et al., 2001b).

4.1.2. Lithology

The 700 cm thick sediment sequence of the core and core catcher is represented by greyish clayey silt and silty clay with sandy lenses in the lowest 2 m of the section and with signs of bioturbation in the rest. Lithologically, it might be roughly subdivided into three sediment units: lower (570–700 cm, extrapolated age 15.4–17.6 ka), middle (50–570 cm, 10.2–15.4 ka), and upper (6–50 cm, 0.6–10.2 ka) (Fig. 3).

The lower sediment unit is impoverished in fossils, but enriched in mica flakes, authigenic concretions, and IRD. Weight percentage of the fraction >63 μm fluctuates between 3 and 10%. The middle sediment unit is highly fossiliferous, contains abundant plant debris, and is almost devoid of IRD. Weight percentage of the fraction >63 μm is highly variable. At the base of this unit, a spike up to 70 wt% corresponds to a sand slide layer. It is represented by uniform sand size particles and contains different nearshore microfossils including brackishwater and freshwater ostracods (Taldenkova et al., 2008). Upwards, the content of >63 μm fraction gradually decreases to 2–3 wt%. The fossiliferous upper sediment unit accumulated under very low sedimentation rates and is distinguished by high average percentage of the fraction >63 μm (20–25 wt%) and high IRD concentrations.

4.1.3. Authigenic concretions

An important feature of the lower sediment unit is the presence of vivianite (Fe₃(PO₄)×8H₂O) and concretions of rhodochrosite (MnCO₃) (Figs. 3 and 4). The number of vivianite grains is quite variable (Fig. 3), so are the grain sizes. Some are less than 1 mm (Fig. 4a), others comprise big agglomerates exceeding 1 cm (Fig. 4b). Rhodochrosite mainly occurs as tubes (Fig. 4c), which are either empty or infilled with clusters of nodules. Some concretions are ramiform (Fig. 4d). The tubes are usually up to 1–2 cm long with a diameter of 1–2 mm. Prevalent tube-like form of rhodochrosite concretions most likely indicates the presence of gas seepage channels. Also, numerous fragments of the walls of rhodochrosite tubes are present in the sediments. Since it is difficult to count these fragmented concretions, only the depth intervals of rhodochrosite occurrence are shown in Fig. 3. Concretions of both types show peak concentrations between 656 and 677 cm (c. 16.8–17.2 ka) and a second small spike in the layer 584–616 cm (c. 15.6–16.2 ka).

4.1.4. Ice-rafted debris (IRD)

There are two IRD-rich intervals in core PS51/154 – the lower (c. 16.2–17.6 ka) and upper (7–0.6 ka) sediment units (Figs. 3, 6 and 7). Within these intervals, IRD occurs in several peaks labeled 1 (oldest) to 7 in Figs. 3 and 6. Both periods of IRD accumulation have abrupt, sharp upper or lower boundaries – Peak 4 at c. 16.2 ka and Peak 5 at 7 ka. The highest IRD concentration of 10 grains/1 g is recorded by Peak 3 (c. 16.7 ka), but bigger fragments (>1000 μm) are more abundant in Peak 6 (3–4 ka) within the upper sediment unit.

The composition of lithic grains varies within different IRD peaks, although quartz fragments, both rounded and angular, are dominant throughout the whole sediment sequence. Rock fragments are most abundant in Peak 1 (c. 17.6 ka) and Peaks 6 (3–4 ka) and 5 (4.5–7 ka) of the upper sediment unit. The share of phyllites from Severnaya Zemlya is highest at the very base of the sequence.
during Peak 1 (up to 20% of the total grains) and remains high during Peak 2 (c. 17.2 ka) and early Peak 3 (c. 16.9 ka), but then drastically diminishes to almost zero at the end of Peak 3 (c. 16.6 ka) and during Peak 4 (c. 16.2 ka). In the upper sediment layer the share of phyllites is high throughout.

In the lower sediment unit, IRD peaks intercalate with intervals enriched in concretions. Obviously, the latter postdate the periods of IRD input. Peak 1 with the highest concentration of rocks and phyllites and no concretions was quickly followed by a rather similar Peak 2, after which the main concretion-formation event occurred c. 16.8–17.2 ka. The next Peak 3 was different from Peaks 1 and 2 as the composition of IRD changed to the predominance of quartz grains and diverse rocks other than phyllites. After Peak 3, there is an interval in the core (624–640 cm, c. 16.3–16.6 ka) with sediments almost devoid of both, the IRD and concretions. Here, the washed residues from the depth of 626–630 cm are represented by rusty-colored blocks of "lithified" clay and silt. These might be fragments of a thin crust formed by diagenetic alteration of sediments similar to oxidized layers found in some deglacial sediments with slow sedimentation (Polyak et al., 1997). This interval is followed by IRD Peak 4 and the subsequent small spike of vivianite.

In summary, the two IRD-rich sediment intervals of the core correspond to essentially different paleoenvironments. The upper IRD-rich interval (c. 16.2–17.6 ka) is somewhat heterogeneous. The composition and abundance of grains in the two lower peaks of this interval (1 and 2) are rather similar to the upper IRD-rich interval and imply the discharge of iceberg-rafted material from the local ice caps on Severnaya Zemlya. This is especially true for Peak 1 which is devoid of authigenic concretions. The two younger IRD peaks (3 and 4) which occurred after the main concretion-formation event contain less phyllites from Severnaya Zemlya, but more quartz and rocks from other sources. This probably gives evidence for the iceberg-rafted material from a disintegrating Barents–Kara ice sheet.

### 4.1.5. Foraminifers

Although the lower IRD-rich sediment interval of the core is almost barren of benthic fossils there are two samples from the depth of 672–677 cm (age 17.2 ka) containing rare tests of benthic foraminifers including *C. neoteretis*. *C. neoteretis* is among the first species to appear in the fossil-rich part of the record at 15.4 ka and is continuously present until 12 ka (Figs. 6 and 7), thus evidencing the constant and strong inflow of chilled subsurface Atlantic-derived waters to the site (cf. Lubinski et al., 2001). It is absent during the Early Holocene, but then re-appears between 5 and 2 ka.

Subpolar planktic foraminifers also indicative of subsurface Atlantic-derived water inflow to the Laptev Sea (Volkmann, 2000) first appear at the very base of the record (age 17.6 ka) and are represented by several tests of *Globigerinita uvula* (Fig. 6). A rather high abundance of subpolar planktic foraminifers (>3000 tests/100 g sediment) is recorded in the same samples which contain benthic foraminifers as well (age 17.2 ka). Most of these specimens have small-sized tests from the fraction 63–125 μm and are largely represented by *Neogloboquadrina pachyderma* dex., but also do include *Globigerina bulloides*, *Turborotalita quinqueloba*, *Globigerinita glutinata*, *G. uvula*, *Globorotalia scitula*, *G. inflata* and *Orbulina*...
In these two samples subpolar planktic foraminifers dominate over the polar species *N. pachyderma* sin., which is otherwise the dominant species throughout the core (Taldenkova et al., 2008). Further up the section until 15.4 ka subpolar planktic foraminifers occur periodically, in low numbers. Usually these are less than 5 tests per sample. Since the tests of planktic foraminifers are well preserved, this confirms that the simultaneous absence of benthic microfossils is not the result of carbonate dissolution. It can be concluded that both groups of microfossils, which indicate a presence of Atlantic-derived waters, are most abundant in the IRD-poor sediments dating back to 12 e 15.4 ka, i.e. the time after the peak Late Glacial IRD input and prior to the onset of extensive shelf flooding (Bauch et al., 2001b). However, they do also occur in the IRD-rich sediments. In the lower IRD-rich interval, the single but very conspicuous event containing highest abundance of small-sized subpolar planktic foraminifers and the presence of rare benthic forms including *C. neoteretis*, altogether indicate a strong, but short-lived intrusion of Atlantic-derived waters. This intrusion was time-coeval with IRD Peak 2 and occurred prior to the major concretion-formation event. In the upper IRD-rich layer the microfossils correlate with the strongest IRD Peaks 5 and 6 dating back to c. 4.5–7 and 3–4 ka, respectively (Figs. 6 and 7).

### 4.2. Core PS51/159: outer shelf

#### 4.2.1. Chronology and sedimentation rates

The core chronology is constrained by eight AMS$^{14}$C dates (Table 2, Fig. 5). The lowermost date was obtained on a sample from the core catcher, and the interpolated age of the core’s base yielded 12.2 ka. Sedimentation rates were highest at about 10.2–10.5 ka (Fig. 5). The average sedimentation rates were 114 cm/kyr during the early stage of site flooding, but sharply decreased down to 14 cm/kyr at approximately 9.6 ka (Fig. 5) due to the southward retreat of the main depocenters as the shelf flooding progressed (Bauch et al., 2001b).

#### 4.2.2. Lithology

The 447 cm-thick sediment sequence of the core and core catcher is represented by dark greyish silty clay, bioturbated, with abundant plant debris below 200 cm core depth. Core PS51/159 is located in a former river valley filled with marine sediments after flooding (Bauch et al., 2001b). The sand size fraction is rather uniform and does not exceed 8.5 wt%. This value is considerably less than found in core PS51/154 from the continental slope (see Section 4.1, Figs. 3 and 5). The highest variability of the sand fraction is observed between 160 and 200 cm core depth (c. 7–10 ka) and corresponds to the time interval when, according to the composition of benthic assemblages, the site turned from an inner shelf to mid-shelf depositional setting (Taldenkova et al., 2009). Accordingly, the site was located close to the fast ice edge and winter polynya at this time, a situation which might have induced the changeable hydrodynamic regime reflected by rapid jumps in the sand size fraction.

#### 4.2.3. Authigenic concretions

No authigenic vivianite and concretions of rhodochrosite occur in the sediments of core PS51/159, but numerous tabular crystals and aggregates of gypsum are present in the samples from the core.
catcher (419–447 cm). The crystals are perfectly preserved thus excluding the possibility of a detrital origin (St. John and Cowen, 2000). The core catcher was sampled after it had been kept in a cool storage for 8 years. In such cases, formation of authigenic gypsum may result from dehydration of the unconsolidated sediment (Alvi and Winterhalter, 2001).

4.2.4. Ice-rafted debris (IRD)

The presence of IRD in the core is largely restricted to the upper 100 cm, i.e. to the last 7 kyrs (Figs. 5–7; Peaks b–f). Only the oldest Peak a occurs in the lower part of the sequence which corresponds to 11.6–11.8 ka. At that time water depths at the site were less than 20 m, and inner shelf benthic assemblages inhabited the region (Bauch et al., 2001b; Taldenkova et al., 2009). Therefore, this peak may not be an indication of iceberg-rafting, but rather a result of bottom erosion which is corroborated by a simultaneous increase of the sand fraction content (Fig. 5).

The upper IRD peaks (b–f) are in time similar to the peaks found in the Mid–Late Holocene sequence of core PS51/154. However, the total abundance of grains here is considerably less (Figs. 3 and 5). Compared to the lower Peak a, Peaks b–f include more large angular fragments and phyllites. Given the modern-like water depths at the site in the Mid–Late Holocene, this may give evidence for a discharge of iceberg-rafted material from Severnaya Zemlya.

4.2.5. Foraminifers

C. neoteretis is absent in the sediments of core PS51/159, because in its modern distribution it is restricted to deeper environments on the continental slope within chilled, subsurface Atlantic-derived waters (Lubinski et al., 2001). However, rare but taxonomically diverse subpolar planktic foraminifers, mainly small-sized tests occur in the core (Fig. 6). These not only comprise those species recorded in core PS51/154 (see Section Fig. 7. Overregional comparison of the western Laptev Sea and North Atlantic-western Arctic records. Shading corresponds to the western Laptev Sea record: IRD peaks (a) in the upper half and peaks in Cassidulina neoteretis percentage (d) in the lower half. Plot (b) shows the IRD record of core 23258-2 (1768 m water depth) from the western Barents Sea continental slope (Sarnthein et al., 2003; http://doi.pangaea.de/10.1594/PANGAEA.144682). Dotted line and the left-hand scale correspond to enlarged record of the last 9 kyrs. Plot (c) shows the NGRIP oxygen isotope record (http://www.iceandclimate.nbi.ku.dk/data). Plot (e) represents the record of C. neoteretis percentage in the western (JM02-440, 240 m) and northern (NP94-51, 400 m) Svalbard shelf (Słubowska et al., 2005; Słubowska-Woldeng et al., 2007, 2008). Positions of cores 23258-2, NP94-51, and JM02-440 are shown in Fig. 1.
4.1.5), but also warm-water forms such as *Globorotalia truncatulinoides*, *Globoturborotalita tenella*, and *Globigerinoides ruber*. It should be noted that these are just single tests, which were possibly advected to the shelf by reversed bottom currents (see Section 2.1). Their presence in the outer shelf sediments may thus indicate a close interaction between the shelf and the open sea water masses during postglacial times.

4.3. Correlation of the Late Holocene IRD peaks

The highest concentrations of IRD Peaks 5–7 in PS51/154 are centered at 5, 3.5, and 0.9 ka (Figs. 6 and 7). The precise age of the youngest peak in core PS51/154 remains uncertain because of low sedimentation rates, and the absence of reliable dates younger than 3.3 ka (Table 2). But the much better resolution of core PS51/159...
allows for a more precise age determination of these younger IRD peaks. Accordingly, the upper three IRD Peaks d–f date back to 5.1–5.6, 2.8–3.7, and 1.8–2.2 ka, with the highest concentrations centered at 5.4, 3, and 2 ka. The two older peaks (b, c) in core PS51/159 have the highest concentrations at 7.2 ka and 6.4, respectively. It might be concluded that enhanced IRD input at both sites started around 7 ka. Further development of ice-rafting shows fairly good correlation between both investigated sites. The highest input of IRD took place during the period between c. 5.4 and 2 ka. The better resolved IRD record of PS51/159 demonstrates a sub-millennial periodicity. The peaks are centered at c. 7.2, 6.4, 5.4, 3, and 2 ka and might be taken as an indication of climate-induced changes in both iceberg production on Severnaya Zemlya as well as wind-controlled changes in ice drift pattern.

5. Discussion: paleoenvironmental implications

Below we present reconstructions of the paleoenvironment for certain time slices and a comparison with evidence from outside the Laptev Sea.

A common problem in Arctic paleoenvironmental reconstruction is the difficulty to distinguish between the two major transporting agents for coarse-grained sediments, namely, sea ice and icebergs both of which contribute to glacial marine sedimentation, especially at continental margins (Nürnberg et al., 1994; Darby et al., 2006; St. John, 2008; Polyak et al., 2010). When reconstructing the history of sea ice and iceberg-rafting for the western Laptev Sea in relation to other climate-driven environmental changes, primarily the postglacial sea-level rise, coast retreat and variability in freshwater runoff, we assume that sea-ice cover was always present in this high Arctic region during its long and cold winter season, but varied during the relatively short and warm summer months. This assumption is also justified by an almost continuous background signal of IRD at both studied sites (Fig. 7). Prominent IRD peaks in our cores are therefore attributed to iceberg-rafting events (Knies et al., 2001; Darby et al., 2006). It is further assumed that in the modern Arctic Ocean with its vast and shallow shelves the entrainment of coarse-grained sediment particles into sea ice largely occurring during stormy autumn periods (see Section 2.2) is more effective than during glacial and deglacial epochs, when these shelves were either non-existent or narrow and steep (Dethleff et al., 1993). This might be another argument for the mainly iceberg-rafted origin of the oldest IRD peaks in the studied core from the continental slope.

Yet another common problem which often limits the paleo-reconstructive significance of marine sediment cores from the more extreme sea-ice covered Arctic regions is a general scarcity of calcitic foraminifers. For the Laptev Sea, even though total foraminiferal abundance is rather low compared to more open-ocean regions farther west (Nørgaard-Pedersen et al., 1998; Hald et al., 1999; Koç et al., 2002; Nørgaard-Pedersen et al., 2003; Ślubowska et al., 2005) we assume that the low abundance but fairly good preservation of the microfossil carbonate is indicative of a generally low productivity in this ice covered region (Taldenkova et al., 2008, 2009).

5.1. Late Glacial and early deglaciation, prior to 15.4 ka

Although the oldest IRD Peaks 1–4 in core PS51/154 (Figs. 3, 6 and 7) could not be directly dated because of the lack of fossils, the extrapolated age of 17.6 ka indicates that these peaks belong by an almost continuous background signal of IRD at both studied sites (Fig. 7). Prominent IRD peaks in our cores are therefore attributed to iceberg-rafting events (Knies et al., 2001; Darby et al., 2006). It is further assumed that in the modern Arctic Ocean with its vast and shallow shelves the entrainment of coarse-grained sediment particles into sea ice largely occurring during stormy autumn periods (see Section 2.2) is more effective than during glacial and deglacial epochs, when these shelves were either non-existent or narrow and steep (Dethleff et al., 1993). This might be another argument for the mainly iceberg-rafted origin of the oldest IRD peaks in the studied core from the continental slope.

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a morainic ridge of apparently LGM age at the entrance to Vilkitski Strait (Polyak et al., 2008). The configuration of glaciogenic deposits on land and on the sea floor would imply a short-lived blockage of the river drainage in the eastern Kara Sea (Polyak et al., 2002, 2008). The exact timing of the northeastern Kara Sea ice sheet extension is not yet determined, and there is still the possibility that it was formed during deglaciation when the main Barents–Kara ice sheet became smaller and gave way to moisture supply from the west (Polyak et al., 2008). It might be provisionally assumed that IRD Peaks 1 and 2 in core PS51/154 were correlative in time to the existence of ice coverage in the northeastern Kara Sea, and the subsequent concretion-formation (meltwater) event corresponded to the local meltwater input which occurred after the short-term ice dam disappeared.

In marine sediments from other cores around Severnaya Zemlya – albeit with rather uncertain age models – the evidence about local ice caps of the LGM age is contradictory (Fig. 1a; Knies et al., 2000, 2001). In core 2741 (2350 m water depth) no IRD was accumulated during the last deglaciation probably implying a perennial sea-ice cover. In adjacent core 2742 (1850 m water depth) a thin layer of diamicton with large dropstones dates back to >12.5 14C ka (c. 14 ka). It is supposed to document either the retreat of small ice caps on Severnaya Zemlya from the coastline to inner fjords or iceberg discharge from the retreating Barents–Kara ice sheet. Cores 2782 and 2778 from 340 m water depth show only minor IRD peaks which are not supportive of an ice advance to the outer coastline. Sediment sequence of core 2719 from 130 m water depth in the western Vilkitski Strait contains a laminated (meltwater?) layer overlain by an IRD spike which both are older than 11.7 14C ka. The laminated layer is supposed to be correlative to the diamicton in core 2742. Summarizing, in the cores around Severnaya Zemlya there is also some evidence for the extension of local ice caps to the coastline which most likely occurred near the end of maximum glaciation and/or during the earliest deglaciation.

Because IRD Peaks 3 and 4, which postdate the major local meltwater event at c. 16.8–17.2 ka, not only differ from the first two peaks in grain abundance, the very little contribution of phylites to the total amount of IRD may also imply a different provenance. Perhaps the ice caps on Severnaya Zemlya already had retreated from the coastline at this time because of increasing insolation and summer melt. Terrestrial evidence from the Laptev Sea region would indeed suggest a distinct climatic turnover to warmer summers starting at about 18 ka, although overall climate conditions remained dry with very cold winters until about 14 ka (Hubberten et al., 2004).

The sharp decrease in IRD input after Peak 4 (c. 16.2 ka) gives evidence for the retreat of the Barents–Kara ice caps from the shelf break and cessation of iceberg production followed by the final release of meltwater probably reflected in core PS51/154 by a small, but distinct vivianite spike. Another assumption could be that IRD Peaks 3−4 simply reflect downslope displacement of alluvial and nearshore material, given the predominance of quartz grains of which many are rounded. However, this explanation is unlikely because further up the core at the depth of 530 cm (15 ka) there is clear evidence of a slump event confirmed by the presence of brackishwater and even freshwater microfossils (Fig. 3; Taldenkova et al., 2008). This layer consists of 60–70% of sand and is devoid of coarse fragments. The sediment unit with the small vivianite spike is distinguished in the record by almost complete absence of any coarse debris, thus probably evidencing a dense sea ice cover. Similarly, during the major concretion-formation event the amount of IRD was also considerably reduced.

In general, it looks as if periods of IRD input (Peaks 1−4) document more seasonally open-water conditions with drifting icebergs, whereas subsequent periods of meltwater-induced surface stratification had a more severe sea-ice cover in summer. However, for the long Arctic winter there probably also existed at this time a narrow fast ice cover and frequently occurring nearshore polynya. Given the rather steep slope and the high probability of predominantly offshore winds blowing from the cold permafrost mainland (Fig. 8a), such conditions resembled LGM conditions reconstructed for the northwestern edge of the Barents Sea shelf, where persistent katabatic winds blowing from the ice sheet maintained open coastal polynya with Atlantic water advection (Nørgaard-Pedersen et al., 1998, Knies et al., 1999; Nørgaard-Pedersen et al., 2003).

An interesting feature recorded in core PS51/154 is a thin layer enriched in taxonomically diverse small-sized subpolar planktic foraminifers together with some benthic microfossils including C. neoteretis (Figs. 6 and 7). The abundance of subpolar planktic foraminifers in this layer having an age of c. 17.2 ka is 150 times higher than anywhere upcore. This finding may indeed point to the episodically strong advection of Atlantic-derived waters from farther upstream in the Nordic seas facilitated by upwelling in coastal polynyas (Bauch et al., 2001a). Our data indicate a farther eastward penetration of cold subsurface Atlantic waters into the Arctic than previously interpreted for the LGM (Knies et al., 1999; Nørgaard-Pedersen et al., 2003; Wollenburg et al., 2004; Rasmussen et al., 2007) or early deglaciation (Ślubowska-Woldengen et al., 2008) in the Svalbard region. Rare subpolar planktic foraminifers which are occasionally present in the lower, IRD-rich sediments of core PS51/154 dating back to c. 17.2−16 ka, show that the paleoenvironment remained unstable at this time, with periodic upwelling of Atlantic-derived waters facilitated through coastal polynyas.

5.2. Late deglaciation and Early Holocene, 15.4−7 ka

The period from the early deglaciation to the middle Holocene encompasses the major transgression-induced transformations of the Laptev Sea continental margin from a glacial to a modern-type environment (Fig. 8a–c). The ongoing climate amelioration and melting of ice caps on Severnaya Zemlya and the Barents–Kara seas shelf resulted in a cessation of iceberg-rafting in the western Laptev Sea after 16 ka. From then on, and until 7 ka, the uniformly low IRD content indicates the presence of a seasonal sea ice cover. The IRD content was also low on the outer shelf − the only small peak found at about 11.7 ka corresponds to the times when the studied site was in the neashore zone and probably under fast ice cover (see Section 4.2.4). In the farther western regions with higher moisture supply IRD input from the decaying Barents–Kara ice sheet and glaciers in the fjords continued until the Early Holocene (Fig. 7b; Sarthein et al., 2003; Hald et al., 2004).

Continuous presence and high relative abundances of C. neoteretis and subpolar planktic foraminifers on the Laptev Sea continental slope between 15.4 and 12 ka (Fig. 7d) indicate the strongest subsurface inflow of Atlantic-derived waters. This postglacial inflow of subsurface Atlantic-derived water is well documented in the regions to the west from the Laptev Sea. It has been shown that the troughs of the western and northern Barents–Kara shelf became deglaciated and then invaded by Atlantic-derived water some time between 16 and 13 ka (Lubinski et al., 2001; Koc et al., 2002; Ślubowska et al., 2005; Ślubowska-Woldengen et al., 2007, 2008). Around Svalbard, highest relative abundances of C. neoteretis indicate the presence of subsurface Atlantic-derived water in the troughs between 15 and 12.5 ka (Fig. 7e). In Franz Victoria and St. Anna troughs the appearance of Atlantic water was intermittent and occurred at c. 15, 13 and 11.5 ka (Lubinski et al., 2001). The high representation of C. neoteretis and O-isotopic composition of foraminiferal tests would suggest that it was mainly the FSBW which occupied the trough areas (Lubinski et al., 2001). The non-glaciated
western Laptev Sea area evidences a similar subsurface Atlantic-derived water inflows for this time interval as demonstrated by a remarkable similarity in the outline of C. neoteretis records in core PS51/154 and core NP94-51 from northern Svalbard shelf (Fig. 7d–e). The difference in the timing of C. neoteretis percentage peaks between 13.5 and 12 ka might be due to different age calibration approaches used. Comparison with the NGRIP ice core temperature record shows that C. neoteretis peaks in the Laptev Sea largely tend to correlate with the intervals of elevated temperatures and, probably, enhanced subsurface advection of Atlantic waters to the Arctic (Fig. 7c–e). However, unlike farther west, environmental conditions in the western Laptev Sea remained glacial-like (similar to the situation shown in Fig. 8a) until about 13–14 ka, i.e. prior to onset of outer shelf flooding. The absence of C. neoteretis at c. 13 ka might be related to a regional episode of surface water freshening previously recorded at the eastern Laptev Sea continental slope (Spielhagen et al., 2005).

The disappearance of C. neoteretis from the sediment record of core PS51/154 after c. 12 ka together with a strong reduction in both contents of subpolar planktic foraminifers and the total abundance of planktic foraminifers (Taldenkova et al., 2008) coincides with the onset of shelf flooding which eventually turned core site PS51/159 on the outer shelf into a nearshore, estuarine setting (Taldenkova et al., 2009). The observed faunal changes at the continental slope are most likely caused by the formation of freshened shelf water masses which mixed with the open sea waters thus creating conditions unfavorable for both groups of microfossils (Lubinski et al., 2001). The close interaction between the shelf and open sea waters is seen in periodically increasing abundance of subpolar planktic foraminifers in the Laptev Sea shelf sediments were reported only from the Late Holocene deposits of its eastern region (Bauch, 1999; Matul et al., 2007) and were tentatively related either to erosion of older marine beds or penetration of transformed Atlantic waters to the shelf. However, a diverse subpolar planktic foraminiferal fauna found in net tows to the north of the Laptev Sea seems indicative of strong advective processes, likely caused by the subsurface inflow of Atlantic waters (Volkmann, 2000).

In the central Nordic seas, the western margins of the Barents shelf and Svalbard an Early Holocene enhancement of the inflow of warmest surface waters is reported for the time after 11 ka lasting until about 6 ka (Bauch et al., 2001a; Sarnthein et al., 2003; Hald et al., 2004, 2007; Rasmussen et al., 2007; Ebbesen et al., 2007; Slubowska-Woldengen et al., 2007, 2008). Because inflowing warm Atlantic water into the Nordic seas suffers substantial heat loss on its way to the north, at the northern Svalbard margin the inflow of Atlantic water is recorded mostly as an increase in salinity and productivity (Slubowska-Woldengen et al., 2008). For the time 11.5–8 ka, the subsurface and bottom waters of the Franz Victoria and St. Anna troughs were apparently occupied by cold BSBW, as suggested by the very low abundance of C. neoteretis and specific changes in the isotopic composition of planktic and benthic foraminifer tests from that area (Lubinski et al., 2001). This apparent discrepancy in temperature between the Nordic seas region and the northern Barents–Kara seas is explained by bathymetric controls. The distribution of Atlantic waters on the Barents shelf during the early to middle Holocene was guided by overdeepening of the northern Barents Sea caused by a thicker glacial load in the north compared to its southern part (Lubinski et al., 2001). Together with specific changes in atmospheric circulation more Atlantic water was forced first into the southwestern Barents Sea (Lubinski et al., 2001). The enhancement of the BSBW influence in the northern Barents shelf troughs might be tentatively related to the disappearance of C. neoteretis in core PS51/154 after 12 ka as this site experiences combined influence of both Atlantic water branches.

5.3. Mid–Late Holocene, 7 ka–recent

Since c. 7 ka the regional sea-level of the Laptev Sea was already close to its modern position (Fig. 8c; Bauch et al., 2001b) which caused the main depocenters to shift into the inner shelf, thereby leading to a strong reduction in sedimentation rates, especially on the continental slope (Figs. 3 and 5). By this time, our core sites were already situated in modern-like water depths, as is also suggested by the taxonomic change in the benthic assemblage (Taldenkova et al., 2008, 2009). Thus, any subsequent environmental changes became increasingly influenced by climate variables other than sea-level-driven factors.

A prominent feature of the Mid–Late Holocene time period in the western Laptev Sea is the millennial-scale input of IRD, after c. 7.2 ka (see Section 4.3; Figs. 6 and 7). The high content of phyllites among coarse fragments found in both cores implies re-growth of local ice caps on Severnaya Zemlya caused by climate cooling due to insolation decrease. The actual ages of ice caps on Severnaya Zemlya remains controversial. For instance, the estimated basal age of the recently drilled Akademii Nauk ice cap on Komsomolets Island varies from Bølling-Allerød (Kotlyakov et al., 2004) to as young as 2500 years (Fritzsche et al., 2005). At present it is difficult to resolve whether the ice caps completely disappeared during the Early Holocene climate warming when northern insolation was very high or whether they had just retracted into the inner regions of the islands. It seems evident, however, that a substantial re-growth must have occurred, reaching the coastline around 7 ka when we observe enhanced iceberg activity in our cores. Upper sections of marine sediment cores from around Severnaya Zemlya (Fig. 1a) also show increased IRD contents (Knies et al., 2000, 2001) confirming our interpretation of an ice cap expansion since the middle Holocene due to progressive climate cooling in the high-northern latitudes (Bauch et al., 2001a).

A general cooling trend is also recorded along the western Barents Sea margin and has been interpreted as decrease in heat advection by Atlantic water during times of diminishing northern insolation intensity (Fig. 7b; Sarnthein et al., 2003; Hald et al., 2004; Rasmussen et al., 2007; Ebbesen et al., 2007; Slubowska-Woldengen et al., 2007, 2008; Miller et al., 2010). In the Barents Sea, the influence of Atlantic water during the Mid–Late Holocene seems more complex due to regional isostatic changes and local water mass developments (Lubinski et al., 2001; Duplessy et al., 2001, 2005). There is evidence that increased inflow of FSBW into the northern troughs (mainly Franz Victoria Trough) occurred during the later Holocene (Lubinski et al., 2001), which correlates in time with the re-appearance of C. neoteretis and subpolar planktic foraminifers after 5 ka in the western Laptev Sea. Thus, the co-occurrence of strongly increased IRD input implies that the existence of Atlantic water was also responsible for the moisture necessary to facilitate re-growth of ice caps and enhanced iceberg production on Severnaya Zemlya during times of progressive cooling.

Since modern-type hydrological and circulation patterns became established in the Late Holocene, the observed minor fluctuations in the flux of Atlantic water to the Barents Sea are supposedly related to atmospheric forcing, i.e. millennial-scale variability in the intensity of westerlies associated with the NAO (Duplessy et al., 2005). Similarly, the centennial- to millennial-scale climate-induced changes in the Arctic, like the shifts in sea-ice drift pathways, freshwater discharge and its spatial distribution, are controlled by the changes in the phase of AO/NAO (Darby and
6. Conclusions

- Two major periods of iceberg-rafting corresponding to essentially very different paleoenvironments were recorded in the western Laptev Sea. The first interval is recognized in sediments older than 15.4 ka. The second interval is recorded in Mid–Late Holocene deposits younger than 7 ka. The time in between is characterized by conditions showing less intense iceberg activity and reduced seasonal sea ice cover.

- The glacial-like environment of the upper continental slope prior to 15.4 ka represented a cold-water marine setting, with water depths of 100–150 m, and a closely located coastline on exposed permafrost shelf. IRD-rich sediments of this age consist of four IRD peaks and two sediment layers enriched in authigenic vivianite and rhodochrosite concretions. The latter are indicative of anaerobic bottom water conditions produced by strong water stratification likely due to meltwater input. The major concretion-formation (meltwater) event occurred c. 16.8–17.2 ka after the second IRD peak, and the smaller event c. 15.6–16.2 ka postdates the fourth IRD peak. IRD peaks document more seasonally open-water conditions with drifting icebergs, whereas subsequent periods of meltwater-induced stratification faced more severe sea ice covered environment in summer. In winter, offshore winds blowing from the cold permafrost mainland maintained open-water leads, so-called coastal polynas.

- The two oldest IRD peaks are supposed to be of Late Glacial age (c. 17.6–17.2 ka) probably correlative in time with the short-term Barents–Kara ice sheet extension in the northeastern Kara Sea. The highest proportion of phyllites from Severnaya Zemlya indicates that local ice caps facing the western Laptev Sea reached the coastline and produced icebergs. The subsequent strong meltwater event might be a result of freshwater release after the retreat of the northeastern end of the Barents–Kara ice sheet. The two younger IRD peaks (c. 16.9–16.2 ka) are almost devoid of phyllites indicating that the local ice caps on Severnaya Zemlya retracted inland due to increasing insolation and summer melt, whereby icebergs from the decaying Barents–Kara ice caps contributed to the glaciomarine sedimentation in the western Laptev Sea.

- Already prior to 15.4 ka Atlantic-derived waters occasionally reached the studied site in the western Laptev Sea continental slope probably being facilitated by upwelling through the coastal polynya system. A strong inflow and upwelling, evidenced by the exceptionally high spike of small-sized subpolar planktic foraminifers and the first appearance of C. neoteretis, is noted at the end of the second IRD peak (c. 17.2 ka) and before the strongest meltwater event occurred c. 16.8–17.2 ka.

- Subsurface Atlantic-derived waters related to the FSBW inflow to the west of the Laptev Sea were constantly present at the Laptev Sea continental slope between 15.4–12 ka. Disappearance of C. neoteretis and strong reduction in the abundance of subpolar planktic foraminifers at 12 ka suggest establishment of freshened shelf water mass on the flooded outer shelf which also affected the water mass structure on the continental slope. These faunal changes might be in parts a result of the simultaneous enhancement of the BSBW at the expense of the FSBW as observed in the Barents Sea.

- After 7 ka climate cooling and enhanced Atlantic-derived water inflow caused re-growth of ice caps on Severnaya Zemlya as evidenced by IRD peaks in the sediments from both the continental slope and outer shelf. The IRD peaks are centered at 7.2, 6.4, 5.4, 3, and 2 ka with the strongest iceberg production recorded between 5.4 and 2 ka. This millennial-scale variability in IRD input might be related to overregional changes in atmospheric circulation, i.e., variations in the mode of the North Atlantic and Arctic oscillations.

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