Palaeoceanographic and environmental changes in the eastern Fram Strait during the last 14,000 years based on benthic and planktonic foraminifera

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Abstract

Benthic and planktonic foraminifera, stable isotopes and other geochemical and sedimentological parameters have been investigated in a sediment core from Vestnesa Ridge (79°N, NW Svalbard margin, 1300 m water depth) in order to reconstruct the palaeoceanographic and palaeoenvironmental evolution of the eastern Fram Strait during the last 14,000 years. Our multiproxy data and, in particular, our high-resolution benthic foraminiferal record allow a detailed reconstruction of the intermediate water flow in the eastern Fram Strait. During the Bølling-Allerød warm interstadials the high proportion of the benthic Atlantic species Cassidulina neoteretis suggests the presence of chilled Atlantic-derived water at the bottom of the Vestnesa Ridge, while the dominance of the polar planktonic species Neogloboquadrina pachyderma (s) indicates cold polar surface conditions.
During the Younger Dryas cold stadial the inflow of Atlantic-derived water was weakened and of lower temperature compared to the previous interval, as suggested by the increase in the glacio-marine indicator *Cassidulina reniforme* and the lower percentages of *C. neoteretis*. After 11,000 cal yrs BP the abrupt increase in relative abundance of the benthic species *Cibicides wuellerstorfi* and *Oridorsalis umbonatus*, indicates a shift to interglacial conditions with deep convection and generation of cold intermediate water. Warm surface water conditions were finally established at about 10,300 cal yrs BP, when *Turborotalita quinqueloba* became the dominant planktonic species. Generally warmer conditions than today were recorded in the middle Holocene, with a mild cooling at about 8200 cal yrs BP and a more pronounced cool interval with an abrupt decrease in *T. quinqueloba* between 7000 and 6200 cal yrs BP. Cold climatic conditions were established around 4100 cal yrs BP when *N. pachyderma* (s) became again the dominant species and the polar benthic species *Cribroelphidium excavatum* appeared.
1 Introduction

The Fram Strait is the only deep-water gateway to the Arctic, where water masses between the Atlantic and the Arctic Ocean can exchange. Two main surface currents dominate the Fram Strait: warm Atlantic Water to the east and cold polar water to the west (Aagaard et al., 1987). Here we present data from sediment core JM10-330GC from the Vestnesa Ridge, eastern Fram Strait. The core site is located under the flow of Atlantic Water in the vicinity of the Arctic Front (Fig. 1), which makes it an ideal location for a detailed reconstruction of the distribution of Atlantic Water and oceanic fronts at high latitude in the past. Several studies have reconstructed climatic and environmental changes of the eastern Fram Strait and western Svalbard margin covering the last deglaciation and the Holocene (e.g., Hald et al., 2004; Ebbesen et al., 2007; Rasmussen et al., 2007; Müller et al., 2009; Werner et al., 2013; 2016; Zamelczyk et al., 2012; Aagaard-Sørensen et al., 2014a, b), but mostly with proxies for surface water conditions. Only a few have presented records of benthic foraminiferal distribution patterns. Benthic foraminifera are indicators of the amount and quality of food reaching the seafloor as well as bottom currents, water mass properties and sedimentary conditions (e.g., Gooday, 1988; Murgese and de Decker, 2005; Murray, 2006 and references therein; Jorissen et al., 2007; Rasmussen and Thomsen, 2017). Here we present a detailed micropalaeontological investigation based on the distribution patterns of benthic and planktonic foraminifera, together with sedimentological and geochemical data, in order to reconstruct the palaeoceanographic and palaeoenvironmental evolution of the eastern Fram Strait during the last 14,000 years.
2 Study area

2.1 Geological setting

The study area is the Vestnesa Ridge, an elongated sediment drift offshore the western Svalbard margin at about 79°N in the eastern Fram Strait (Figs 1, 2a). The crest of the Vestnesa Ridge is perforated by pockmarks (Vogt et al., 1994; Hustoft et al., 2009; Fig. 2b), where seepage of gas and fluids occurs or has occurred (Judd and Hovland, 2007). Active venting has recently been observed on the eastern part of the ridge as acoustic gas flares (Hustoft et al., 2009; Bünz et al., 2012). Sediment core JM10-330GC has been retrieved from the deeper (ca. 1300 m water depth), western part of the ridge within a presently inactive pockmark (Bünz et al., 2012; Consolaro et al., 2015) (Fig. 2b).

2.2 Modern oceanographic setting

Two main surface currents dominate the Fram Strait: the northward flowing West Spitsbergen Current (WSC) to the east and the southward flowing East Greenland Current (EGC) to the west (Aagaard et al., 1987) (Fig. 1). The relatively warm and saline WSC (T ≥2°C and S ≥35 PSU; Hopkins 1991) is the northern branch of the North Atlantic Current (NAC). The WSC is the major source of heat and salt to high northern latitudes, and contributes to the generation of deep waters in the Nordic Seas (Aagaard et al., 1985).

Conductivity-Temperature-Depth (CTD; Fig. 3) data collected in June 2010 near the coring site (Fig. 2b) show the relatively warm and salty Atlantic Water (AW) underneath a thin, mixed layer of surface water of low salinity. The AW overlies the Greenland Sea Intermediate Water (GSIW), which is generated from convection in the Nordic Seas and represents one of the deeper components of the WSC (Aagaard et al., 1987). Cold, less saline and ice-loaded polar water (T 0-2°C, S 33–34.4 PSU; Hopkins, 1991) is transported southwards in the western Fram Strait along the East Greenland coast into the North Atlantic Ocean by the
EGC, and into the Barents Sea by the East Spitsbergen Current (ESC) (Fig. 1). Where polar water and Atlantic Water meet they mix and form Arctic water masses (T ~ 0.5°C, S ~34.8 PSU; Hopkins, 1991) characterised by seasonal sea-ice cover. The difference in temperature, salinity and sea-ice cover between the three main surface water masses creates steep climatic gradients: the Arctic Front between Arctic and Atlantic Waters and the Polar Front between polar and Arctic water masses (Hopkins, 1991). The Arctic Front is associated with the maximum extension of the sea-ice margin in winter, while the Polar Front corresponds to the average summer sea-ice margin (Vinje, 1977; NSIDC data, 2016) (Fig. 1). The area between the two fronts defines the marginal ice zone (MIZ), an area with very high seasonal productivity (Smith et al., 1987; Gradinger and Baumann, 1991; Hirche et al., 1991). The seasonal ice retreat and associated primary production is strictly dependent on the strength and temperature of the inflowing Atlantic Water (e.g., Knies et al., 1999, 2000; Nørgaard-Pedersen et al., 2003; Wollenburg et al., 2004). The outer Vestnesa Ridge area is now free of sea ice during most of the year, but is still close to the modern winter ice margin (Fig. 1) (NSIDC data, 2016).

### 3 Material and methods

Gravity core JM10-330GC (79.13°N, 5.6°E; 420 cm long) was retrieved during a cruise with the RV Jan Mayen (now RV Helmer Hanssen) in June 2010, from 1297 m water depth (mwd) at a pockmark located on the western part of the Vestnesa Ridge (Figs 1, 2). The handling and sampling procedures for JM10-330GC have already been described by Consolaro et al. (2015). Visual inspection and X-ray images revealed no disturbances from or evidence of presence of gas. The colour scan images of the core, together with the lithological log, the
magnetic susceptibility, the concentration of IRD >1 mm and the sedimentation rate have
been published previously by Consolaro et al. (2015).

The samples were collected in 1-cm slices every 5 cm, freeze-dried, weighed, wet-sieved over
mesh sizes of 63 µm, 100 µm, and 1 mm, dried at 40°C and weighed again. Eighty-five
samples were analysed giving an average time resolution of 165 years. Mineral grains,
biogenic silica remains (diatoms, radiolarians and sponge spicules) and pyritized burrows
were counted in the 100–1000 µm size fraction.

The weight percentages (wt%) of total carbon (TC), total organic carbon (TOC) and calcium
carbonate (CaCO₃) were determined every 5 cm. TC and TOC were measured from bulk
samples using a Leco CS-200 induction furnace. TC was measured directly on the dried and
crushed samples, while TOC was measured after the inorganic carbon was removed with HCl
(10%) at room temperature. The CaCO₃ content (wt%) was calculated using the equation of
Espitalié et al. (1977):

\[ \text{CaCO}_3 = \left( \text{TC} - \text{TOC} \right) \times 100 / 12 \]

Oxygen and carbon isotope data measured on the planktonic foraminiferal species
*Neogloboquadrina pachyderma* sinistral (s) and the benthic species *Cassidulina neoteretis*
have been previously published by Consolaro et al. (2015), while the isotope data generated
using *Cibicides wuellerstorfi* are new (Supplementary Table S1). Stable isotope analyses were
carried out at the Leibniz-Laboratory for Radiometric Dating and Isotope Research in Kiel,
Germany, using a Finnigan MAT-253 mass spectrometer with Kiel IV system (analytical
precision of ±0.05‰ for δ¹³C and ±0.1‰ for δ¹⁸O estimated by certified standard NBS-19).
Isotope results are reported in standard delta notation relative to Vienna Peedee Belemnite
(VPDB). The δ¹⁸O isotopic values were corrected for ice volume effect (δ¹⁸O_{IVC}) with a
correction of 0.11‰ δ¹⁸O per ten meters sea level change (Fairbanks, 1989). The values
obtained for C. wuellerstorfi were corrected for disequilibrium with seawater by adding 0.64‰ (Duplessy et al., 1980).

For this study, benthic and planktonic foraminifera were picked from the 100 µm to 1 mm size fraction, counted and identified to species level following the guidelines from Knudsen (1998). A minimum of 300 benthic and 300 planktonic foraminifera specimens were counted for each sample. When necessary, the residues were split and the total number of foraminifera calculated. In the lower part of the record (between 420 and 280 cm) one sample for planktonic and eight samples for benthic foraminifera contained between 150 and 270 specimens. These data were, therefore, treated with caution but they were still included in the calculations for relative and absolute abundances (Supplementary Table S2). The taxonomy employed follows Loeblich and Tappan (1987). According to Darling et al. (2006), if the percentage of right coiling *Neogloboquadrina pachyderma* at high-latitudes is less than 3%, they should be considered as aberrant morphotypes of *N. pachyderma* sinistral (s) and should be referred to as *N. pachyderma* dextral (d). Since the percentage of right-coiling specimens in our sediment samples is close to 3% (lower in the early part of the record and slightly higher during the Holocene), we decided not to distinguish *N. pachyderma* (d) from *N. incompta* and named all the right coiling specimens as *N. pachyderma* (d).

Relative abundance (%) and flux (no. specimens/cm²/yr) of the most abundant species were calculated for each sample (Supplementary Table S2). Percentages of the planktonic species *N. pachyderma* (s) and *Turborotalita quinqueloba* have been published previously by Consolaro et al. (2015). Flux, or foraminiferal accumulation rates, were calculated using the formula of Ehrmann and Thiede (1985):

\[
\text{Flux} (\text{#/cm}^2/\text{yr}) = (\text{absolute abundance (#/g}) \times \text{dry bulk density (g/cm}^3) \times \text{SAR (cm/ka)})/1000,
\]
where \( # \) is the number of foraminiferal specimens, and SAR is the sediment accumulation rate. Dry bulk density was calculated based on sediment water content and wet bulk density, acquired by GEOTEK multisensor core logging.

The age model has been published by Consolaro et al. (2015) and is based on eight AMS-\(^{14}\)C dates. In the following, ages are presented as cal yrs BP or cal kyrs BP.

### 4 Results

The record of core JM10-330GC is subdivided into five time intervals based on the age model, the benthic and planktonic foraminiferal distribution, the lithological features and the stable isotope records (Figs 4–7). The first interval coincides with the upper part of the Bølling interstadial and the Allerød interstadial (from the base of the core at 14,120 to 12,850 cal yrs BP). The second interval corresponds to the Younger Dryas (YD) stadial (12,850–11,650 cal yrs BP) and is characterised by a greenish-grey, sandy layer at the base (12,700–12,500 cal yrs BP). The early Holocene interval starts at 11,650 cal yrs BP and consists of fine-grained, silty clay with high concentration of biogenic siliceous remains, especially diatoms (*Coscinodiscus* spp.), termed diatom-rich mud (11,100–8800 cal yrs BP). The middle Holocene interval is defined by high % of *T. quinqueloba* and low % of *N. pachyderma* (s) (8800–4100 cal yrs BP). The late Holocene is defined by the presence of the benthic foraminiferal species *Cribrorhphidium excavatum* and high relative abundance of *N. pachyderma* (s) from 4100 to 150 cal yrs BP (age of the top of the core). More details about the stratigraphy of the core JM10-330GC, including the sandy layer and the diatom-rich layer, can be found in Consolaro et al. (2015).
4.1 Stable isotope analysis

Stable isotope analyses ($\delta^{18}$O and $\delta^{13}$C) of the planktonic foraminifera *N. pachyderma* (s) and the infaunal benthic foraminifera *C. neoteretis* (Fig. 5) have been previously published in Consolaro et al. (2015). Their $\delta^{13}$C records show two Carbon Isotope Excursions (CIEs) attributed to influence by seeping gas (for details see Consolaro et al., 2015). Excluding the CIE events, both $\delta^{13}$C records present low values in the Younger Dryas and in the early Holocene (Fig. 5). The new Holocene $\delta^{13}$C record of *C. wuellerstorfi* shows low values in the early Holocene with two negative peaks at 11,000 and 10,000 cal yrs BP, while in the middle and late Holocene the values are generally higher, with small negative peaks at 8800 and 8000 cal yrs BP and a gradual decrease in the last 3000 years (Fig. 5). The $\delta^{18}$O record of *C. wuellerstorfi* shows relatively high values in the diatom-rich layer, and peaks of high values around 8000 and 4500 cal yrs BP (Fig. 5).

4.2 Foraminiferal abundance and distribution

The planktonic foraminifera assemblage is dominated by two species: *Neogloboquadrina pachyderma* (s) (mean value 55%) and *Turborotalita quinqueloba* (mean value 40%) (Fig. 6). Other less abundant species are *N. pachyderma* (d), *Globigerinita uvula*, *Globigerinita glutinata* and *Globigerina bulloides*. Before 10,800 cal yrs BP *N. pachyderma* (s) dominates the assemblage with *T. quinqueloba* as a minor component. After 10,300 cal yrs BP, in the early and middle Holocene, *T. quinqueloba* dominates the assemblage. The percentage of *N. pachyderma* (s) shows a gradual increase starting at about 4900 cal yrs BP. The number of planktonic foraminifera per gram of dry sediment and the total flux of planktonic foraminifera show similar trends with very low abundance in the lower part of the record, with the
exception of the sandy layer at 12,700 cal yrs BP, and a sharp increase at about 9800 cal yrs BP (Fig. 6g).

A total number of 73 benthic species were found (Appendix). The most abundant species are: Cassidulina neoteretis, Cassidulina reniforme, Melonis barleeanus, Lobatula lobatula, Cibicides wuellerstorfi, Epistominella arctica, Cribroelphidium excavatum, Oridorsalis umbonatus, Ioanella tumidula, Stainforthia loeblichi, Nonionella iridea, and Nonionellina labradorica (Fig. 7). Cassidulina neoteretis dominates in the Bølling-Allerød interstadials (Fig. 7a). The sandy layer at about 12,700 cal yrs BP, contains an assemblage with dominance of M. barleeanus, O. umbonatus and I. tumidula, while just above S. loeblichi and N. iridea show major peaks (see discussion in section 5.2) (Fig. 7c, h–k). The diatom-rich layer (11,100–8800 cal yrs BP) is dominated by Cassidulina reniforme together with Cibicides wuellerstorfi, M. barleeanus and Oridorsalis umbonatus, with Epistominella arctica as a minor component (Fig. 7). After 4100 cal yrs BP C. wuellerstorfi and E. arctica show a decrease in their relative abundance and Cribroelphidium excavatum appears (Fig. 7g). The number of benthic foraminifera per gram dry sediment and the total flux of benthic foraminifera present low values in the deglacial part of the record and gradually increase after 11,000 cal yrs BP (Fig. 7p).

5 Reconstruction of palaeoceanographic and palaeoenvironmental conditions in the last 14,000 years

5
5.1 Upper part of Bølling and Allerød interstadials (14,120–12,850 cal yrs BP): cold surface water conditions with Atlantic-derived water at the bottom

The Bølling and Allerød interstadials (B-A) are described as warm periods in northern Europe and from the start of the Bølling period (14,650–14,025 cal yrs BP), the Svalbard-Barents Sea ice sheet retreated rapidly (e.g., Vorren and Kristoffersen, 1986; Mangerud et al., 1992; Lubinski et al., 2001). The oldest sediments in core JM10-330GC contain a relatively high concentration of sand (>63 µm) and IRD (>100 µm and >1 mm) deposited rapidly and indicating unstable climatic conditions characterised by melting and re-advancing phases of glaciers and sea-ice, as also described by e.g., Landvik et al. (1998) (Fig. 4a, c). Low values of planktonic and benthic foraminifera total flux can indicate low primary productivity (Figs 6g, 7p) supported by low %CaCO$_3$ (Fig. 4e). The presence of aragonitic pteropods indicates good preservation, as aragonite is more vulnerable to dissolution than calcite (e.g., Cherns and Wright, 2009) (Fig. 6h). The dominance of the polar planktonic foraminiferal species *N. pachyderma* (s) (mean value 90%; Fig. 6a) indicates polar conditions and presence of cold Arctic and polar surface water (Table 1). This is supported by the high planktonic δ$^{18}$O$_{IVC}$ values that also indicate cold conditions at the subsurface (Fig. 5). Similar assemblages with very high relative abundance of *N. pachyderma* (s) and low planktonic flux have previously been reported during the B-A from the western Svalbard margin (Ebbesen et al., 2007; Rasmussen et al., 2007; Aagaard-Sørensen et al., 2014a; Fig. 8) and the south-western Barents Sea (Sarnthein et al., 2003; Aagaard-Sørensen et al., 2010).

The benthic foraminiferal assemblage is dominated by *C. neoteretis* (mean value 55%), which indicates the presence of chilled Atlantic-derived Water (Table 1) (Fig. 7a). Such high relative abundance of *C. neoteretis* occurs in the modern sediments of the Barents and Kara seas in areas with subsurface inflow of Atlantic Water below a layer of polar water at the surface
(Lubinski et al., 2001). It is also common in sediments overlain by Atlantic Intermediate Water in Greenland fjords (Jennings and Weiner, 1996; Jennings et al., 2006) and in the Arctic Ocean (Wollenburg and Mackensen, 1998). The very high proportion of *C. neoteretis* during the B-A suggests a very strong inflow of chilled Atlantic-derived water at the bottom. Similar conditions, based on a high proportion of *C. neoteretis*, have been suggested by Lubinski et al. (2001) for the same time interval in the northern Barents Sea. *Melonis barleeanus* is the second most abundant species during the B-A (22%; Fig. 7c) possibly indicating high sedimentation rates with steady food supply (Table 1), and it has been associated with the presence of Atlantic-derived water in the Arctic (Polyak et al., 2002). Less abundant species are *C. reniforme* (11%; Fig. 7b), an Arctic species common in cooled Atlantic Water and in glacio-marine environments (Table 1), and *Lobatula lobatula* (4%; Fig. 7d), an indicator of strong bottom current activity (Table 1).

Our reconstruction (Fig. 10) agrees well with previous findings from the Western Svalbard margin (Rasmussen et al., 2007; Ślubowska-Woldengen et al., 2007; Aagaard-Sørensen et al., 2014a), the northern Svalbard margin (Koç et al., 2002; Ślubowska et al., 2005), the marginal Arctic Ocean (Wollenburg et al., 2004) and the northern Barents Sea (Polyak and Solheim, 1994; Lubinski et al., 1996, 2001; Klitgaard-Kristensen et al., 2013), where Atlantic Water was found to be continuously present below polar surface water from the Bølling interstadial onwards.

### 5.2 Younger Dryas stadial (12,850–11,650 cal yrs BP): a not very cold interval with a sandy layer associated with a meltwater event

The low planktonic and benthic δ¹⁸O values during the YD correlating with the deposition of the sandy layer are most likely the result of reduced salinity due to a local meltwater event
(Fig. 5). Similar short local melt water pulses during the YD have previously been recorded from Storfjorden Trough (Rasmussen et al., 2007), central (Zameczyk et al., 2012) and western Fram Strait (Bauch et al., 2001), and eastern Greenland shelf (Jennings et al., 2006). A slight decrease in sand (>63 µm) and IRD (>100 µm and >1 mm) with lower sedimentation rate above the sandy layer (Fig. 4), can probably be linked to increased sea-ice cover and reduced iceberg transport, as previously documented on the western and northern Svalbard margin (Koç et al., 2002; Wollenburg et al., 2004; Ebbesen et al., 2007; Ślubowska-Woldengen et al. 2007; Aagaard-Sørenesen et al., 2014a).

During the Younger Dryas the total flux and concentration of planktonic foraminifera was very low, except in the sandy layer, indicating, together with the low CaCO₃ content, a very low productivity (Figs 4e, 6g). The planktonic assemblage was still dominated by *N. pachyderma* (s), but with slightly lower mean values (86%) compared to the Bølling-Allerød interstadials, indicating slightly less severe conditions (Fig. 6a). Similar surface conditions were also found in records from the western Svalbard margin (Ebbesen et al., 2007; Rasmussen et al. 2007; Aagaard-Sørenesen et al., 2014a; Fig. 8).

The total flux and concentration of benthic foraminifera show here the lowest values of the record, except in the sandy layer (Fig. 7p), confirming that the low productivity also affected the deep benthic community. Before, and during, the deposition of the sandy layer the high relative abundance of *M. barleeanus* (Fig. 7c) and the presence of *L. lobatula* (Fig. 7d) indicate high sedimentation rates and strong current activity, while the peaks of *Oridorsalis umbonatus* and *Joanella tumidula* (Fig. 7h-i) can indicate oligotrophic conditions (Table 1), as also suggested by the very low values of TOC registered during the event (Fig. 4e). The organic-rich top part of the sandy layer was colonised by the opportunistic species *Nonionella iridea* and *Stainforthia loeblichii* (Table 1). These two species probably were able to exploit
the high organic content in the sediment (Fig. 7k, j; Table 1). Above the sandy layer the TOC values increase, possibly indicating a progressive eutrophication of the ecosystem (Fig. 4e). Increased TOC and marine biomarker values during the late YD have been reported in the Fram Strait/Yermak Plateau (Birgel and Hass, 2004; Müller and Stein, 2014). After the melt water event the benthic assemblage is characterised by high percentages of the glacio-marine indicator *C. reniforme* with lower percentages of *C. neoteretis*, *M. barleeanus* and *L. lobatula* (Fig. 7a–d). This assemblage suggests reduced influence of Atlantic Water and increased influence of colder bottom water with weaker current activity (Fig. 10).

The YD cold period is not particularly conspicuous in our data (Fig. 10). The sea surface was still influenced by cold polar water, but with open water conditions that allowed substantial primary production, while the influx of Atlantic-derived Water at the bottom was weaker, with low current velocity and probably of lower salinity (low δ¹⁸O benthic values, Fig. 5). The relatively low planktonic and benthic δ¹³C values are an indication of poor ventilation and probably strong stratification of the water column (Figs 5, 10). Similar low δ¹³C planktonic values have been registered in other western Svalbard margin records during the same interval (Ebbesen et al., 2007; Aagaard-Sørensen et al. 2014a; Fig. 9i, j). These data are consistent with previous studies from the northern and western Svalbard margin that reported seasonal open water conditions and reduced influence of Atlantic Water at the bottom (e.g., Ślubowska et al., 2005; Rasmussen et al., 2007; Ślubowska-Woldengen et al., 2007; Aagaard-Sørensen et al. 2014a; Müller and Stein, 2014). Colder conditions with almost permanent ice cover, very little Atlantic-derived water influence and low-abundance or foraminifera-barren sediments, have been reported from the eastern Svalbard margin and the Barents Sea (Polyak and Solheim, 1994; Hald et al., 1999; Lubinski et al., 2001; Aagaard-Sørensen et al. 2010; Klitgaard et al., 2013).
5.3 Early Holocene (11,650–8800 cal yrs BP): unstable with shift to stable interglacial conditions around 11,000 cal yrs BP in the benthic environment and delayed warming in the planktonic environment

The transition from the YD to the early Holocene is characterised by unstable conditions with presence of IRD and high sedimentation rates reflecting increased sediment transport by melting icebergs and/or sea ice (Fig. 4). Similar surface conditions have been recorded both west and north of Svalbard during this period (Ebbesen et al., 2007; Ślubowska-Woldengen et al., 2007; Aagaard-Sørensen et al., 2014a). In our record the early Holocene is marked by the deposition of a thick (1.1 m) diatom-rich mud layer characterised by abundant Coscinodiscus spp. in the 63–100 µm fraction and high abundance of radiolarian and sponge spicules in the >100 µm fraction (Fig. 4d). Similar diatom-rich mud layers have been observed in the Fram Strait further to the north of our site (Stabell, 1986), in the southern Norwegian Sea (Stabell, 1986; Koç-Karpuz and Jansen, 1992; Birks and Koç, 2002) and western Svalbard margin (Jessen et al., 2010; Aagaard-Sørensen et al., 2014a). The diatom-rich layer is time transgressive and has been attributed to the northward movement of the Polar Front (Jessen et al., 1983; Stabell, 1986). It is considered a chronostratigraphic marker for the western Svalbard slope with an early Holocene age (10,100–9840 cal yrs BP in core JM03-373PC, Jessen et al., 2010; and 10,500–9800 cal yrs BP in core MSM5/5-712-2, Aagaard-Sørensen et al., 2014a). In our record the diatom-rich layer is thicker and more expanded in time (11,100–8800 cal yrs BP), possibly indicating prolonged influence of the marginal ice zone over outer Vestnesa Ridge than in other records further south (Fig. 2).

In the early part of the early Holocene (11,650–10,800 cal yrs BP) the surface conditions were still strongly influenced by cold polar water (71–75% of N. pachyderma (s); Fig. 6a).
Thereafter, a very unstable period occurred with strong fluctuations in the relative abundance of *N. pachyderma* (s) (from 26% to 82%) and *T. quinqueloba* (from 14% to 86%), until warm surface conditions were finally established around 10,300 cal yrs BP, when *T. quinqueloba* became the dominant species (Fig. 6a, b). Similar unstable conditions with fluctuating sea subsurface temperature (SST-100 m) have been observed in the neighbouring core MSM5/5-723-2 until 10,600 cal yrs BP (Werner et al., 2016), but with smaller amplitude in fluctuations of the planktonic assemblages (Fig. 8e). A similar pattern in the reconstructed SST based on Mg/Ca can be observed in core MSM5/5-712-2, but not in the faunal assemblage (Aagaard-Sørensen et al., 2014a; Fig. 8c). Further south in the Storfjord Trough strong fluctuations in the planktonic record are registered earlier between 11,600 and 11,000 cal yrs BP suggesting a ‘Pre-Boreal Type’ event (Rasmussen et al., 2007; Fig. 8a). Despite the differences between the records during the unstable transition from the YD to the Holocene, which might have been caused by local conditions, a common pattern with a clear shift to warmer surface/subsurface conditions occurs between 10,500 and 10,300 cal yrs BP in all records (Fig. 8; Ebbesen et al., 2007; Rasmussen et al., 2007; Aagaard-Sørensen et al., 2014a; Werner et al., 2016). The lowest peak in abundance and flux of *N. pachyderma* (s), together with a peak of *G. uvula* and a sharp increase in planktonic foraminiferal productivity at 9800 cal yrs BP, mark the beginning of Holocene optimum climatic conditions in our record (Fig. 6a, d, g). The pronounced peak of *G. uvula* between 10,000 and 9,500 cal yrs BP, is a common feature in records from the western Svalbard margin (Fig. 8; Ebbesen et al., 2007; Rasmussen et al., 2007; Aagaard-Sørensen et al., 2014a), and can indicate the presence of highly productive surface waters in proximity of the oceanic fronts (Table 1). The concomitant increase in abundance and flux of the relatively thermophile species *N. pachyderma* (d) and *G. bulloides* demonstrates a stronger influence of warm Atlantic Water (Fig. 6c, f), as already observed in the Nordic Seas (Risebrobakken et al., 2011) and in the neighbouring core MSM5/5-712-2.
(Aagaard-Sørensen et al., 2014a). Maximum oceanic heat advection through northward transport of Atlantic Water around 10,000 cal yrs BP (Risebrobakken et al., 2011) led to cessation of iceberg transport, as suggested by the absence of IRD after 10,000 cal yrs BP in our record (Fig. 4c) and in other records along the western Svalbard margin (Ebbesen et al., 2007; Rasmussen et al., 2007, 2014; Jessen et al., 2010; Aagaard-Sørensen et al., 2014a). The planktonic δ18O record shows first warming signals at about 10,300 cal yrs BP and minimum values around 9600 cal yrs BP (Figs 5, 10). A similar pattern with minimum δ18O values after 10,000 cal yrs BP has been observed in the neighbouring cores MSM5/5-712-2 and MSM5/5-723-2 (Aagaard-Sørensen et al., 2014a; Werner et al., 2016; Fig. 9f, h) confirming warm climatic conditions and maximum Atlantic Water advection in the area during this time (Werner et al., 2016). Holocene climatic optimum conditions with higher SST than present have been reported in the western Fram Strait (Bauch et al., 2001), western Svalbard margin (Ebbesen et al., 2007; Hald et al., 2007; Aagaard-Sørensen et al., 2014a; Werner et al., 2016) and Barents Sea (Sarnthein et al., 2003; Berben et al., 2014) at about the same time. The northward heat transport gradually decreased after 10,000 cal yrs BP (Risebrobakken et al., 2011), but conditions warmer than today with high relative abundance of *T. quinqueloba* continued until about 8800 cal yrs BP (Ebbesen et al., 2007; Rasmussen et al., 2007; Aagaard-Sørensen et al., 2014a; Werner et al., 2016; Figs 8, 10).

The species composition of the benthic foraminiferal assemblages also shifted during the early Holocene (Fig. 7). In the beginning, the assemblage was still dominated by *C. reniforme, C. neoteretis* and *M. barleeanus*, but around 11,000-10,800 cal yrs BP, the increase in relative abundances of *C. wuellerstorfi* and *Oridorsalis umbonatus* indicate interglacial conditions similar to the modern environment on the Nordic Seas deeper slopes characterised by well oxygenated deep-waters (Table 1). The appearance of *Epistominella arctica*, an opportunistic phytodetritus feeder (Table 1), can indicate highly variable food productivity...
during the deposition of the diatom-rich layer, as also shown by the high TOC values (Figs 7 and 4e). Cold deep-water conditions and decreased influence of Atlantic-derived water at the bottom is also indicated by the low proportion of *C. neoteretis* between 10,800 and 8200 cal yrs BP (Figs 7a, 10). Both benthic δ¹⁸O records show relatively high values confirming cold bottom water conditions, with some signal of warming at 10,500 and 10,000 cal yrs BP, when the *C. neoteretis* record registers two minima (Fig. 5). Relatively low δ¹³C values in *C. wuellerstorfi* have been observed around 11,000 and 10,000 cal yrs BP in our record (Figs 5, 9n, 10), and before 11,000 cal yrs BP in the neighbouring core MSM5/5-723-2 (Werner et al., 2016; Fig. 9o). Low epibenthic δ¹³C values before 11,000 cal yrs BP have been reported in the Faroe-Shetland Channel by Risebrobakken et al. (2011), who argued that during the maximum of northward Atlantic Water advection some δ¹³C-depleted bottom waters up-welled affecting the planktonic δ¹³C signal in the Nordic Seas. This would explain the low planktonic δ¹³C values registered during the early Holocene in our record and in other records along the western Svalbard margin (Fig. 9i–l; Ebbesen et al., 2007; Aagaard-Sørensen et al., 2014a; Werner et al., 2016).

The shift towards interglacial conditions with the generation of cold intermediate water occurred around 11,000-10,800 cal yrs BP in the benthic environment, while warm surface water conditions finally established around 10,300 cal yrs BP (Fig. 10). Similar results with interglacial conditions established first in the benthic environment and a delayed appearance of Atlantic Water at the surface, has been described from the southwestern Svalbard margin (Rasmussen et al., 2007; Ślubowska-Woldengen et al., 2007).
5.4 Middle Holocene (8800–4100 cal yrs BP): warm conditions with a mild cooling around 8200 cal yrs BP and pronounced cooling between 7000 and 6200 cal yrs BP.

Warm climatic conditions with strong inflow of Atlantic Water at the surface lasted until about 7000 cal yrs BP, as indicated by the high relative abundance of *T. quinqueloba*, the presence of subpolar species and the high concentration and flux of planktonic foraminifera (Figs 6, 10). In contrast to neighbouring cores MSM5/5-712-2 and MSM5/5-723-2, where the ‘8200 cal yrs BP’ climate anomaly is indicated by pronounced shifts in surface and deep-water proxy records (Werner et al., 2013, 2016), we see only a slight increase in flux and relative abundance of *N. pachyderma* (s) (Figs 6a, 8n), two higher peaks in δ¹⁸O values and a δ¹³C minimum in *C. wuellerstorfi* around 8000 cal yrs BP (Figs 5, 9b, n). This could indicate colder conditions and reduced deep-water ventilation, respectively.

A short episode with increased abundance of *N. pachyderma* (s) at the expense of *T. quinqueloba*, and with low planktonic foraminiferal concentration, occurred between 7000 and 6200 cal yrs BP (Figs 6, 8, 9, 10). Temperature lows at about 6900 and 6100 cal yrs BP (Fig. 8c) have been described in core MSM5/5-712-2 (Werner et al., 2013; Aagaard-Sørensen et al., 2014b) and, less pronounced, in core MSM5/5-723-2 (Werner et al., 2016; Fig. 8e). These cooling events have been linked to advances and retreats of the sea-ice margin connected to the movement of the Arctic Front (Werner et al., 2013). A general cooling trend from about 7000 cal yrs BP has been observed in several other records in the Nordic Seas (Sarnthein et al., 2003; Knudsen et al., 2004; Hald et al., 2007; Rasmussen et al., 2007) caused by reduced northward advection of Atlantic Water and cooling of the surface water (Hald et al., 2007; Risebrobakken et al., 2011). This cooling is also seen in our core (Fig. 5) and in general over the western Svalbard margin at about the same time (Ebbesen et al., 2007;
After 6200 cal yrs BP *T. quinqueloba* returns to being the dominant species, although with lower relative abundance compared to the Holocene climatic optimum interval (Figs 6b, 8m). Subpolar species are present and together with the high concentration of planktonic foraminifera can indicate a fairly strong inflow of Atlantic Water and proximity to the oceanic fronts until about 4100 cal yrs BP (Fig. 10). A Similarly high proportion of *T. quinqueloba* has been described in core MSM5/5-712-2 (Werner et al., 2013; Fig. 8j) and in core MSM5/5-723-2 (although with lower abundance; Werner et al., 2016; Fig. 8p). These high relative abundances of *T. quinqueloba* differs from most other studies in the region (e.g., Bauch et al., 2001; Sarnthein et al., 2003; Ebbesen et al., 2007; Hald et al., 2007; Rasmussen et al., 2007; Risebrobakken et al., 2010), and can be attributed to a closer location of the core site to the Arctic Front compared to other studies, or to a better preservation of the more delicate *T. quinqueloba* shells. Similar results have been reported from Kveithola Trough in the western Barents Sea (Berben et al., 2014).

In the benthic environment, *C. wuellerstorfi* has a lower relative abundance than in the early Holocene (around 12%), while the opportunistic species *E. arctica* (10%) together with *M. barleeanus* (7%) and the phytodetritus feeders *Nonionella iridea* (5.8%) increase, pointing to more variable seasonal productivity and proximity to the oceanic fronts (Table 1), as also indicated by the high TOC values (Fig. 4e). The decrease in percentage of *O. umbonatus* towards the end of the time interval and the appearance of *N. labradorica* at the same time, also suggests an increase in productivity (Table 1). Generally high benthic δ13C values can indicate stronger ventilation and deep-sea convection that, together with the increasing trend of the δ13C in the planktonic record, also suggest a general increase in productivity (Figs. 5,
During the 8200 and the 7000–6200 cal yrs BP events only a small decrease in the relative abundance of *C. wuellerstorfi* can be noticed (Fig. 5), suggesting that these cooling events were more pronounced in the planktonic environment (see discussion above).

### 5.5 Late Holocene (4100–150 cal yrs BP): a general cooling trend and a small warming in the last 2000 years

*Neogloboquadrina pachyderma* (s) becomes dominant again in the late Holocene interval, showing a cooling trend started at about 4900 cal yrs BP (Figs 6a, 10). A similar, but more abrupt climate shift occurred around 5200 cal yrs BP in core MSM5/5-712-2 (Werner et al., 2013; Fig. 8c) and more gradually from 5000 cal yrs BP in core MSM5/5-723-2 (Werner et al., 2016; Fig. 8e), showing a common cooling trend culminating with minimum temperatures between 4000 and 3000 cal yrs BP (Werner et al., 2013; 2016; Aagaard-Sørensen et al., 2014b). Increasing planktonic δ¹⁸O values from about 5000 to 3000 cal yrs BP in our record (Figs 5, 10) and in other records from the western Svalbard margin (Fig. 9h–l; Ebbesen et al., 2007; Rasmussen et al., 2007; Werner et al., 2013; 2016; Aagaard-Sørensen et al., 2014b) reflect colder subsurface conditions in the whole area, while increased sea-ice biomarker concentration indicates more severe sea-ice conditions (Müller et al., 2012; Werner et al., 2013). A cooling trend with spreading of polar conditions associated with the so-called Neoglacial starting at about 4000 cal yrs BP has also been reported south of Svalbard and in Svalbard fjords (Sarnthein et al., 2003; Hald et al., 2004, 2007; Rasmussen et al., 2012). The advance of the Polar Front and spreading of polar water caused the disappearance of planktonic foraminifera over the north and west shelf of Svalbard (Ślubowska et al., 2005;...
Ślubowska-Woldengen et al., 2007; Skirbekk et al., 2010; Rasmussen et al., 2014). In sediments from the central and eastern Fram Strait and southwestern Svalbard dissolution became more important, seen as reduced shell weight of planktonic foraminifera (Zamelczyk et al., 2012), poor preservation and absence of pteropods in our record (Fig. 6h), and increased fragmentation of planktonic foraminiferal shells (Rasmussen et al., 2007; Werner et al., 2016).

The cooling trend is also reflected in the benthic environment, where the Arctic-polar species *C. excavatum* appears, reaching its highest relative abundance (about 20%) around 3500 cal yrs BP (Figs 7g, 10; Table 1). The increase in percentage in *L. lobatula* indicates strong bottom current activity (Fig. 7d). Deep-sea convection and good bottom water ventilation are also indicated by the high δ\(^{13}\)C values in *C. wuellerstorfi* (Figs 5, 10). A late Holocene cooling is also evident in several records of the Nordic Seas at lower latitudes (e.g., Birks and Koç, 2002; Calvo et al., 2002; Jennings et al., 2002; Andersen et al., 2004; Kaplan and Wolfe, 2006; Justwan et al., 2008; Ólafsdóttir et al., 2010; Rasmussen and Thomsen, 2015).

The slow increase in the opportunistic species *G. uvula* during the last 3000 years, and especially after 2000 cal yrs BP, points to increased contribution of cold, productive surface waters (Fig. 6d; Table 1). The late Holocene increase of *G. uvula* is a common occurrence at the western Svalbard margin (Fig. 8a, c; Rasmussen et al., 2007; Werner et al., 2013) and western Barents Sea (Berben et al., 2014). Increase in IRD content (Fig. 4c) and in sea-ice biomarker concentration indicates a more extensive sea-ice cover and increased freshwater supply (Müller et al., 2012; Werner et al., 2013; 2016). The gradual decrease in both δ\(^{13}\)C and δ\(^{18}\)O values of *N. pachyderma* (s) in the last 3000 years observed in our record (Fig. 5) confirms a general pattern already described in the neighbouring cores MSM5/5-712-2 and MSM5/5-723-2 (Fig. 9), indicating less ventilated subsurface water with higher temperature, possibly in connection with enhanced Atlantic Water advection (Werner et al., 2013; 2016;...
Aagaard-Sørensen et al., 2014b). These authors explain this apparent contradiction with the presence of a strong pycnocline separating cold surface waters from the warm Atlantic layer below and with the migration of *N. pachyderma* (s) to the deeper Atlantic layer, where conditions were more favourable. Evidence for increased Atlantic Water inflow and strong stratification of the upper water column has been found elsewhere in the Nordic Seas in the same period (Lubinski et al., 2001; Sarnthein et al., 2003; Hald et al., 2007; Risebrobakken et al., 2003; 2011; Berben et al., 2014).

The gradual decrease in relative abundance of *C. excavatum* and the small increase of *N. labradorica* and *C. wuellestorfi* (Fig. 7), together with a slight decrease in δ¹⁸O values in both benthic records during the last 2000 years (Fig. 5), can indicate warmer intermediate waters conditions. Increased inflow of Atlantic Water and periodic stratification at the surface in the last 2000 years has been described in Isfjorden, Svalbard (Rasmussen et al., 2012) and in the western Barents Sea (Chistyakova et al., 2010; Dylmer et al., 2013; Groot et al., 2014). However, the general conditions remained cold and unstable with episodes of reduced salinity and concomitant increases in sea-ice and decreases in surface temperature (e.g., Ślubowska-Woldengen et al., 2007, 2008; Risebrobakken et al., 2010; Werner et al., 2013; Berben et al., 2014).

### 7. Summary and conclusions

Sediment core JM10-330GC from the Vestnesa Ridge (NW Svalbard margin) has been investigated with regard to planktonic and benthic foraminiferal assemblages, stable isotopes, and sedimentological parameters in order to reconstruct the palaeoceanographic and palaeoenvironmental evolution of the eastern Fram Strait in the last 14,000 years.
The sediment record has been divided into five time-intervals reflecting different environmental and climatic conditions (Fig. 10):

1. The end of Bølling and Allerød interstadial (14,120–12,850 cal yrs BP) was relatively cold with unstable climatic conditions, influence of polar surface water, and dominance of the polar species *Neogloboquadrina pachyderma* (s). However, beneath the cold polar surface waters, chilled Atlantic-derived intermediate water was flowing at the bottom of the Vestnesa Ridge, as suggested by the high abundance of the Atlantic Water species *Cassidulina neoteretis*.

2. The Younger Dryas (12,850–11,650 cal yrs BP) was characterised in its early part by the deposition of a sandy layer associated with a meltwater event and increase in opportunistic benthic foraminiferal species (*Stainforthia loeblichi* and *Nonionella iridea*). After the meltwater event the surface water conditions were still cold with a strong influence of polar surface water, but with open water conditions that allowed seasonal, pulsed primary production. The influx of Atlantic Water at the bottom was still present, but with lower temperature and lower current velocity.

3. The early Holocene (11,650–8800 cal yrs BP) was characterised by unstable and quite cold conditions at the beginning. The shift to interglacial conditions with cold intermediate water and deep-sea convection occurred at about 11,000 cal yrs BP in the benthic environment, while in the planktonic environment the shift to warm conditions occurred later around 10,300 cal yrs BP. Climate optimum conditions were finally established around 9800 cal yrs BP.

4. The middle Holocene (8800–4100 cal yrs BP) was warmer than today with strong inflow of Atlantic water, which persisted until about 7000 cal yrs BP. The 8200 cal yrs BP climate anomaly was inconspicuous in our record. A more pronounced cool event with abrupt decrease in *Turborotalita quinqueloba* occurred between 7000 and 6200
cal yrs BP. After 6200 cal yrs BP relatively warm conditions were established again, but generally cooler than before. The persistence of warmer conditions until the end of the period suggests a delayed influence of polar water at the core site in comparison to other records in the Nordic Seas.

5. The late Holocene (4100–150 cal yrs BP) was characterised by cold climatic conditions with *N. pachyderma* (s) as the dominant species again. The polar benthic species *Cribrorhphidium excavatum* appeared, indicating colder and more variable conditions. Warmer intermediate waters conditions occurred from about 2000 cal yrs BP, with periodic increases of Atlantic Water inflow.

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**Figure captions**

Fig. 1. Bathymetric map showing major surface currents in the Nordic Seas and Barents Sea, and the core location (black circle). Black dashed line: modern position of the Polar Front. Black line: modern position of the Arctic Front. Abbreviations: NAC: North Atlantic Current; WSC: West Spitsbergen Current; ESC: East Spitsbergen Current; EGC: East Greenland Current; DS: Denmark Strait. Figure is modified after Consolaro et al. (2015).
Fig. 2. A) Bathymetric map of the W-Svalbard margin and eastern Fram Strait. Abbreviations: MFZ: Molloy Fracture Zone; MD: Molloy Deep; MR: Molloy Ridge; SFZ: Spitsbergen Fracture Zone; YP: Yermak Plateau. B) Overview swath bathymetry map of the Vestnesa Ridge. Black circles indicate the location of the studied core JM10-330GC and CTD. White circles indicate location of other cores discussed in the paper: MSM5/5-712-2 (Werner et al., 2013; Aagaard-Sørensen et al., 2014a,b), MSM5/5-723-2 (Werner et al., 2016); core 1 (JM03-373PC2: Rasmussen et al., 2007; Jessen et al., 2010), and core 2 (MD99-2304: Ebbsen et al., 2007). Figures are modified after Consolaro et al. (2015).

Fig. 3. Conductivity-temperature-depth (CTD) data of temperature and salinity from the core site taken on June 27, 2010. Abbreviations: AW: Atlantic Water; GSIW: Greenland Sea Intermediate Water.

Fig. 4. Sedimentological and geochemical data plotted versus cal kyrs BP. (a) Calculated sediment accumulation rate. (b) Grain size distribution in %. (c) Number of ice-rafted debris (IRD) >1 mm (black) and concentration of IRD >100 µm per gram dry weight sediment (red line). (d) Concentration of diatoms (magenta), siliceous remains (radiolarian and sponge spicules; blue), and pyritized burrows (brown line) >100 µm per gram dry weight sediment. (e) CaCO3 wt% (grey) and TOC wt% (black line). (f) Stratigraphy obtained for JM10-330GC with chronological subdivisions (black lines). White diamonds on the y axis indicate radiocarbon dated levels. Abbreviations: B-A: Bølling-Allerød; YD: Younger Dryas; sl: sandy layer; E-H: early Holocene.

Fig. 5. Isotopic records plotted versus cal kyrs BP. Left: δ¹⁸O_{IVC} (ice volume corrected) record of Cassidulina neoteretis (red line), Cibicides wuellerstorfi (green line) and N. pachyderma (s) (blue line). Right: δ¹³C record of C. neoteretis (red line), C. wuellerstorfi (green line) and N. pachyderma (s) (blue line). Dotted vertical lines mark average stable isotope values of species. White diamonds on the y axis indicate radiocarbon dated levels. Shaded horizontal bars indicate carbon isotope excursions CIE I and CIE II (Consolaro et al., 2015). Abbreviations: IVC: ice volume corrected; B-A: Bølling-Allerød; YD: Younger Dryas; sl: sandy layer.
Fig. 6. Relative abundance (%; grey) and flux (black line) of planktonic foraminifera plotted versus cal kyrs BP; (a) Neogloboquadrina pachyderma (s); (b) Turborotalita quinqueloba; (c) N. pachyderma (d); (d) Globigerinita uvula; (e) Globigerinita glutinata; (f) Globigerina bulloides; (g) Concentration of planktonic foraminifera per gram dry weight sediment (grey) and planktonic foraminifera flux (black line); (h) Presence of pteropods. White diamonds on the y axis indicate radiocarbon dated levels.

Fig. 7. Relative abundance (%; grey) and flux (black line) of benthic foraminifera plotted versus cal kyrs BP. (a) Cassidulina neoteretis; (b) C. reniforme; (c) Melonis barleeanus; (d) Lobatula lobatula; (e) Cibicides wuellerstorfi; (f) Epistominella arctica; (g) Cribroelphidium excavatum; (h) Oridorsalis umbonatus; (i) Ioanella tumidula; (j) Stainforthia loeblichi; (k) Nonionella iridea; (l) Nonionellina labradorica; (m) Miliolids; (n) Agglutinated; (o) Concentration of benthic foraminifera per gram dry weight sediment (grey) and benthic foraminifera flux (black line). White diamonds on the y axis indicate radiocarbon dated levels.

Fig. 8. Comparison of planktonic foraminiferal records of JM10-330GC with other records from the western Svalbard margin. (a) Core JM03-373PC (b) core MD99-2304 (c) core MSM5/5-712-2 (d) core JM10-330GC and (e) core MSM5/5-712-2.

Fig. 9. Comparison of C. wuellerstorfi and N. pachyderma (s) isotopic records of JM10-330GC with other records from the western Svalbard margin. (a–c) $\delta^{18}$O$_{IVC}$ records of C. wuellerstorfi; (d–h) $\delta^{18}$O$_{IVC}$ records of N. pachyderma (s); (i–l) $\delta^{13}$C records of N. pachyderma (s); (m–o) $\delta^{13}$C records of C. wuellerstorfi. Data for a, m (Werner et al., 2013); b, g, k, n (this study); c, h, i, o (Werner et al., 2016); d (Rasmussen et al., 2007); e, j (Ebbesen et al., 2007); f, j (Werner et al., 2013; Aagaard-Sørensen et al., 2014a).

Fig. 10. Schematic overview with the main parameters of core JM10-330GC plotted versus cal kyrs BP, and their interpretation. (a) $\delta^{18}$O$_{IVC}$ record of Cibicides wuellerstorfi (green) and
N. pachyderma (s) (blue); Relative abundance (%) of: (b) T. quinqueloba (magenta) and N. pachyderma (s) (blue, inverse scale); (c) C. neoteretis (red) and C. reniforme (light blue, inverse scale); (d) C. wuellerstorfi (dark green) and C. excavatum (deep purple, inverse scale); (e) δ¹³C record of Cibicides wuellerstorfi (green) and N. pachyderma (s) (blue); (f) Concentration of planktonic (green) and benthic (orange) foraminifera per gram dry weight sediment; (g) Interpretation of the environmental conditions at the surface (S) and at the bottom (B). Blue star: omitted data because the isotopic signal in the B-A is disturbed by CIE. Green star: omitted data because the number of planktonic foraminifera in the sandy layer is much higher. Abbreviations: AW: Atlantic Water; Convection: deep-sea convection.

Fig. 11. Scanning electron microscope images of important benthic foraminiferal species from core JM10-330GC (scale bar = 100 µm). 1, 2. Adercotryma glomerata 0–1 cm. 3, 4. Ammoglobigerina globigeriniformis 0–1 cm. 5, 6. Trochammina nana 300–3001 cm. 7, 8. Veleroninoides wiesneri 5–6 cm. 9. Lagenammina micacea 0–1 cm. 10. Textularia earlandi 0–1 cm. 11, 12. Bolivinella pseudopunctata 300–301 cm. 13. Stainforthia feylingi 0–1 cm. 14. Stainforthia loeblichii 355–356 cm. 15, 16. Cassidulina neoteretis 0–1 cm. 17, 18. Cassidulina reniforme 0–1 cm. 19. Epistominella arctica 0–1 cm. 20. Eilohedra vitrea 300–3001 cm.

Fig. 12. Scanning electron microscope images of important benthic foraminiferal species from core JM10-330GC (scale bar = 100 µm). 21, 22. Eilohedra nipponica 0–1 cm. 23, 24. Lobatula lobatula 0–1 cm. 25, 26. Cibicides wuellerstorfi 0–1 cm. 27. Ioanella tumidula 0–1 cm. 28. Ioanella tumidula 15–16 cm. 29. Oridorsalis umbonatus 0–1 cm. 30. Pullenia osloensis 355–356 cm. 31, 32. Pullenia bulloides 0–1 cm. 33, 34. Melonis barleeanus 0–1 cm. 35, 36. Cribroelphidium excavatum 0–1 cm. 37. Nonionellina labradorica 5–6 cm. 38. 39. Nonionella stella 355–356 cm. 40, 41. Nonionella iridea 5–6 cm. Scale bar is 100 µm.

**Taxonomic Appendix**

**Planktonic foraminifera**

*Globigerina bulloides* d’Orbigny, 1826

*Globigerinita glutinata* (Egger, 1893) = *Globigerina glutinata* Egger, 1895
Globigerinita uvula (Ehrenberg, 1861) = Globigerina bradyi Wiesner, 1931

Neogloboquadrina pachyderma (Ehrenberg, 1861) = Planulina pachyderma Ehrenberg, 1854

Turborotalita quinqueloba (Natland, 1938) = Globigerina quinqueloba Natland, 1938

Benthic foraminifera

Adercotryma glomerata (Brady, 1878) = Lituola glomerata Brady, 1878

Ammoglobigerina globigeriniformis (Parker and Jones, 1865) = Lituola nautiloides var.
globigeriniformis Parker and Jones, 1865

Astronomion gallowayi Loeblich and Tappan, 1953

Bolivinella pseudopunctata (Höglund, 1947) = Bolivina pseudopunctata Höglund, 1947

Buccella frigida (Cushman, 1921) = Pulvinulina frigida Cushman, 1921

Cassidulina laevigata d’Orbigny, 1826

Cassidulina neoteretis Seidenkrantz, 1995

Cassidulina reniforme Nørvang, 1945

Cibicides sp. de Montfort, 1808

Cibicides wuellerstorfi (Schwager, 1866) = Anomalina wuellerstorfi Schwager, 1866

Cribroelphidium excavatum (Terquem, 1875) = Polystomella excavata Terquem, 1875

Cribrostromoides subglobosus (Cushman, 1910) = Haplophragmoides subglobosum Cushman, 1910

Cornuspira involvens (Reuss, 1850) = Operculina involvens Reuss, 1850

Dentalina sp. d’Orbigny, 1826

Discorbis sp. Lamarck, 1804

Earlandammina inconspicua Earland, 1934

Eilohedra nipponica Kuwano, 1962

Eilohedra vitrea (Parker, 1953) = Epistominella vitrea Parker, 1953

Elphidium sp. de Montfort, 1808
Epistominella arctica Green, 1959 (with this name we include all the intergradational series of Stetsonia horvathi Green, 1959 and E. arctica, according to Scott and Vilks, 1991).

Epistominella exigua (Brady, 1884) = Pulvinalina exigua Brady, 1884

Evolvocassidulina bradyi (Norman, 1881) = Cassidulina bradyi Norman, 1881

Fissurina sp. Reuss, 1850

Fursenkoina sp. Loeblich and Tappan, 1961

Gavelinopsis praegeri (Heron-Allen and Earland, 1913) = Discorbina praegeri Heron-Allen and Earland, 1913

Glandulina d’Orbigny, 1839

Globobulimina Cushman, 1927

Guttulina sp. d’Orbigny, 1839

Gyroidina lamarckiana (d’Orbigny, 1839) = Rotalina lamarckiana d’Orbigny, 1839

Haynesina germanica (Ehrenberg, 1840) = Nonionina germanica Ehrenberg, 1840

Haynesina orbiculare (Brady, 1881) = Nonion orbiculare Brady, 1881

Ioanella tumidula (Brady, 1884) = Truncatulina tumidula Brady, 1884

Islandiella helenae Feyling-Hanssen and Buzas, 1976

Islandiella norcrossi (Cushman, 1933) = Cassidulina norcrossi Cushman, 1933

Lagena sp. Walker and Boys, 1798

Lagena striata (d’Orbigny, 1839) = Oolina striata d’Orbigny, 1839

Lagenammina micacea (Cushman, 1918) = Proteonina micacea Cushman, 1918

Lenticulina sp. Lamarck, 1804

Marginulina sp. d’Orbigny, 1826

Melonis barleeanus (Williamson, 1858) = Nionionina barleeana Williamson, 1858

Miliolinella sp. Wiesner, 1931

Nodosaria sp. Lamarck, 1816

Nonionella sp. Rhumbler, 1949
Nonionella auricula Heron-Allen and Earland, 1930
Nonionella iridea Heron-Allen and Earland, 1932
Nonionella stella Cushman and Moyer, 1930
Nonionellina labradorica (Dawson, 1860) = Nonionina labradorica Dawson, 1860
Nonionoides turgida (Williamson, 1858) = Rotalina turgida Williamson, 1858
Oolina sp. d’Orbigny, 1839
Oolina apiopleura (Loeblich and Tappan, 1953) = Lagena apiopleura Loeblich and Tappan, 1953
Oridorsalis umbonatus (Reuss, 1851) = Rotalina umbonatus Reuss, 1851
Patellina corrugata Williamson, 1858
Pullenia bulloides (d’Orbigny, 1846) = Nonionina bulloides d’Orbigny, 1846
Pullenia osloensis Feyling-Hanssen, 1954
Pyrgo Defrance, 1824
Quinqueloculina sp. d’Orbigny, 1826
Quinqueloculina arctica Cushman, 1933
Quinqueloculina seminula (Linnaeus, 1758) = Serpula seminulum Linnaeus, 1758
Reophax sp. de Montfort, 1808
Reophax scorpiurus de Montfort, 1808
Reussoolina laevis (Montagu, 1803) = Vermiculum laeve Montagu, 1803
Robertinoides charlottensis (Cushman, 1925) = Robertina charlottensis Cushman, 1925
Stainforthia concava (Höglund, 1947) = Virgulina concava Höglund, 1947
Stainforthia feylingi Knudsen and Sedenkrantz, 1994
Stainforthia fusiformis (Williamson, 1848) = Bulimina fusiformis Williamson, 1848
Stainforthia loeblichii Feyling-Hanssen, 1954
Textularia earlandi Parker, 1952
Trifarina angulosa (Williamson, 1858) = Uvigerina angulosa Williamson, 1858
| 1074 | *Trifarina carinata* (Cushman, 1923) = *Angulogerina carinata* Cushman, 1927 |
| 1075 | *Trochammina* sp. Parker and Jones, 1859 |
| 1076 | *Trochammina nana* (Brady, 1881) = *Haplophragmium nanum* Brady, 1881 |
| 1077 | *Trochammina pseudoinflata* Scott and Vilks, 1991 |
| 1078 | *Veleronoides wiesneri* (Parr, 1950) = *Labrospira wiesneri* Parr, 1950 |
Depth (m) - Temperature (°C)

Depth (m) - Salinity (PSU)

- Mixed layer
- AW
- GSIW

June 27, 2010
Sedimentation rate (cm/kyr)

Grain size (%)

#IRD >1mm

# diatoms >100μm/g

CaCO₃ wt%

TOC wt%

Diatom rich mud

Pyritized burrows >100μm/g

CaCO₃ wt%

TOC wt%

Diatom rich mud

Pyritized burrows >100μm/g

CaCO₃ wt%

TOC wt%

Diatom rich mud

Pyritized burrows >100μm/g

CaCO₃ wt%

TOC wt%

Diatom rich mud

Pyritized burrows >100μm/g

CaCO₃ wt%

TOC wt%

Diatom rich mud

Pyritized burrows >100μm/g

CaCO₃ wt%

TOC wt%

Diatom rich mud

Pyritized burrows >100μm/g

CaCO₃ wt%

TOC wt%

Diatom rich mud

Pyritized burrows >100μm/g

CaCO₃ wt%

TOC wt%
Diatom rich mud

Early Hol.

Mid- Holocene

Late Holoc.

N. pachyderma T. quinqueloba N. pachyderma G. uvula G. glutinata G. bulloides

PF flux

#PF/g

Pteropods

Diatom rich mud

Late Holocene

Mid- Holocene

Early Hol.

Cal kyrs BP

(a) (b) (c) (d) (e) (f) (g) (h)

0 4 8 12 0 4 8 0 0.5 1 0 0.5 0 0.3 0 10 20

0 40 80 0 40 80 0 4 8 0 5 0 3 0 300 600

T. quinqueloba

N. pachyderma

G. bulloides

G. glutinata

G. uvula

N. pachyderma (s) %

N. pachyderma (d) %

G. uvula %

G. glutinata %

G. bulloides %

PF flux (#/cm²/yr)

Flux #/cm²/yr
Conditions at the surface (S) and at the bottom (B)

- S: > cold polar, productive water
- B: cooling, ice-margin conditions
- S: strong, stable AW inflow, high productivity
- short cold interval
- B: > productivity, strong convection
- S: > AW inflow
- B: cold interm. water, convection
- unstable transition
- S: polar water
- B: weaker AW
- S: polar water
- B: strong AW

(a) $\delta^{18}$O$_{IVC}$ (% VPDB)

(b) $T. quinquelaoboa$ %

(c) $C. reniforme$ %

(d) $C. excavatum$ %

(e) $\delta^{13}$C (% VPDB)

(f) $#PF/g$

(g) $#BF/g$