Late Weichselian and Holocene paleoceanography of Storfjordrenna, southern Svalbard

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Abstract

Multiproxy analyses (incl. benthic and planktonic foraminifera, $\delta^{18}O$ and $\delta^{13}C$ records, grain-size distribution, ice-rafted debris, XRF geochemistry and magnetic susceptibility) were performed on a ^{14}C dated marine sediment core from Storfjordrenna, off southern Svalbard. The sediments in the core cover the termination of Bølling-Allerød, the Younger Dryas and the Holocene, and they reflect general changes in the oceanography/climate of the European Arctic after the last glaciation. Grounded ice of the last Svalbard- Barents Sea Ice Sheet retreated from the coring site c. 13,950 cal yr BP. During the transition from the sub-glacial to glaciomarine setting, Arctic Waters dominated the hydrography in Storfjordrenna. However, the waters were not uniformly cold and experienced several warmer spells. A progressive warming and marked change in the nature of hydrology occurred during the early Holocene. Relatively warm and saline Atlantic Water started to dominate the hydrography from approx. 9600 cal yr BP. Even though the climate in eastern Svalbard was milder at that time than at present (smaller glaciers), there were two slight coolings observed in the periods of 9000 - 8000 cal yr BP and 6000 - 5500 cal yr BP. A change of the Storfjordrenna oceanography occurred at the beginning of late Holocene (i.e. 3600 cal yr BP) synchronously with glacier growth on land and enhanced bottom current velocities. Although cooling was observed in the surface water, Atlantic Water remained present in the deeper part of water column of Storfjordrenna.

1 Introduction

The northward flowing North Atlantic Current (NAC) is the most important source of heat and salt in the Arctic Ocean (Gammelsrod and Rudels, 1983; Aagaard et al., 1987; Schauer et al., 2004; Fig. 1b). The main stream of Atlantic Water (AW) flowing north to Fram Strait as the West Spitsbergen Current (WSC) causes the dramatic reduction of sea ice extent and thickness through the warming of the intermediate water layer in this region of the Arctic Ocean (Quadfasel et al., 1991; Serreze et al., 2003). Paleoceanographic (e.g., Spielhagen et al., 2011; Dylmer et al., 2013) and instrumental (Walczowski and Piechura 2006, 2007; Walczowski et al., 2012) investigations provide evidence of a recent intensification of the flow of AW in the Nordic Seas and the Fram Strait.

The Svalbard archipelago is influenced by two water masses: AW flowing northward from the North Atlantic and Arctic Water (ArW) flowing southwest from the northern Barents Sea (Fig. 1b). An oceanic front arising at the contact of different bodies of water is an excellent area to research contemporary and past environmental changes. Intensification of AW flow and associated climate warming cause decreased sea-ice cover in the Svalbard fjords during winter (Berge et al., 2006), increased sediment accumulation rate (Zajączkowski et al., 2004; Szczuciński et al., 2009) and influences pelage-benthic carbon cycling (Zajączkowski et al., 2010).

Paleoceanographic records indicate that AW was present along the western margin of Svalbard, at least, during the last 12,000 years (e.g. Ślubowska et al., 2007; Werner et al., 2011; Rasmussen et al., 2013); occasionally reaching the Hinlopen Trough and Kvitøya Trough, thus transporting warmer and more saline water to the eastern part of Svalbard from the north (Ślubowska-Woldengen et al., 2007; Ślubowska et al., 2008; Kubischta et al., 2010; Klitgaard Kristensen et al., 2013). Periods of enhanced inflow of AW during the Holocene led to the expansion of marine species being absent or only rarely occurring at present. This includes the mollusc *Mytilus edulis* whose fossil remains are widely distributed in raised beach deposits on the western and northern coasts of Svalbard (e.g. Feyling-Hanssen and Jørstad, 1950; Hjort et al., 1992). *Mytilus edulis* spawn at temperatures above 8 to 10 °C (Thorarinsdóttir and Gunnarson, 2003) and thus is considered to indicate higher surface-water temperature related to stronger AW inflow during the early Holocene (11,000 – 6800 cal yr BP) (Feyling-Hanssen, 1955; Salvigsen et al., 1992; Hansen et al., 2011). Although the progressive development of *Mytilus edulis* is well documented by the periods of warming and inflow of AW to Hinlopen Trough, the presence of this species in Storfjorden (W Edgeøya;

Fig. 1) is unclear. Hansen et al. (2011) suggested that a small branch of warm AW could have reached eastern Spitsbergen from the south at that time.

In the 1980s and 1990s, Storfjorden was regarded to be exclusively influenced by the East Spitsbergen Current (ESC), carrying the cold and less saline ArW from the Barents Sea (Quadfasel et al., 1988; Piechura et al., 1996). More recent studies suggested that the hydrography in Storfjorden is affected by the production of brine-enriched shelf waters (e.g., Haarpaintner et al., 2001; Rasmussen and Thomsen, 2009), the creation of a coastal polynya (e.g., Skogseth et al., 2005; Geyer et al., 2010) or the overflow of dense waters to the continental shelf (e.g., Fer et al., 2003). However, hydrological data obtained from conductivity-temperature sensors attached to a *Delphinapterus leucas* showed a substantial and topographically steered inflow of AW to Storfjorden through the Storfjordrenna (Lydersen et al., 2002). Recently, Akimova et al. (2011) reviewed typical water masses for Storfjorden, where the AW was located between 50 and 70 meters.

Storfjordrenna is a sensitive boundary area (Fig. 1) where two contrasting water masses form an oceanic polar front, separating colder, less saline and isotopically lighter ArW from warmer, high saline and $\delta^{18}O$ heavier AW. An abrupt cooling (e.g. Younger Dryas, Little Ice Age) and warming (e.g. early Holocene warming) of the European Arctic might be linked to relatively small displacements of this front (Sarnthein et al., 2003; Hald et al., 2004; Rasmussen et al., 2014).

Two sediment cores taken at the mouth of Storfjordrenna, reveal a continuous inflow of AW to the south western Svalbard shelf since the deglaciation of Svalbard-Barents Ice Sheet (Rasmussen et al,. 2007), while inner Storfjorden basins undergo a shift from being occupied by continental ice to ice proximal condition (Rasmussen and Thomsen, in press). Nevertheless a limited amount of paleoceanographical data is available from this region, thus the reconstruction of the Svalbard-Barents Ice Sheet retreat and further development of Storfjordrenna oceanography is often speculative.

In this paper we present results from multi-proxy analyses of a sediment core retrieved 100 km east of the mouth of Storfjordrenna (Fig. 1a). We provide a new age for the retreat of the last Svalbard-Barents Sea Ice Sheet from Storfjordrenna and discuss the interaction of oceanography and deglaciation, as well as the postglacial history of Atlantic Water inflow onto the shelf off southern Svalbard. Since the studied sediment core was retrieved from an oceanographic frontal zone, sensitive to larger-scale changes, we believe that the presented data show the general climatic/oceanographic trends in the eastern Arctic.

2 Study area

Storfjorden is an approx. 190 km long and up to 190 m deep glacial trough located between the landmasses of Spitsbergen to the west, Edgeøya and Barentsøya to the east, and the shallow Storfjordenbanken to the south-east (Fig. 1a). It is not a fjord sensu stricto, as the sounds of Heleysundet and Freemansundet to the north and northeast, respectively, connect the head of Storfjorden to the north western Barents Sea. A sill of 120 m depth crosses the mouth of Storfjorden. The 254 km long Storfjordrenna, a continuation of the trough that extends towards the shelf break, is located beyond this sill. Bottom depth along the trough axis varies between 150 m and 420 m (Pedrosa et al., 2011).

2.1 Water masses

The water column of Storfjorden and Storfjordrenna is composed of two main water masses transported with currents from east and south and mixed waters which are formed locally (Table 1. after Skogseth et al., 2005). Warm and saline Atlantic Water (AW) enters Storfjordrenna in a cyclonic manner (Schauer, 1995; Fer et al., 2003), flowing into the trough parallel to its southern margin and flowing towards the trough mouth along its northern slope. The AW occurs between 50 and 70 m in Storfjorden and extends to a depth of 200 m in Storfjordrenna (Akimova et al., 2011). The origin of AW entering Storfjordrenna is an eastward branch of the North Atlantic Current (NAC) following the topography of the Barents Sea Shelf Break. However, approx. 50% of AW flowing northward also penetrate into Bjørnøyrenna (Smedsrud et al., 2013; for location see Fig. 1). The AW in Storfjordrenna is cooler and fresher than in Bjørnøyrenna as an effect of distance and mixing processes (O'Dwyer et al., 2001). AW may occasionally propagate even further east of Svalbard, where it fills the depressions below 180 m (Schauer, 1995). Relatively cold Arctic Water (ArW) is transported to Storfjorden and Storfjordrenna by the East Spitsbergen Current (ESC). The ESC enters the fjord through the tidally influenced sounds of Heleysundet and Freemansundet in the north and northeast (Norges Sjøkartverk, 1988), as well as from the southeast with a coastal current flowing around Edgøya (Loeng, 1991). AW and ArW mix to form Transformed Atlantic Water (TAW), which dominates on the shelf off west Spitsbergen (Svendsen et al., 2002; Table 1). Dense, brine-enriched Shelf Water (BSW) in Storfjorden is produced through high polynya activity and results from intense formation of sea ice (Haarpaintner et al., 2001; Skogseth et al., 2004, 2005). The BSW fills the fjord to the top of the sill (120 m) and initiates a gravity driven overflow (Quadfasel et al., 1988; Schauer, 1995; Schauer and Fahrbach, 1999; Fer et al., 2003, 2004; Skogseth et al., 2005). BSW is characterized by salinity greater than 34.8 and temperature at or slightly above the freezing point (Table 1). Surface Water (SW) in the upper 50 m is cold and fresh during the autumn and warm and fresh due to ice melting during the summer. In winter, the water column in Storfjorden is homogenized due to wind and tidal mixing and is considered to be close to the freezing point (Skogseth et al., 2005).

3 Material and methods

Multi-proxy analyses of the gravity core JM09-020-GC provided the basis for this study. The core was retrieved with R/V Jan Mayen (University of Tromsø – The Arctic University of Norway, UiT) in November 2009 from the Storfjordrenna (76°31489' N, 19°69957' E), from a bottom depth of 253 m (Fig. 1a). The coring site was located in an area above the continuous presence of BSW and was selected after an echo-acoustic investigation in order to identify the greatest possible area of flat bottom with minimum disturbance of sediments. Conductivity-temperature-depth (CTD) measurements were performed prior to coring (Fig. 2a) and in summer 2013 (Fig. 2b).

Prior to sediment core opening, the magnetic susceptibility (MS) was measured using a loop sensor installed on a GEOTEK Multi Sensor Core Logger at the Department of Geology, UiT. Core sections were stored in the laboratory for one day before measurements thereby allowing the sediments to adjust to room temperature and to avoid measurement errors related to temperature changes (Weber et al., 1997). X-radiographs and digital images were taken from half of the core to define sedimentary and biogenic structures. Sediment colour was defined according to the Munsell Soil Color Charts (Munsell Products, 2009). Qualitative element-geochemical measurements were performed with an Avaatech X-ray fluorescence (XRF) core scanner using the following settings: 10 kV; 1000 µA; 10 sec. measuring time; no filter. Both core halves were subsequently cut into 1-cm slices and transported to the Institute of Oceanology, Polish Academy of Sciences in Sopot for further analyses.

Sediment samples for foraminiferal analyses were freeze-dried, weighed, and wet sieved using sieves with mesh-sizes of 500 µm and 100 µm. Residues were dried, weighted again and then split on a dry micro-splitter. Where possible, at least 300 specimens of foraminifera were counted in every 5 cm of sediment. Species identification under a binocular microscope (Nikon SMZ1500) was supported using classification of Loeblich and Tappan (1987), with few exceptions. Percentages of the 8 indicator species were applied. The number of species per sample and Shannon-Wiener Index were calculated in the program Primer 6. The benthic foraminiferal abundance and ice-rafted debris (IRD; grains >500 μm) were counted under a stereo-microscope and expressed as flux values (no. of specimens/grains cm^{-2} ka^{-1}) using the bulk sediment density and sediment accumulation rate.

Stable oxygen and carbon isotope compositions of tests of the infaunal foraminifer species *Elphidium excavatum* f. *clavata* were determined at the Department of Geological Sciences, University of Florida (Florida, USA). All values are calibrated to the PeeDee Belemnite (PDB) scale and corrected for ice volume changes. In our study we discuss the $\delta^{18}O$ and $\delta^{13}C$ record as a relative measure for changes in the water mass characteristics (temperature-salinity) and/or the supply of meltwater/freshwater to the area. Therefore, we haven't corrected the values for vital effect.

Grain size (<2 mm) analyses were performed every 1 cm using a Malvern Mastersizer 2000 laser particle analyser and presented as volume percent. To examine relative variability in the near-bottom currents the mean grain size distribution of the <63 μm fraction was calculated, to avoid effect of ice-rafted coarse fraction. Mean grain size was calculated in the program GRADISTAT 8.0 by the geometric method of moments (Blott and Pye, 2001).

3.1 Age control

The chronology for this study is based on high-precision AMS ^{14}C measurements of fragments from nine calcareous bivalve shells. Measurements were performed in the Poznań Radiocarbon Laboratory, which is equipped with the 1.5 SDH-Pelletron Model "Compact Carbon AMS" (Czernik and Goslar, 2001; Goslar et al., 2004). The surface layer of shells was scraped off to avoid contamination with younger carbonate encrustation. The AMS ^{14}C dates were converted into calibrated ages using the calibration program CALIB 6.1 (Stuiver and Reimer, 1993; Stuiver et al., 2005) and the Marine13 calibration curve (Reimer et al., 2013). The difference ΔR in reservoir age correction of the model ocean and region of Svalbard was reported by Mangerud et al. (2006) to be 105±24 or 111±35; we used the first value; calibrated ages are presented in Table 2. It should be noted that the reservoir age is based on few data points from western Spitsbergen, and the age may be different for the eastern coast. However, no data are available from the latter region.

4 Results

4.1 Modern hydrology

In November 2009 the surface water at the coring site (upper ~27 m) had already cooled down (1.24 °C; Fig. 2a). However, its salinity was still low (34.24 °C). Transformed AW was observed in the layer between 60 and 160 m. The lowermost part of water column shows gradual cooling reaching a minimum temperature of 0.76 °C near the bottom. The lack of BSW at the bottom indicates gradual water mixing during summer and fall. In August 2013, the surface waters had slightly lower salinity, but the temperature was ~5 °C higher than in November 2009 (Fig. 2b). TAW occupied the same depths as in 2009. However, an almost 50 m thick layer of BSW was present close to the seafloor.

4.2 Age model

The ^{14}C ages and calibrated ages are reported in Table 2. The calibration gives an age distribution, not a single value, so the 2-sigma range presented and Fig. 3 shows age probability distribution curves. Ages of samples generally increase with sediment depth except in the case of one sample: St 20A 39, which provided an older age than the sample below. That shell was most likely re-deposited and was thus not used for the age model. However, because all the samples used for dating were shell fragments, it must be taken into account that it is possible that more samples could be subjected to re-deposition, but on the basis of the available data this is not possible to confirm. The age model is based on assuming linear sediment accumulation rates between data points. The highest probability peaks from calibrated age ranges were used as input values for the model. For the lowermost and uppermost parts of the core, we adopted sediment accumulation rates for the neighbouring parts. It is common to observe the loss of the sediment surface layer during coring with heavy gravity cores. In the case of core JM09-020-GC it is likely that at least the top 40 cm of sediments were lost during coring. This conclusion is supported by analysis of a box corer collected prior to coring (Łącka et al., in prep.). The extrapolated age model for the sediment surface is, therefore, 1200 cal yr BP.

4.3 Sedimentological and geochemical parameters

The core JM09-020-GC is 426 cm long and consists of four lithological units L1 (bottom of the core to 370 cm; >13,450 cal yr BP), L2 (370 cm to 272 cm; ~13,450 cal yr BP to ~11,500 cal yr BP), L3 (272 cm to 113 cm; ~11,500 cal yr BP to ~3600 cal yr BP) and L4 (113 cm to core top; ~3600 cal yr BP to ~ 1200 cal yr BP). The lithological log was created based on the X-radiographs, grain-size analysis data and foraminiferal flux (Fig. 4). Grains >2 mm are referred to as "clasts" and are marked in the lithological logs as individual features.

Unit L1 consists of compacted massive dark grey (5Y 4/1) sandy mud with various amounts of clasts. Bioturbation and foraminifera were generally absent. However, one shell fragment was found at approx. 395 cm.

Unit L2 contains massive dark grey (5Y 4/1) sandy mud with some coarser material and generally lower amounts of clasts than unit L1. The mean grain size (<63 μm) ranged from 7-10 μm. The highest IRD flux and Fe/Ca ratio for the entire core occur in this unit. The mass accumulation rate (MAR) is 0.043 g cm^{-2} yr^{-1}. The first signs of bioturbation occur in this unit and the flux of foraminifera increases rapidly up to ~5700 individuals cm^{-2} ka^{-1} (Fig. 4).

The unit L3 is composed of massive dark olive grey mud (5Y 3/2) and is characterized by decreasing MAR values (0.019 g cm^{-2} yr^{-1} to 0.002 g cm^{-2} yr^{-1}), moderate sand content and clearly increasing mean grain size (<63 μm). IRD flux is low and the Fe/Ca ratio decreases gradually until c. 9200 cal yr BP and then remains low (between 3 and 4; Fig. 4) Continuous bioturbation and variable foraminiferal fluxes, with maxima in the intervals 9000-8000 cal yr BP and 6000-5500 cal yr BP, are observed.

The uppermost unit L4 is mostly composed of the same material as the underlying unit- massive dark olive grey mud (5Y 3/2). However, the sand content is occasionally higher. MAR increases to 0.024 g cm^{-2} yr^{-1}. The mean grain size (<63 μm) through this interval is even higher than in L3 and reaches up to 15 μm and Fe/Ca ratio is increasing. The bioturbation continues, numerous shell fragments are presented and foraminifera flux reaches high values throughout the entire unit.

4.4 Foraminiferal fauna

A total of 54 calcareous and 6 agglutinated species were identified. The foraminiferal assemblages were dominated by calcareous fauna. Agglutinated species occurred only in 14 sediment samples, and their abundance did not exceeded 4%. The only exception is the sample dated to c. 11,350 cal yr BP (262.5 cm depth) with 25% of agglutinated foraminiferal fauna. However, in this sample the total foraminifera abundance was low (13 specimens g^{-1} sediment). In general, species richness, number of agglutinated foraminifera, as well as rare and fragile species, increase towards the top of the core. Benthic foraminiferal fauna is dominated by *Elphidium excavatum* f. *clavata*, *Cassidulina reniforme*, *Nonionellina labradorica*, *Melonis barleeanum*, *Islandiella* spp. (*Islandiella norcrossi*/*Islandiella helenae*) and *Cibicides lobatulus*. Percentages of *E. excavatum* f. *clavata* show an inverse relationship to *C. reniforme* with the almost constant dominance of the latter species in the periods: ~12,450 cal yr BP to ~12,000 cal yr BP and ~ 9600 cal yr BP to ~2800 cal yr BP (Fig. 5). Planktonic foraminifera are represented by three species, *Neogloboquadrina pachyderma* (sinistral), *Neogloboquadrina pachyderma* (dextral) and *Turborotalita quinqueloba*. However, the two later species are very rare. In general, the abundance of planktonic fauna is low in the older parts of the core and slightly increases approx. 10,000 cal yr BP reaching maximum values c. 2000 cal yr BP (Fig. 6).

Based on the most significant changes in the foraminiferal species abundances, species diversity and $\delta^{18}O$ and $\delta^{13}C$ in *E. excavatum* f. *clavata* tests the core was divided into the four foraminiferal zones F1-F4: ~13,450 cal yr BP to 11,500 cal yr BP (F1); 11,500 cal yr BP to 9200 cal yr BP (F2); 9200 cal yr BP to 3600 cal yr BP (F3); 3600 cal yr BP to 1200 cal yr BP (F4) (Fig. 5, Fig. 6). Zones correspond to lithological division: the age of unit F4 is the same as L4, units F3 and F2 correspond to L3 and unit F1 is linked to unit L2. In unit L4 foraminifera are rare to absent.

Zone F1 is dominated by the opportunistic *E. excavatum* f. *clavata* and *C. reniforme.* The latter one dominates over *E.excavatum* f.*clavata* between 12,250 cal yr BP and 11,950 cal yr BP. High percentages of *C. lobatulus* (up to 57%) and *Astrononion gallowayi* (up to 2.5%) occur occasionally. Planktonic foraminifera flux was low at the beginning of this section (mean value of 9 specimens cm^{-2} ka^{-1}) and completely disappeared for almost 1500 years from approx. 11,500 cal yr BP (Fig. 6). Species richness as well as Shannon-Wiener index show, compared to the upper part of the core, low biodiversity (mean values of 8 and 1.26, respectively). Furthermore, maxima of $\delta^{18}O$ and $\delta^{13}C$ occur in this interval.

In zone F2 the contribution of *E. excavatum* f. *clavata* and *C. reniforme* is slightly lower, and *N. labradorica* becomes the most abundant species (Fig. 5). There is also an increase in *Islandiella* spp. percentage. Planktonic foraminifera appeared again c. 10,000 cal yr BP. Biodiversity significantly increased and $\delta^{18}O$ reached its minimum value of 2.61 ‰ vs VPDB approx. 10,000 cal yr BP.

Zone F3 is characterized by the minimum mass accumulation rates of sediment and consequently, low temporal resolution. *C. reniforme* dominates over *E. excavatum* f. *clavata* throughout. *M. barleeanum* has its maximum abundance in this zone, and *N. labradorica* is abundant in the lower parts of this zone, decreasing at approx. 7000 cal yr BP. *Islandiella* spp. increases upcore. Planktonic foraminifera occur in the entire zone, and the fluxes are higher than those of previous units (Fig. 6). Biodiversity remains high in this zone, and $\delta^{18}O$ and $\delta^{13}C$ remain generally stable, however marked peaks occurred at approx. 6800 cal yr BP, 6500 cal yr BP and 5700 cal yr BP, respectively.

A consistently high foraminiferal flux of up to ~4900 no. of specimens cm^{-2} ka^{-1} characterises zone F4. The fluxes of *Islandiella* spp. and *Buccella* spp. increase significantly and from 2850 cal yr BP *Islandiella* spp. dominated the assemblage with *E.excavatum* f.*clavata*. Additionally, the fluxes of *C. lobatulus* and *A. gallowayi* increase. However, their abundances are lower than those of zone F2. A maximum abundance of planktonic foraminifera occurs in this unit. Foraminifera biodiversity continues to increase towards the core top (up to 2.33; Fig. 6). $\delta^{18}O$ and $\delta^{13}C$ increase slightly, however, with numerous fluctuations.

5 Discussion

Based on the most pronounced changes in sedimentological and foraminiferal data as well as comparison to previous studies from adjacent areas, we have distinguished 5 units in the studied core: a sub-glacial unit (>13,450 cal yr BP), glacier-proximal unit (13,450 cal yr BP to 11,500 cal yr BP), glaciomarine unit I (11,500 cal yr BP to 9200 cal yr BP), glaciomarine unit II (9200 cal yr BP to 3600 cal yr BP) and glaciomarine unit III (3600 to 1200 cal yr BP).

5.1 Sub-glacial unit (>13,450 cal yr BP)

The lowermost unit L1 (Fig. 4) was significantly coarser, compacted and devoid of foraminifera, which indicates its likely of sub-glacial origin. During the late Weichselian Glacial Maximum, Storfjorden and Storfjordrenna were covered by an ice stream draining the Svalbard-Barents Ice Sheet (SBIS; e.g., Ottesen et al.,2005). The SBIS deglaciation occurred as a response to sea-level rise and increased mean annual temperature (Siegert and Dowdeswell, 2002). Rasmussen et al. (2007) noted that the outer part of Storfjordrenna (389 m depth; Fig. 1a) was deglaciated before 19,700 cal yr BP. The bivalve shell fragment from

395.5 cm in our core suggests that the centre part of Storfjordrenna was ice-free before ~13,950 cal yr BP. This indicates that the ~100 km long retreat of the grounding line from the shelf break to the central part of Storfjordrenna occurred in approx. 5700 years. The deglaciation of the inner Storfjorden basin occurred c.11,700 cal yr BP (Rasmussen and Thomsen, 2014), while the coasts of east Storfjorden islands, Barentsøya and Edgeøya, which are located over 100 km north from the coring site, occurred some 500 years later, i.e., 11,200 cal yr BP (recalibrated after Landvik et al., 1995). Siegert and Dowdeswell (2002) noted that, during the Bølling-Allerød warming (c. 14,700-12,700 cal yr BP), some of the deeper bathymetric troughs (e.g., Bjørnøyrenna) had deglaciated first, forming large embayments of ice around them. Probably, Storfjordrenna was one of such embayments at that time. Our data is in agreement with ice stream retreat dynamics presented by Rüther et al. (2012) and refines the recent models of the Barents Sea deglaciation (e.g. Winsborrow et al., 2010; Hormes et al., 2013; Andreassen et al., 2014).

5.2 Glacier-proximal unit (13,450 cal yr BP to 11,500 cal yr BP)

The transition from a subglacial to the glaciomarine setting is observed as a distinct change in sediment colour, several peaks of IRD, decreased amount of clasts and the appearance of foraminifera. The sediment accumulation rate (0.043 g cm^{-2} yr^{-1}) was in the same order of magnitude as modern proximal and central parts of west Spitsbergen fjords (see Szczuciński et al., 2009 for review). Textural and compositional analyses of L2 recorded bimodal grain-size distribution and low abundance of microfossils, suggesting that deposition during the deglaciation occurred from suspension settling from sediment-laden plumes and ice rafting (Lucchi et al., 2013; Witus et al., 2014). This unit in our core is limited to ~60 cm and is characterized by a lack of bioturbation in its lower part.

The high flux of IRD supported by the high Fe/Ca ratio and depleted $\delta^{18}O$ values correlates well with the abundance of *C. lobatulus* and *A. gallowayi* (Fig. 4 and Fig. 5), two species connected with high energy environments (Østby and Nagy, 1982) indicating that the coring site was likely located proximal to one or several ice fronts during the time of deposition of this unit.

During an early phase of the deglaciation of Storfjorden, the East Spitsbergen Current was still not active, because the ice sheet grounded between Svalbardbanken and Storfjordbanken blocked the passage between eastern and western Svalbard (Rasmussen et al., 2007; Hormes et al., 2013). Thus, the first foraminiferal propagules (juvenile forms) were transported by sea currents (Alve and Goldstein, 2003) from the south and west and settled on the seafloor that was exposed after the retreat of grounded ice. The proximal glaciomarine environment affected foraminiferal assemblages and resulted in low species richness, biodiversity and low foraminiferal abundance. Consequently, foraminifera assemblages became dominated by fauna typical for the glacier proximal settings: *E. excavatum* f. *clavata*, *C. reniforme* and *Islandiella* spp. (e.g., Vilks, 1981; Osterman and Nelson, 1989; Polyak and Mikhailov, 1996; Hald and Korsun, 1997). Dominance of *E. excavatum* f. *clavata* confirms the proximity to the ice sheet, decreased salinity and high water turbidity (e.g., Steinsund, 1994; Korsun and Hald, 1998; Włodarska-Kowalczuk et al., 2013).

The upper part of unit L2 (c. 12,800-11,500 cal yr BP) spans the Younger Dryas (YD) stadial. Records of marine sediments from Nordic and Barents Sea (e.g., Rasmussen et al., 2007; Ślubowska-Woldengen et al., 2007, 2008; Zamelczyk et al., 2012; Groot et al., 2014), as well as δ^{18}O records from Greenland ice cores (e.g., Dansgaard et al., 1993; Grootes et al., 1993; Mayewski et al., 1993; Alley, 2000) show that the YD was characterised by a rapid and short-term temperature decrease. This event was likely driven by weakened North Atlantic Meridional Overturning Circulation, a result of the Lake Agassiz outburst (e.g., Gildor and Tziperman, 2001; Jennings et al., 2006; Murton et al., 2010; Cronin et al., 2012) or interaction between the sea ice and thermohaline water circulation (Broecker, 2006), which led to a reduction of AW transport to the north and a dominance of fresher Arctic Water. Our data shows that heavier δ^{18}O recorded e.g., 12,720 cal yr BP and 12,100 cal yr BP, correlate with reduced to absent IRD fluxes, while the peaks of lighter δ^{18}O, e.g., 12,450 cal yr BP, 12,150 cal yr BP and 11,780 cal yr BP, occurred synchronously with significant enhanced IRD fluxes (Fig. 7). Absence of IRD, occasionally for several decades, might reflect temporarily polar conditions (Dowdeswell et al., 1998; Gilbert, 2000) characterized by the formation of perennial pack ice in Storfjorden locking icebergs proximal to their calving fronts and preventing their movement over the coring site (Forwick and Vorren, 2009). On the other hand, warmer periods resulted in massive iceberg rafting and delivery of IRD to Storfjordrenna, thus reflecting more sub-polar conditions. Hydrological variability during Younger Dryas was previously noted in some circum-North Atlantic deep-water records (Bakke et al., 2009; Elmore and Wright, 2011 and references therein; Pearce et al., 2013). Moreover, oxygen stable isotopes record from an ice-core GISP2 shows some warmer spells during that time (Stuiver et al., 1995), which coincides with higher ice-rafting in Storfjordrenna (Fig. 7). Bakke et al. (2009) noted that the earlier part of YD was colder and more stable, whereas later part of this period was characterized by alternations between sea-ice cover and influx of warmer, salty North Atlantic waters. Our record shows that during the late YD $\delta^{18}O$ were slightly shifted towards lighter values. Temporal resolution of our record do not allow for more detailed comparison with available data, nevertheless it clearly indicate that the Younger Dryas was not uniformly cold and that at least some warmer spells occurred on eastern Svalbard.

We also conclude that the data on $\delta^{18}O$ presented in Fig. 7 reflects temperature variations at the coring site according to the isotopically lighter ArW paleotemperature model (Duplessy et al., 2005). Another explanation of the heavier $\delta^{18}O$ periods during the YD could be intermittent inflow of warmer AW. However, this is unlikely to cause the synchronous disappearance of IRD.

5.3 Glaciomarine unit I (early Holocene; 11,500 cal yr BP to 9200 cal yr BP)

During the early Holocene foraminiferal fauna, although low in abundance, was dominated by species related to the glaciomarine environment (*E. excavatum* and *C. reniforme*; Fig. 5). Increasing species richness and biodiversity of foraminifera point to amelioration of environmental conditions and a progressive increase in the distance to the glacier front (Korsun and Hald, 2000; Włodarska-Kowalczuk et al., 2013). Decrease of the Fe/Ca ratio is suggested to reflect increased marine productivity and reduced supply of terrigenous material (Croudace et al., 2006). The mean grain size (>63 μm; Fig. 4) indicates weaker bottom currents at the beginning of the early Holocene and stronger bottom currents at the end of this period, which might have been related to the ongoing isostatic uplift of the land masses of Svalbard, as well as sea level rise (e.g., Forman et al., 2004).

Significant fluctuations of the $\delta^{18}O$ and $\delta^{13}C$ and increasing abundance of *N. labradorica* and *Islandiella* spp. suggest that Storfjordrenna was under the influence of various water masses at this time (Fig. 6). Comparison of our $\delta^{18}O$ record with records from the Storfjorden shelf (400 m depth; Rasmussen et al., 2007; Fig. 1a) and the northern shelf of Svalbard (400 m depth; Ślubowska et al., 2005; Fig. 1b) show that all the records are shifted towards lighter values in the early Holocene (Fig. 8a) with the record from our core being the most depleted (from c. 13,000 cal yr BP). We suggest that the records located on the western and northern shelf of Svalbard directly mirror the effect of warmer Atlantic water inflow, while record from Storfjordrenna is under influence of isotopically lighter Arctic Water from the Barents Sea (Duplessy et al., 2005). The shift from the Arctic water domain to the Atlantic water domain during the end of the early Holocene is also visible on a scatter plot of $\delta^{13}C$

against $\delta^{18}O$ (Fig. 8b). The results grouped to the left indicate Arctic water domination, while the results grouped to the right shows Atlantic water domination.

According to Kaufman et al. (2004), the early Holocene is characterized by higher summer solar insolation at 60°N (10% higher than today), leading to a reduction in sea-ice cover (Sarnthein et al., 2003). As ice cover decreased, more solar energy was stored in summer and then re-radiated during the winter (e.g., Gildor and Tziperman, 2001). This process accelerated the ice sheet melting and finally, its retreat towards the fjord heads (Forwick & Vorren, 2009; Jessen et al., 2010; Baeten et al., 2010). Our data suggest that the iceberg calving to Storfjordrenna was significantly reduced or even disappeared approx. 10,800 cal yr BP. However, supply of turbid meltwater from land to the study area still resulted in relatively high sediment accumulation rate.

According to Risebrobakken et al., (2011) and Groot et al., (2014) the presence of Arctic water suppressed the warming signal in the western Barents Sea. This is in agreement with our data on planktonic foraminifera reappearing at the termination of the early Holocene (c. 9600 cal yr BP; Fig.6). During this period *N. pachyderma* (sin.) dominated, however some peaks of *N. pachyderma* (dex.) and *T. quinqueloba* were noted. The two latter species are regarded as subpolar species (Bé and Tolderlund, 1971), although *T. quinqueloba* could be also related to oceanic frontal conditions separating Atlantic and Arctic water (Johannessen et al., 1994; Matthiessen et al., 2001). The peaks of *T. quinqueloba* around 9600 cal yr BP were noted previously in western Barents Sea margin (e.g. Hald et al., 2007; Risebrobakken et al., 2010).

Increasing foraminiferal biodiversity in Storfjordrenna (Fig. 6), as well as the occurrence of the thermophilous mollusc *Mytilus edulis* on western Edgeøya (Salvigsen et al., 1992) suggest that the inflow of AW crossed Storfjordrenna and continued northward to the inner fjord by 9600 cal yr BP.

5.4 Glaciomarine unit II (mid-Holocene; 9200 cal yr BP to 3600 cal yr BP)

The mid-Holocene was characterized by relative stable environmental conditions, low sediment accumulation rates (0.002 g $cm^{-2}yr^{-1}$) and slight delivery of IRD (Fig. 4), reflecting very limited ice rafting and reduced supply of fine-grained material to Storfjordrenna. Low sedimentation rates and the low Fe/Ca ratio reflect reduced glacial conditions on Svalbard during the mid-Holocene (Elverhøi et al., 1995; Svendsen and Mangerud, 1997). In contrast, Hald et al. (2004) noted that in the record from Van Mijenfjorden, an enhanced tidewater glaciation occurred during this period; it was thus argued that IRD is a more reliable indicator of glaciation than sedimentation rates. However, ice rafting in Storfjordrenna was generally low.

Shifts between the dominant species *C. reniforme* and *E. excavatum* f. *clavata* (Fig. 5) reflect environmental/hydrological changes (Hald and Korsun, 1997). The decrease of *E. excavatum* f. *clavata* (percentage and flux), which prefers colder bottom waters (Sejrup et al., 2004; Saher et al., 2009) and increase of *C. reniforme* points to the constant inflow of less modified AW and reduction in sedimentation (e.g., Schröder-Adams et al., 1990; Bergsten, 1994; Jennings and Helgadóttir, 1994; Hald and Steinsund, 1996; Hald and Korsun, 1997). Furthermore, the relative abundance of *M. barleeanum* (Fig. 5) indicates that environmental conditions in Storfjordrenna were similar to contemporary Norwegian fjords that are dominated by AW with a temperature of 6 - 8 °C and salinities of 34 - 35 (Husum and Hald, 2004). High total foraminiferal flux at the beginning of this period, as well as high foraminiferal species richness and biodiversity clearly point to AW conditions at the bottom (Hald and Korsun, 1997; Majewski and Zajączkowski, 2007; Włodarska-Kowalczuk et al., 2013). These conclusions are also supported by the heavier δ^{18}O, showing AW dominance and significant reduction in the amount of freshwater and ArW in Storfjordrenna (Fig. 8). The continuous presence of *Mytilus edulis* during the entire mid-Holocene points to the reduced inflow of the East Spitsbergen Current on account of the AW inflow (Feyling-Hansen, 1955; Forman, 1990; Salvigsen et al., 1992. The pathway and range of AW inflow to the western and north-eastern Svalbard during mid-Holocene were well described by Ślubowska-Woldengen et al. (2008) and Groot et al. (2014). Together with our results it is suggested that one of the main ways of AW inflow to the eastern Svalbard may have occurred trough Storfjordrenna.

Even though sediment accumulation rates were low, and grain size, as well as geochemical proxies, remain relatively constant during the mid-Holocene, the foraminiferal flux (including planktonic foraminifera) increased in two periods: of 9000 - 8000 cal yr BP and 6000 - 5500 cal yr BP, respectively (Fig. 4 and 6). In both cases the increase in IRD and *I. norcrossi* fluxes was followed by a slight depletion in δ^{18}O and heavier δ^{13}C suggesting minor cooling and likely seasonal sea-ice formation leading to beach sediment transport by shore ice. Our observations support earlier studies of the overall mid-Holocene shifts towards colder environment (Skirbekk et al., 2010; Rasmussen et al., 2012; Berben et al., 2014; Groot et al., 2014) and fluctuations in the glacial activity in the Svalbard region (e.g., Forwick and Vorren, 2007, 2009; Beaten et al., 2010; Ojala et al., 2014). Our data shows an increased supply of

IRD fraction to Storfjordrenna sediment followed by variation of $\delta^{18}O$, however, high flux of *M. barleeanum* associated with Atlantic-derived waters (Steinsund, 1994; Jennings et al., 2004; Fig. 5) indicates AW condition in southern Storfjorden throughout the whole mid-Holocene. The similar ameliorated condition with consistent AW inflow prevailed over the mid-Holocene also in the Kveithola Trough south of Storfjordrenna (Berben et al., 2014; Groot et al., 2014). To a small extent these two signals (AW inflow and higher IRD flux) are not necessarily in contradiction, since snow accumulation on land and inconsiderable glaciers advance depend on humid air transport from the ocean. Thus slight change in the atmospheric frontal zone over Svalbard could cause fluctuation of the glaciers range.

5.5 Glaciomarine unit III (late Holocene; 3600 cal yr BP to 1200 cal yr BP)

The late Holocene is characterized by a gradual increase in sediment accumulation rates followed by numerous sharp peaks of sand content and minor peaks of IRD flux, as well as increased Fe/Ca ratio, indicating ice growth on land (compare with e.g. Svendsen and Mangerud, 1997; Hald et al., 2004; Forwick and Vorren, 2009; Kempf et al., 2013), slightly enhanced iceberg calving and/or ice rafting over the core site. The IRD record shows few irregular small peaks in the late Holocene (Fig. 7), which, according to Hass (2002), could be correlated with enhanced sea currents increasing the drift of the icebergs. Forwick et al. (2010) suggested several glacier front fluctuations during the past two millennia in Sassenfjorden and Tempelfjorden (W Spitsbergen), hence we suppose increased iceberg calving occurred at Storfjordrenna during this time. However, increased IRD flux can also reflect deposition related to enhanced shore ice rafting. The latter explanation is in agreement with heavier $\delta^{18}O$ record (Fig. 6) indicating a minor cooling.

The mean grain size (<63µm) increases in late Holocene (Fig. 4) and may indicate stronger bottom current velocities and winnowing of fine grained sediments. Andruleit et al. (2006) observed similar increased erosive activity of bottom currents during late Holocene on the SW Svalbard shelf. This sudden increase in current velocities may be connected with (1) postglacial reorganization of oceanographic conditions, (2) relative lowering of the sea level during the postglacial isostatic rebound and/or (3) more intensive sea-ice formation enhancing formation of BSW, forming seasonal near-bottom dense water mass flowing over the coring site (Andruleit et al., 1996). Nevertheless, this process is still not fully understood.

The sharp increase in the foraminiferal flux (Fig. 4) pointing to the increased nutrient advection/upwelling and biological productivity at the coring site during the late Holocene was probably caused by variable hydrological conditions and most likely strong gradients leading to the formation of hydrological fronts. Our data shows increased fluxes of opportunistic species *E. excavatum* and *C. reniforme* as well as *N. labradorica* and *Islandiella* spp. *N. labradorica* and *Islandiella* spp. are abundant in areas with a high biological productivity in the upper surface waters (e.g. Hald and Steinsund, 1996; Korsun and Hald, 2000; Knudsen et al., 2012). Abundant, though variable *M. barleeanum*, documented in organic-rich mud within troughs of the Barents Sea (Hald and Steinsund, 1996) and in temperate fjords of Norway (Husum and Hald, 2004) points to high productivity in the euphotic zone leading to enhanced export of organic material/nutrients to the sea floor. Our data also shows high *N. pachyderma* flux throughout this unit, reflecting a significant increase of euphotic productivity at the coring site. However, low percentage of dextral specimens and *T.quinqueloba* point to low sea-surface temperatures (Fig. 6). This is in agreement with Rasmussen et al. (2014), who noted that after c. 3700 cal yr BP, Atlantic Water was only sporadically present at the surface. Cooling at the sea surface reflects the general trend in the Northern Hemisphere related to orbital forcing and reduction of summer insolation at high latitudes over the late Holocene (Wanner et al., 2008).

The last evidence of AW inflow to Edgøya area based on *M. edulis* is dated to 5000 cal yr BP (Hjort et al., 1995). After that time *M. edulis* remained absent until present days. However, its disappearance can rather be related to the freshening of surface water (Berge at al., 2006) and sea ice forcing as opposed to the extinction of AW in Storfjorden over the late Holocene (Rasmussen et al., 2007).

6 Conclusions

Multi-proxy analyses of one sediment core provide new information about the environmental development of the central part of Storfjordrenna off southern Svalbard since the late Bølling-Allerød. The main conclusions of our study are:

- Central Storfjordrenna was deglaciated before ~13,950 cal yr BP. The new data may help refine the future models of Svalbard-Barents Ice Sheet deglaciation.

- Between c. 13,450 to 11,500 cal yr BP, Storfjordrenna remained under the influence of Arctic Water masses with periodical sea-ice cover limiting the drift of icebergs. Nevertheless, at least three peaks of temperature increase during Younger Dryas stadial (12,800-11,500 cal yr BP) presumably led to seasonal disappearance of sea ice and significantly enhanced IRD flux indicating more sub-polar conditions.

- Atlantic Water started to flow onto the shelves off Svalbard and into Storfjorden during the early Holocene leading to a progressive warming and significant glacial melting. From c. 9600 cal yr BP, the Atlantic Water dominated the water column in Storfjordrenna.

- Environmental conditions off eastern Svalbard remained relatively stable from 9200-3600 cal yr BP with glaciers smaller than those of today. However, some small-scale cooling events (9000 - 8000 cal yr BP and 6000 - 5500 cal yr BP) indicate minor fluctuations in climate/oceanography of Storfjordrenna.

- A surface-water cooling and freshening occurred in Storfjordrenna during the late Holocene, synchronously with glacier growth and cooling on land. Even though, AW was still present in the deeper part of Storfjordrenna. The late Holocene in Storfjordrenna has been characterized also by increased bottom currents velocities however the driving mechanism is not fully understood.

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Table 1

Water mass characteristics in Storfjorden and Storfjordrenna (Skogseth et al., 2005, modified). The two main water masses are in bold.

Watermass names	Watermass characteristics	
	Temperature (°C)	Salinity
Atlantic Water (AW)	**>3.0**	**>34.95**
Arctic Water (ArW)	**<0.0**	**34.3-34.8**
Brine-enriched Shelf Water (BSW)	<-1.5	>34.8
Surface Water (SW)	>0.0	<34.4
Transformed Atlantic Water (TAW)	>0.0	>34.8

Table 2

AMS ^{14}C dates and calibrated ages.

Sample No	*Depth [cm]*	*Lab No.*	*Raw AMS ^{14}C BP*	*Calibrated years BP ± 2σ*	*Cal yr BP used in age model*	*Dated material*
St 20A 5/6	5	Poz-46955	1835 ± 30 BP	1200 – 1365	1285	*Cilliatocardium cilliatum*
St 20A 39	38.5	Poz-46957	2755 ± 30 BP	2245 – 2470	Not used	*Astarte crenata*
St 20 78/79	78	Poz-46958	2735 ± 30 BP	2177 – 2429	2320	*Astarte crenata*
St 20 110	109.5	Poz-46959	3450 ± 30 BP	3079 – 3323	3220	*Astarte crenata*
St 20 142	141.5	Poz-46961	6580 ± 40 BP	6850 – 7133	6970	*Astarte crenata*
St 20A 152	151.5	Poz-46962	7790 ± 40 BP	8018 – 8277	8160	*Astarte crenata*
St 20 157	156.5	Poz-46963	8610 ± 50 BP	8989 – 9288	9120	*Bathyarca glacialis*
St 20 251/252/253	252	Poz-46964	10,200 ± 60 BP	10,895 – 11,223	11,230	*Thracia sp*
St 20 396	395.5	Poz-46965	12,570 ± 60 BP	13,780 – 14,114	13,950	Bivalvia shell

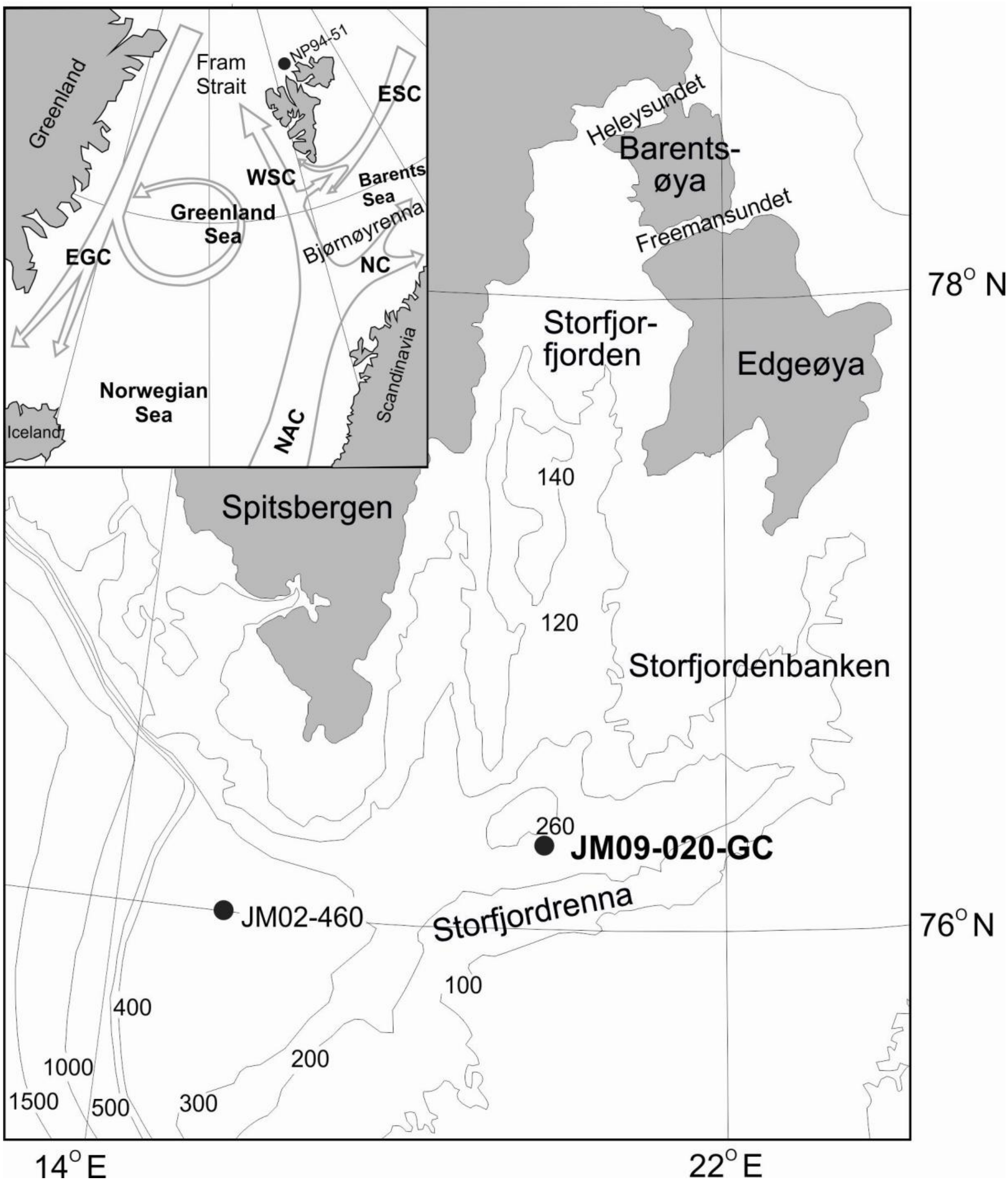


Fig. 1. Location map (a) showing the core site from this study (JM09-020-GC) and core site of JM02-460 (Rasmussen et al., 2007). The inlet map (b) shows the modern surface oceanic circulation in Nordic Seas and location of a core NP94-51 (Ślubowska et al., 2005). Abbreviations: NAC- Norwegian-Atlantic Current; WSC- West Spitsbergen Current; ESC- East Spitsbergen Current; EGC- East Greenland Current; NC- Norwegian Current. The cores JM02-460 and NP94-51 are discussed in the text.

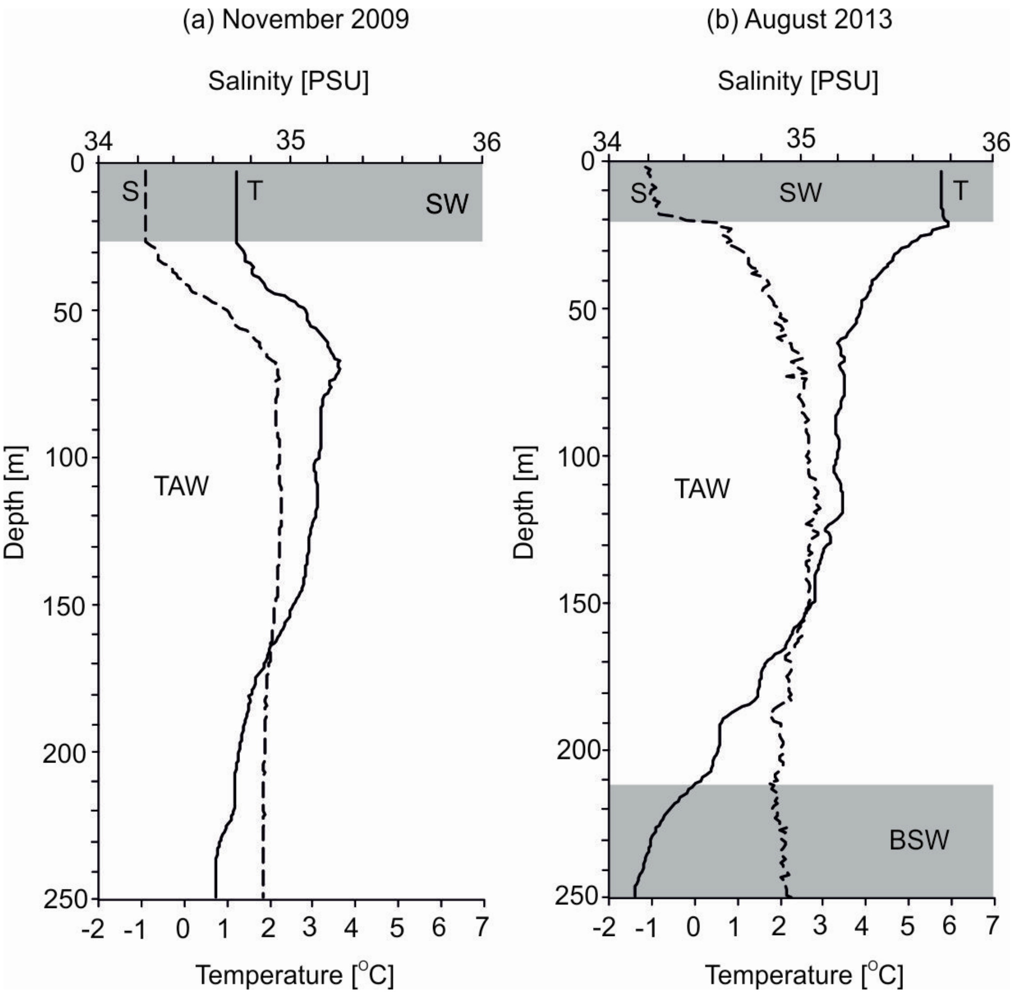


Fig. 2. Temperature and salinity versus depth, measured in November 5th 2009 (a) and in August 13th 2013 (b) at the site of core JM09-020GC. SW - Surface Water, TAW - Transformed Atlantic Water, BSW - Brine-enriched Shelf Water.

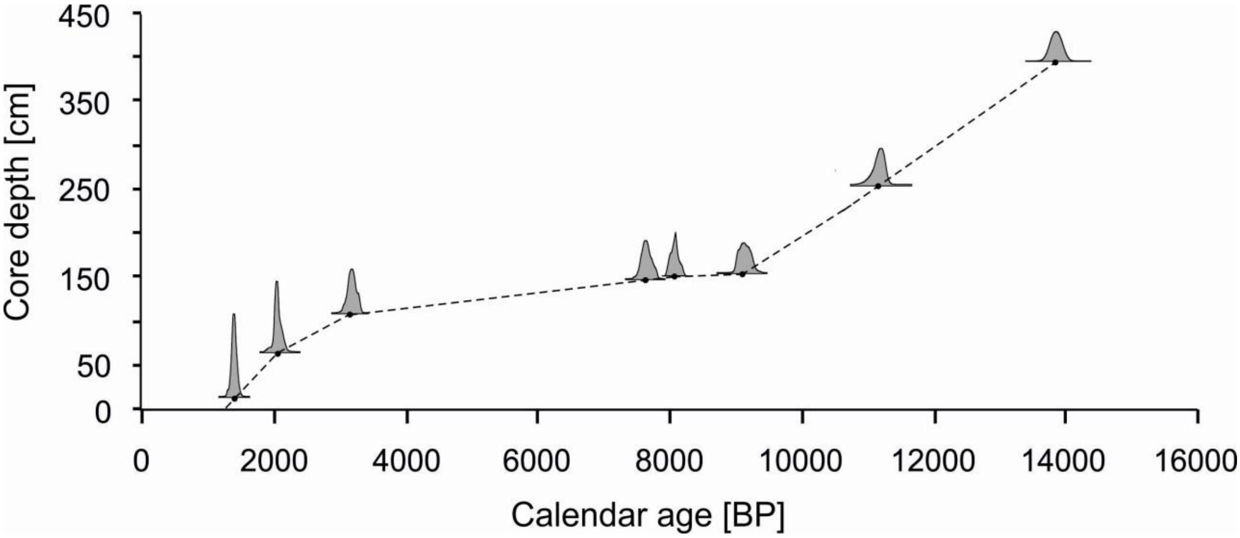


Fig. 3. Age-depth relationship for JM09-020-GC based on 8 AMS ^{14}C calibrated ages with 2-sigma age probability distribution curves. The chronology is established by linear interpolation between the calibrated ages.

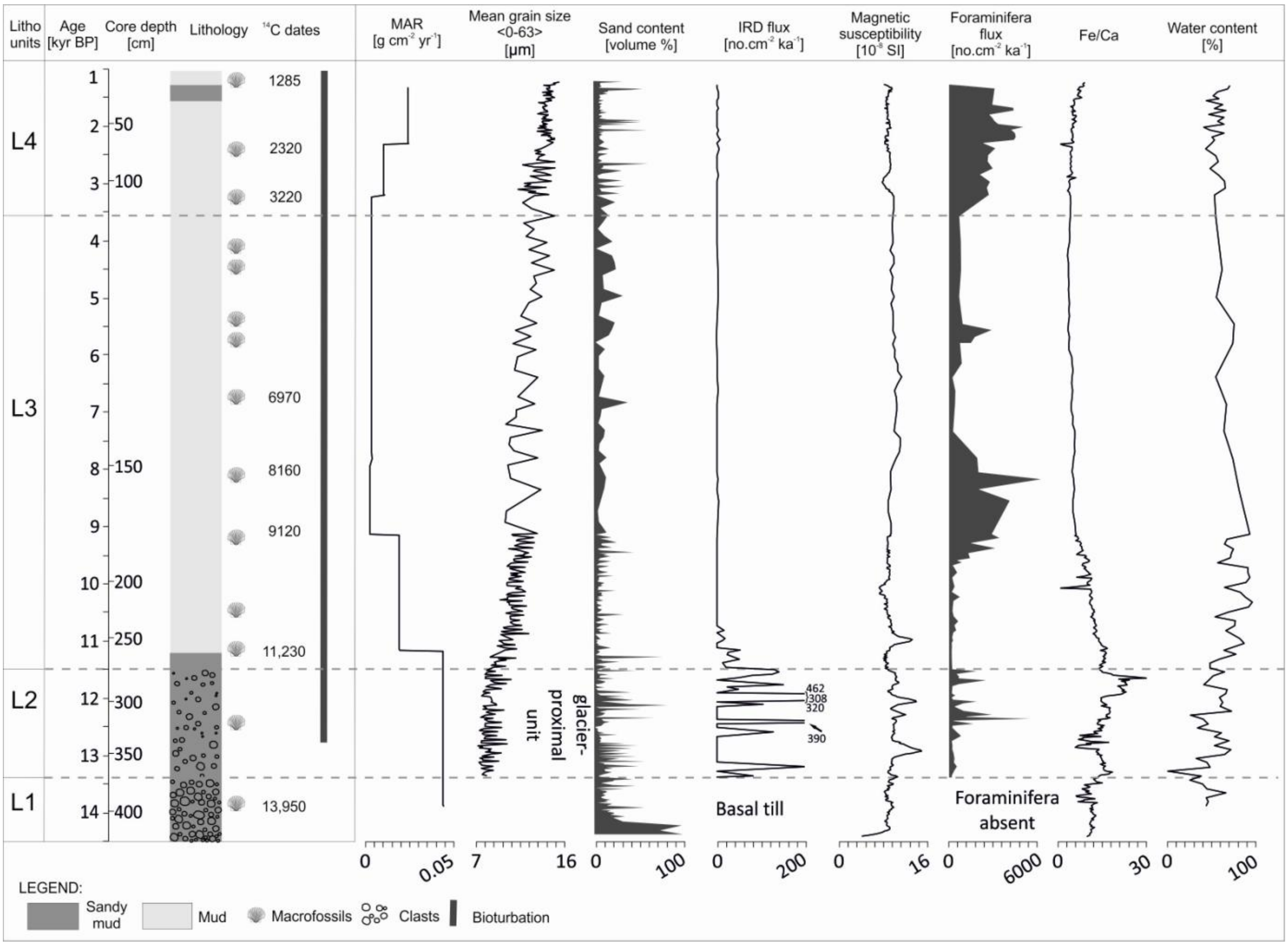


Fig. 4. Lithological log of core JM09-020GC. Lithology, ^{14}C dates, occurrence of bioturbation, mass-accumulation rates, mean grain size in the range of 0-63 µm, sand content, ice-rafted debris flux, magnetic susceptibility, foraminifera flux as well as Fe/Ca ratio and water content. The results are presented with lithostratigraphic units (L1-L4), versus calendar years (cal kyr BP) and core depth (cm).

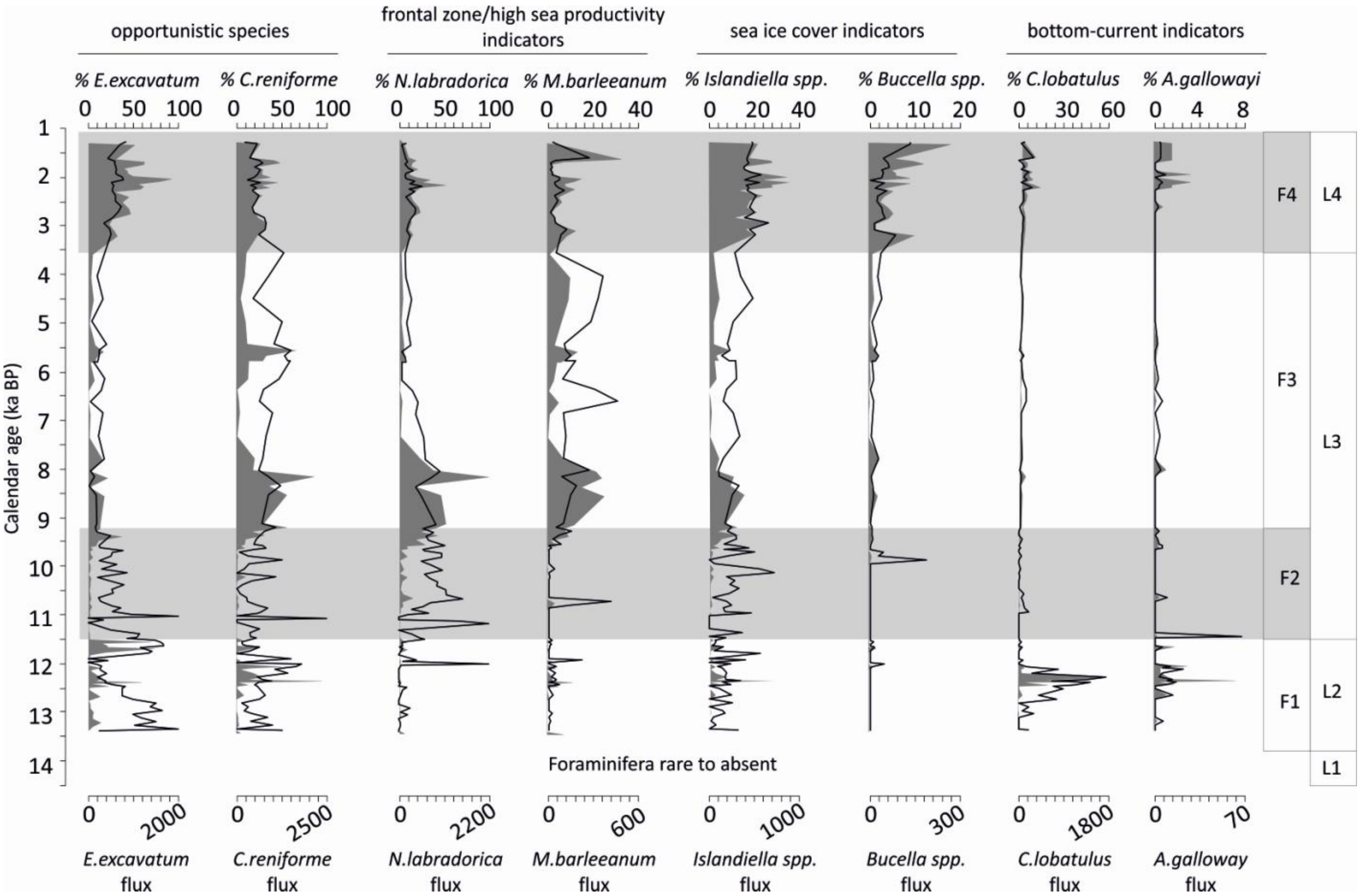


Fig. 5. Percentage distributions (upper scale; black line) and fluxes (no. cm^{-2} ka^{-1}; bottom scale; grey shading) of the most dominant benthic foraminiferal species plotted versus thousands of calendar years with indicated foraminiferal zonation (zones F1-F4) and lithostratigraphic units (L1-L4). Foraminiferal taxa are grouped based on their ecological tolerances described in the text.

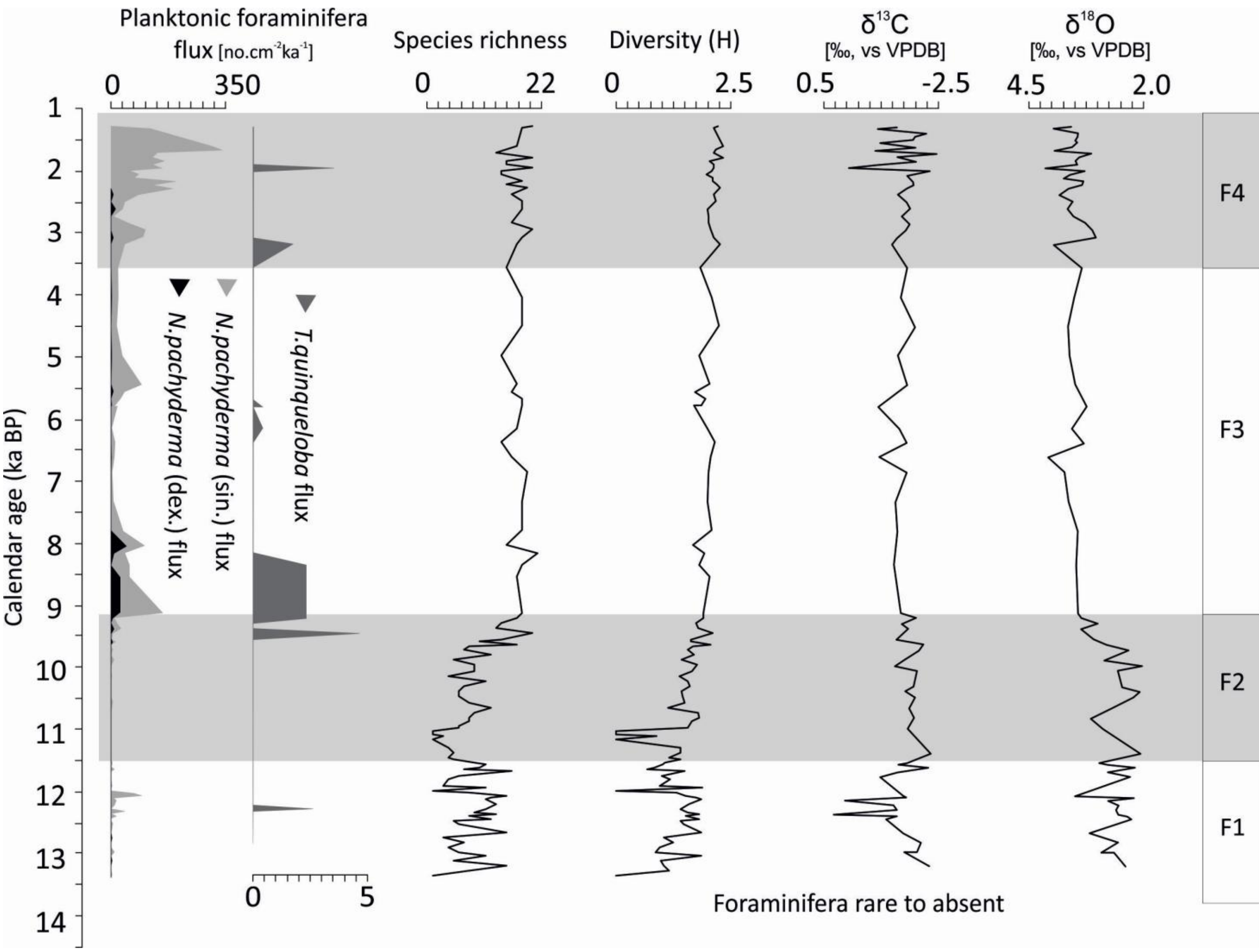
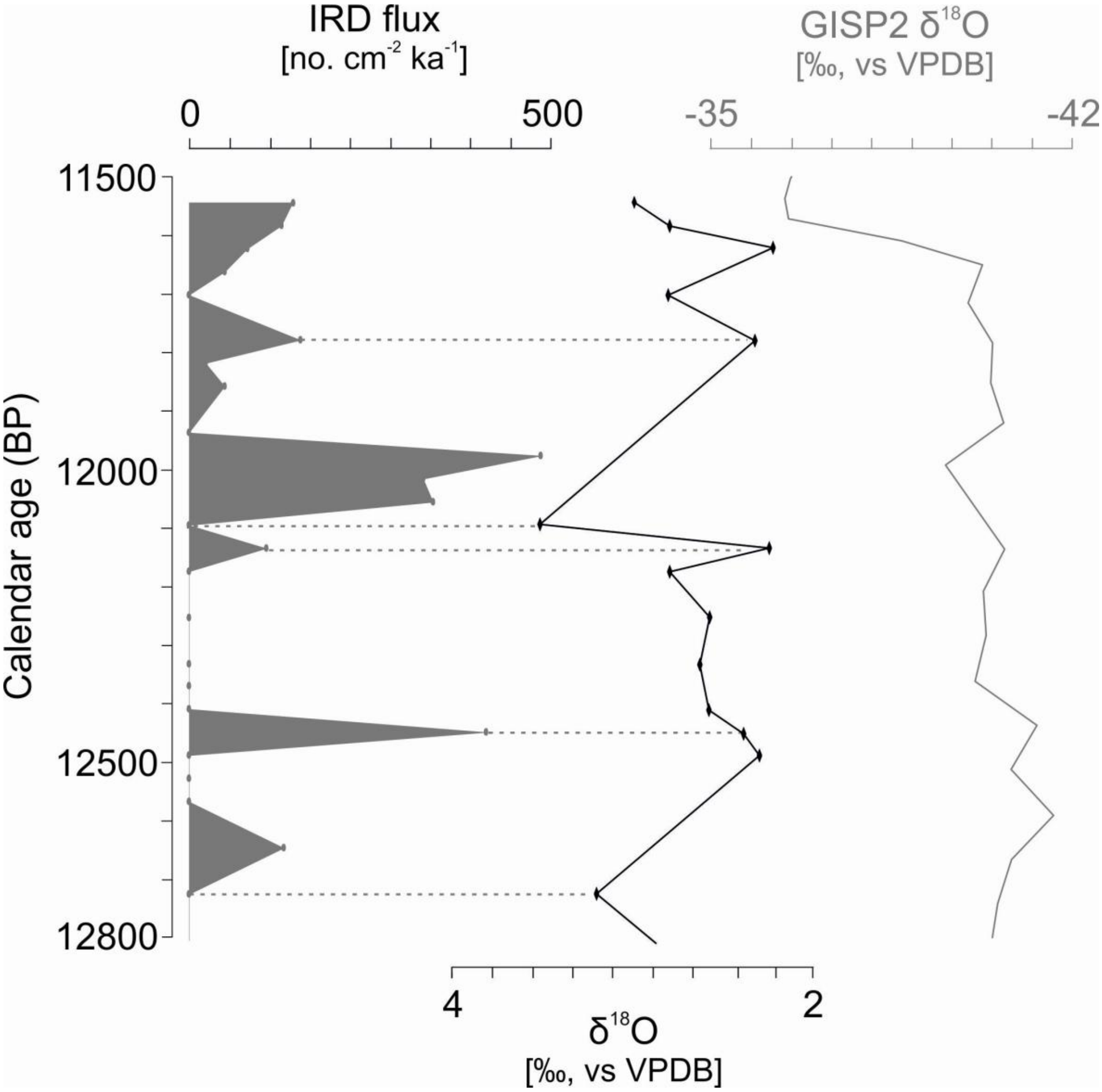


Fig. 6. Fluxes of planktonic foraminifera (no.cm^{-2}ka^{-1}), diversity parameters (species richness and Shannon - Wiener index) and stable oxygen and carbon isotope data ($\delta^{18}O$ and $\delta^{13}C$) plotted versus thousands of calendar years. The foraminiferal zonation (zones F1-F4) and lithostratigraphic units (L1-L4) are indicated.

1153 

1154

1155 Fig. 7 IRD flux (upper scale, grey shading) and oxygen stable isotopes records (bottom scale,
1156 black line) compared with oxygen stable isotopes records from ice core GISP2 from
1157 Greenland during the Younger Dryas period (12,800 cal yr BP to 11,500 cal yr BP).

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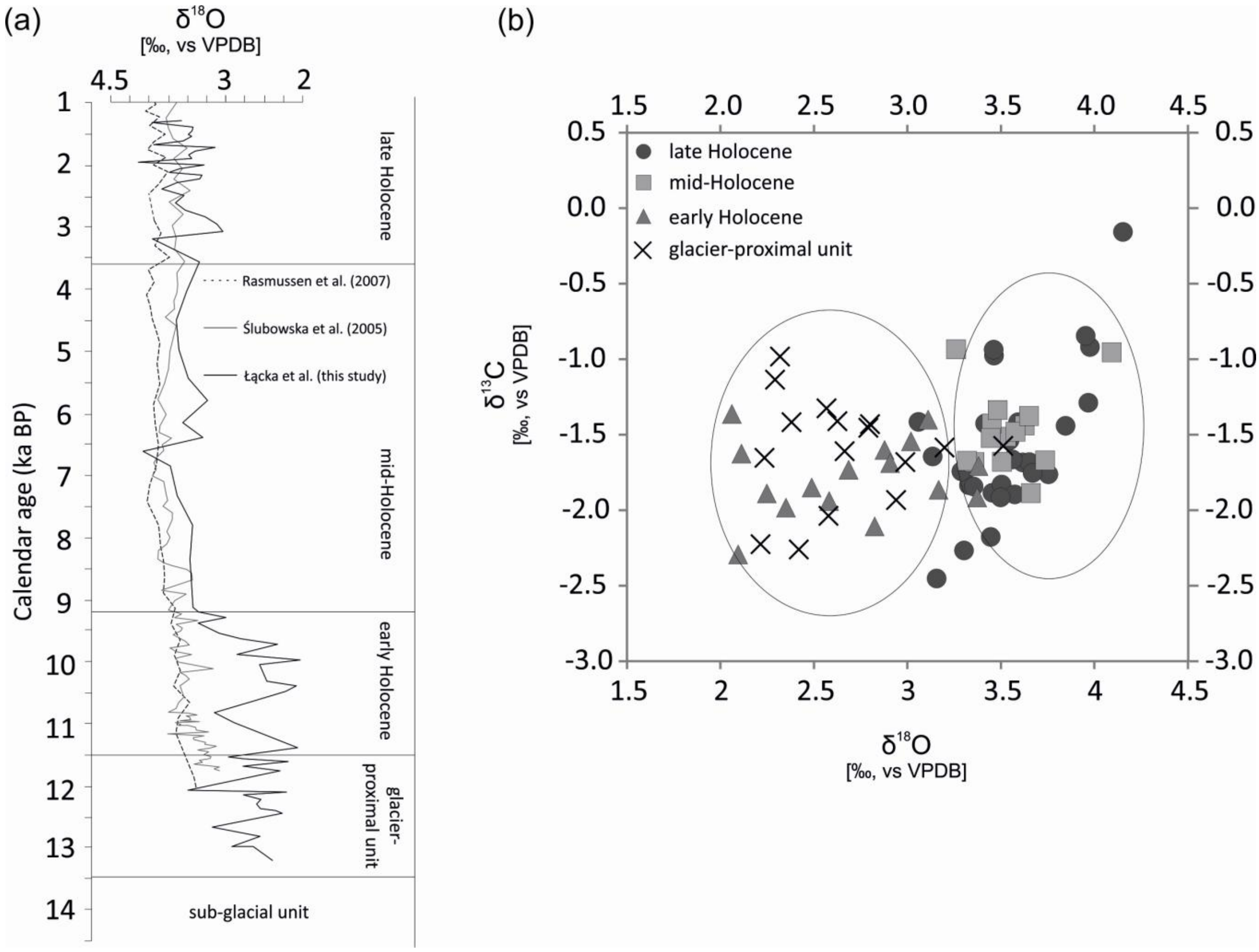

Fig. 8 (a) The comparison of δ^{18}O records (corrected for ice volume changes) between Łącka et al. (this study; black solid line) and Ślubowska et al. (2005; grey solid line) and Rasmussen et al. (2007; black dashed line) plotted versus thousands of calendar years. The δ^{18}O records after Łącka et al. (this study) were measured on *E.excavatum* f. *clavata* and the two latter ones (Ślubowska et al., 2005 and Rasmussen et al., 2007) were measured on *M.barleeanum*. (b) Scatter plot showing δ^{13}C versus δ^{18}O values from core JM09-020-GC (this study).