Holocene variability in sea-ice conditions in the eastern Baffin Bay-Labrador Sea – A north–south biomarker transect study

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Reconstructions of sea-surface conditions during the Holocene were achieved using three sediment cores from northeastern Baffin Bay (GeoB19948-3 and GeoB19927-3) and the Labrador Sea (GeoB19905-1) along a north–south transect based on sea-ice IP$_{25}$ and open-water phytoplankton biomarkers (brassicasterol, dinosterol and HBI III). In Baffin Bay, sea-surface conditions in the Early Holocene were characterized by extended (early) spring sea ice cover (SIC) prior to 7.6 ka BP. The conditions in the NE Labrador Sea, however, remained predominantly ice-free in spring/autumn due to the enhanced influx of Atlantic Water (West Greenland Current, WGC) from 11.5 until ~9.1 ka BP, succeeded by a period of continued (spring–autumn) ice-free conditions between 9.1 and 7.6 ka BP corresponding to the onset of Holocene Thermal Maximum (HTM)-like conditions. A transition towards reoccurring ice-edge and significantly reduced SIC conditions in Baffin Bay is evident in the Middle Holocene (~7.6–3 ka BP) probably caused by the variations in the WGC influence associated with the ice melting and can be characterized as HTM-like conditions. These HTM-like conditions are predominantly recorded in the NE Labrador Sea area shown by (spring–autumn) ice-free conditions from 5.9–3 ka BP. In the Late Holocene (last ~3 ka), our combined proxy records from eastern Baffin Bay indicate low in-situ ice algae production; however, enhanced multi-year (drifted) sea ice in this area was possibly attributed to the increased influx of Polar Water mass influx and may correlate with the Neoglacial cooling. The conditions in the NE Labrador Sea during the last 3 ka, however, continued to remain (spring–autumn) ice-free. Our data from the Baffin Bay–Labrador Sea transect suggest a dominant influence of meltwater influx on sea-ice formation throughout the Holocene, in contrast to sea-ice records from the Fram Strait area, which seem to follow predominantly the summer insolation trend.

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Sea ice with its strong seasonal and interannual variability plays a fundamental role in the Earth’s climate system as it influences the heat, moisture, albedo and gas exchange between the ocean and the atmosphere as well as deep-water formation (Dieckmann & Hellmer 2003; McPhee et al. 2009; Thomas & Dieckmann 2010; Morison et al. 2012). Over the last few decades, the Arctic sea ice has undergone a considerable reduction in its extent and thickness due to modern climate warming (Kinnard et al. 2011; Serreze & Stroeve 2015; NSIDC 2020) and indeed this has raised considerable concerns about sea-ice variability in the future (Bhatt et al. 2014). It is expected that this ongoing reduction of Arctic sea ice will continue and amplify, further decreasing the Arctic Ocean albedo thus increasing sea-surface temperatures (SSTs; Manabe et al. 1992; Screen & Simmonds 2010). To understand these rapid changes in sea ice cover from a long-term perspective, there is a need for a better understanding of the sea-ice variability as well as its dynamics and interaction with climate throughout the geological past (Jakobsson et al. 2010).

Various proxies preserved in marine sediments may indicate indirect changes in sea-ice occurrence and variability, as inferred from micropalaeontology (e.g. dinoflagellate cysts, diatoms, and foraminifera assemblages) (Jennings et al. 2002; Gersonde et al. 2005; de Vernal et al. 2013a, b), geochemistry (Hillaire-Marcel & de Vernal 2008) and sedimentology (e.g. ice-rafted debris; Polyak et al. 2010). The biomarker proxy ‘IP$_{25}$’, exclusively synthesized by sea-ice algae and well preserved in the sediments, however, seems to be a direct indicator for the presence of seasonal Arctic sea ice as shown in the pioneer study by Belt et al. (2007) (see Belt 2018 for a recent review). Even more detailed and semi-quantitative reconstructions of sea ice cover can be obtained when combining IP$_{25}$ with open-water phytoplankton biomarkers such as brassicasterol, dinosterol and/or a highly branched unsaturated triene (HBI III) (‘PIP$_{25}$ index’ Muller et al. 2011; Belt et al. 2015, 2019; Smik et al. 2016).

Arctic outflow through Baffin Bay, via the Labrador Sea, acts as a substantial contributor of fresh water to the North Atlantic, as suggested by modelling and observational studies (Bunker 1976; Tang et al. 2004; Serreze et al. 2006; Aksenov et al. 2010). Baffin Bay is characterized by a sea ice cover during most time of the year; however, further to the south, the eastern Labrador Sea is predominantly ice-free (Tang et al. 2004). Thus the Baffin
The surface and subsurface waters in Baffin Bay and the Labrador Sea form a counter-clockwise gyre due to interaction of northward-flowing warm high salinity water transported by the West Greenland Current (WGC) with southward flowing cold polar sourced Arctic Water transported via the Baffin Current (BC) (Fig. 1A; Tang et al. 2004; Ribergaard et al. 2008). The WGC carries water from the warmer more saline western branch of the Irminger Current, modified by the less saline Arctic Water from the East Greenland Current (EGC) and local meltwater discharge along the SW Greenland coast. The WGC flows northwards along the west Greenland coast and typically turns west in the Melville Bight before merging with the BC in northeastern Baffin Bay. The BC flows south along Baffin Island through Davis Strait before joining the Labrador Current (LC). In the western Labrador Sea, the LC overlies the WGC, which partly turns west and south again near Davis Strait.

Sea ice cover in Baffin Bay is variable and may range from 0 to ~10 months per year, with much of Baffin Bay covered by near-continuous sea ice during winter (Fig. 1B) (Tang et al. 2004). Ice growth first starts in northwestern Baffin Bay at the end of September and expands southeastwards, with a maximum extent in March, sometimes even reaching the NW Labrador Sea. After melting during the summer months it attains a minimal extent in September (Wang et al. 1994; Tang et al. 2004). The warm WGC negatively affects sea-ice formation and prevents its further growth along eastern Baffin Bay. A large interannual variability in sea ice has a major impact not only on ice conditions but on phytoplankton blooms and therefore marine productivity as well, which is coupled with the strong seasonality in surface air temperatures and wind patterns (Sakshaug 2004; Tang et al. 2004).

Environmental setting

The three cores used in this study were collected during RV ‘Maria S. Merian’ cruise MSM44 (Dorschel et al. 2015). Site information, core lengths, and sampling intervals are provided in Table 1. Cores GeoB19948-3 and GeoB19927-3 were taken from northeastern Baffin Bay, whereas core GeoB19905-1 was taken from the northeastern Labrador Sea (Fig. 1, Table 1). Based on visual description and colour measurements of core GeoB19948-3 the upper 280 cm are mainly composed of olive-grey to silty clay homogeneous sediments (Dorschel et al. 2015). In core GeoB19927-3 the lowermost part is characterized by non-homogenous silt and fine sand sediments embedded with dropstones and sharp colour boundaries. The upper part of this core can be mainly described as homogenous olive-grey silty-clay sediments (Dorschel et al. 2015). In core GeoB19905-1, sediments are mainly composed of olive-grey muds with some sandy layers at the lower part of the core and sharp colour transitions displayed by distinct lithological units (LU); for a more detailed description, however, see Dorschel et al. (2015).

Table 1. Coordinates of the three cores used in this study.

| Core ID      | Region                | Latitude (N) | Longitude (W) | Water depth (m) | Core length (cm) |
|-------------|-----------------------|--------------|---------------|-----------------|------------------|
| GeoB19948-3 | Northern Baffin Bay   | 75°46.10’    | 64°08.57’     | 778             | 1018             |
| GeoB19927-3 | Northeastern Baffin Bay| 73°35.26’    | 58°05.66’     | 932             | 1147             |
| GeoB19905-1 | Eastern Labrador Sea  | 64°21.68’    | 52°57.70’     | 485             | 1045             |
Chronostratigraphy

The chronostratigraphical framework of the upper 280 cm of core GeoB19948-3 (Fig. 2) is based on four AMS $^{14}$C dates measured on 0.57–1.12 mg of foraminifera from the >100 µm fraction (Table 2), performed directly on the CO$_2$ gas with the MICADAS (Mini Carbon Dating System) at the Alfred Wegener Institute in Bremerhaven, Germany (for methodological details see Wacker et al. 2013). Radionuclide analyses ($^{210}$Pb, $^{40}$K, $^{137}$Cs) were performed on core top sediments (depths 2, 6 and 10 cm) at the Bremen State Radioactivity Measurements Laboratory. Subsequently, the AMS dates were calibrated to calendar years using PaleoDataView (PDV) (Langner & Mulitza 2019), which uses modelled reservoir ages (Butzin et al. 2017; Table 1) to calibrate radiocarbon ages against IntCal13 (Reimer et al. 2013). The appropriate reservoir ages are chosen

![Fig. 2. Age models of GeoB19948-3 (orange), GeoB19927-3 (green) and GeoB19905-1 (red). For GeoB19927-3 and GeoB19905-1 the bold lines show weighted mean age after modelling in BACON, while the lighter shades indicate 95% confidence intervals (CIs). For GeoB19948-3 the age-depth relationship is given by linear interpolation between the AMS $^{14}$C dates, shown as bold line. Stippled lines (GeoB19905-1) indicate areas of less robust age control. Rectangles denote 95% CIs of the AMS $^{14}$C dates after calibration with PDV. In the centre and bottom panels the derived sedimentation and accumulation rates are shown, including 5-point running averages where applicable.](image-url)
automatically based on the sample’s radiocarbon age and location of the core. The calibrated ages were subsequently used to construct the final age model by linear interpolation between the calibrated ages and linear extrapolation between the topmost age and the core top. Due to the limited number of AMS 14C dates, the age model is still preliminary.

The chronostratigraphy of core GeoB19927-3, previously published by Saini et al. (2020) (Fig. S1) is based on 12 AMS 14C dates and radionuclide analyses (210Pb, 40K, 137Cs) and was developed using ‘BACON’ software (Bläauw & Christen 2011) assuming a constant reservoir age of 140±35 years. However, to ensure consistency between the age models, we recalibrated the AMS dates (Fig. 2) using PDV (Butzin et al. 2017), which uses modelled reservoir ages to calibrate radiocarbon ages against IntCal13 (Reimer modelled reservoir ages to calibrate radiocarbon ages is based on 12 AMS14C dates, previously published by only about 0.2 ka younger (Fig.S1). BP, which is almost the same as the original age model, i.e. GeoB19927-3 covers a time interval from 8.333, where 8.333 is the stratigraphical framework. against IntCal13 (Reimer et al. 2013) (see Fig. S1 for comparison). According to the new age model, core GeoB19927-3 covers a time interval from c. 0 to 10.1 ka BP, which is almost the same as the original age model, i.e. only about 0.2 ka younger (Fig. S1).

The chronostratigraphy of core GeoB19905-1 (Fig. 2) is based on 12 AMS 14C dates, previously published by Weiser et al. (2021) (see data sets available at https://doi.org/10.1016/j.quascirev.2021.106833). The age model was constructed using a combination of the PDV program (Langner & Mulitza 2019) and the open source software package BACON (Bläauw & Christen 2011). According to the published age model, core GeoB19905-1 covers a time interval from c. 0 to 11.5 ka BP, with a hiatus at depth 640 cm between 5.9 to 7.6 ka BP. For details on the methodological approach, the reader is referred to the original publication.

Bulk parameters (TOC and CaCO3)

For organic-geochemical analyses, freeze-dried and homogenized subsamples were taken (n = 60, at 4-cm intervals for core GeoB19948-3 and n = 207, at 5-cm intervals for core GeoB19905-1) and stored in glass vials at −20 °C. TOC contents were measured using a carbon-sulphur ELTRA analyser (CS-800, ELTRA) after carbonate removal with hydrochloric acid (37%, 500 µL). The error of our TOC measurements is ±0.02%. The total amount of carbon (TC) was measured on untreated samples using a carbon-nitrogen-sulphur analyser (Elementar-III, Vario). Assuming the predominant carbonate phase is calcite, carbonate contents were calculated as CaCO3 = (TC − TOC)×8.333, where 8.333 is the stoichiometric calculation factor.

Sea-ice biomarkers (IP25, HBI III and sterols)

For biomarker analyses, we extracted ~4 g of freeze-dried and homogenized sediment using dichloromethane: methanol (2:1 v/v) as a solvent for ultrasonication (3×15 min). Prior to extraction, 9-octyleptadec-8-ene (9-OHD; 0.1 µg per sample), 7-hexynonadecane (7-HND; 0.076 µg per sample), 5α-androstan-3β-ol (androstanol; 10.7 µg per sample) and 2,6,10,15,19,23-hexamethyldieicosane (squalane; 3.2 µg per sample) were added for quantification of biomarkers. The extracts were separated by open silica (SiO2) column chromatography with n-hexane (5 mL) and ethyl-acetate: n-hexane (9 mL, 2:8 v/v) as eluent into the hydrocarbon and sterol fractions, respectively. The sterol fraction was silylated using 200 µl bis-trimethylsilyl-trifluoroacet-amide (BSTFA; 60 °C, 2 h).

Two different gas chromatography-mass spectrometers (GC-MS) with similar basic configuration were used to qualify and quantify the hydrocarbon and sterol fractions. The quantification of the HBI fraction was carried out with a gas chromatograph (Agilent Technologies GC6850, 30 m DB-1MS column, 0.25 mm id, 0.25 µm film) coupled to an Agilent Technologies 5977C VL MDSD mass selective detector (Triplet-Axis Detector, 70 eV constant ionization potential, scan 50–550 m z−1, 1 scan s−1, ion source temperature 230 °C). The quantification of the sterols (quantified as trimethylsilyl ethers) was carried out with a GC Agilent 6850 (30 m DB-1MS column, 0.25 mm id, 0.25 µm film) coupled to an Agilent Technologies 5975C VL MSD mass selective detector (Trip-Axis Detector, 70 eV constant ionization potential, scan 50–550 m z−1, 1 scan s−1, ion source temperature 230 °C). The quantification of the sterols was carried out with a gas chromatograph (Agilent Technologies GC6850, 30 m DB-1MS column, 0.25 mm id, 0.25 µm film) coupled to an Agilent Technologies 5975C VL MSD mass selective detector. GC measurements were carried out with the following temperature programme: 60 °C (3 min), 150 °C (15 °C min−1), 320 °C (10 °C min−1), 320 °C (15 min isothermal) for the hydrocarbons and 60 °C (2 min), 150 °C (15 °C min−1), 320 °C (3 °C min−1), 320 °C (20 min isothermal) for the sterols. Helium served as a carrier gas.

Table 2. Results of the AMS dating and subsequent calibration from core GeoB19948-3. Reservoir ages are in respect to IntCal13. min. and max. ages respectively correspond to the 5% and 95% confidence intervals defined by PDV. The last column denotes the final ages used for the stratigraphical framework.

| AMS dating, MICADAS, AWI Bremerhaven | Reservoir correction and calibration, PaleoDataView |
|------------------------------------|-----------------------------------------------|
| Lab. ID  | Depth (cm) | 14C age (a BP) | 14C error (a) | Material | Reservoir age (a) | Min. age (cal. a BP) | Max. age (cal. a BP) | Weighted mean age (cal. a BP) |
| AWI-     |            |               |              |         |                |                  |                    |                                |
| 1252.1.1 | 14         | 2443          | 108          | Mixed foraminifera | 734±64     | 1354            | 1890             | 1625                |
| 1253.2.1 | 150        | 5431          | 198          | Mixed benthic foraminifera | 818±58     | 4824            | 5874             | 5281                |
| 1253.1.1 | 150        | 5427          | 114          | N. pachyderma sin  | 818±58     | 4892            | 5591             | 5285                |
| 1475.1.1 | 274        | 7789          | 184          | Mixed foraminifera | 660±61     | 7623            | 8336             | 7965                |
carrier gas (1 mL min\(^{-1}\) constant flow). Specific compound identification was based on the comparison of gas chromatography retention times with those of reference compounds and published mass spectra (Boon et al. 1979; Volkman 1986; Belt et al. 2007; Brown & Belt 2016). For the quantification of IP\(_{25}\) and HBI III (Z-isomer, generally referred to as ‘HBI III’) their molecular ion ratio (m/z 350 for IP\(_{25}\) and m/z 346 for HBI III) in relation to the abundant fragment ion m/z 266 of internal standard (7-HND) was used (in selected ion monitoring mode, SIM). The different responses of these ions and a detailed quantification method are given by Fahl & Stein (2012). For the quantification of the sterols, the molecular ions m/z 470 for brassicasterol (as 24-methylcholesta-5,22E-dien-3\(\beta\)-ol) and m/z 500 for dinosterol (4α,23,24R-trimethyl-5α-cholest-22E-en-3\(\beta\)-ol) were used in relation to the molecular ion m/z 348 for the internal standard androstanol.

The PIP\(_{25}\) indices were calculated by combining IP\(_{25}\) with different phytoplankton markers for semi-quantitative sea-ice reconstruction according to Müller et al. (2011):

\[
PIP_{25} = \frac{[IP_{25}]}{([IP_{25} + c \times p])}
\]

where c = mean [IP\(_{25}\)]/mean [p] and p is the phytoplankton biomarker content (p = B (brassicasterol) or D (dinosterol) or III (HBI III)). Recently, based on surface sediments from Baffin Bay, Kolling et al. (2020) suggested that sea-ice indices P\(_D\)IP\(_{25}\) and P\(_B\)IP\(_{25}\) may indicate late spring and/or autumn conditions, while P\(_II\)IP\(_{25}\) may record more the early spring and/or late winter (ice-edge) conditions. Here we concentrate on the P\(_II\)IP\(_{25}\) as for this proxy other published records from Baffin Bay are also available. For P\(_D\)IP\(_{25}\) and P\(_B\)IP\(_{25}\) see Saini et al. (2020) and the complete data at https://doi.pangaea.de/10.1594/PANGAEA.911365.3

Calculation of accumulation rates

Biomarker contents were converted into accumulation rates by using the following equations (e.g. Stein & Macdonald 2004):

\[
\text{Bulk AR} = LSR \times DBD
\]

\[
\text{TOC AR} = \text{Bulk AR} \times \text{TOC} / 100
\]

\[
\text{CaCO}_3 \text{ AR} = \text{Bulk AR} \times \text{CaCO}_3 / 100
\]

\[
\text{BMAR} = \text{Bulk AR} \times \text{BM}
\]

where LSR = sedimentation rate (cm ka\(^{-1}\)), DBD = dry bulk density (g cm\(^{-3}\)), TOC = total organic carbon content (%), CaCO\(_3\) = carbonate content (%), BM = biomarker content (µg g\(^{-1}\)), Bulk AR = total sediments accumulation rate (g cm\(^{-2}\) ka\(^{-1}\)), TOC AR = total organic carbon accumulation rate (g cm\(^{-2}\) ka\(^{-1}\)), CaCO\(_3\) AR = carbonate accumulation rate (g cm\(^{-2}\) ka\(^{-1}\)), and BMAR = biomarker accumulation rate (µg cm\(^{-2}\) ka\(^{-1}\)).

Results

Core chronology and sedimentation rates

Core GeoB19948-3. – Given the scarcity of calcareous material in the sediment resulting in only four AMS \(^{14}\)C dates for core GeoB19948-3, only a preliminary chronological framework could be constructed. Still, this framework suggests continuous sedimentation since \(~8\) ka BP (Fig. 2), with sedimentation rates in the order of 40–50 cm ka\(^{-1}\). No excess \(^{210}\)Pb and/or \(^{137}\)Cs were found in the top sediments and thus the age of the topmost date (1.6 ka BP, 14 cm core depth) suggests that the core top is not of recent age, but instead \(~1.2\) ka old.

Core GeoB19927-3. – The updated age model of core GeoB19927-3 suggests continuous sedimentation since \(~10.1\) ka BP (Fig. 2). Sedimentation rates vary between 40 and 200 cm ka\(^{-1}\), although high sedimentation rates of about 100–200 cm ka\(^{-1}\) prevail in the majority of the record, except for a prominent decrease starting at 2.6 ka BP leading to low values of \(<50\) cm ka\(^{-1}\) after 0.7 ka to present.

Core GeoB19905-1. – The age model of core GeoB19905-1 (Weiser et al. 2021) suggests continuous sedimentation from 11.5 to 7.6 ka BP and since 5.9 ka BP (Fig. 2). A hiatus spanning about 1.7 ka in between 11.5 and 9.5 ka BP is followed by a strong increase in sedimentation rates – with high values of \(~100\) cm ka\(^{-1}\) between 11.5–8.2 ka BP and decreasing to about 80 cm ka\(^{-1}\) from 8.2–7.6 and 5.9–2.4 ka BP. A strong increase in sedimentation rates to 80–100 cm ka\(^{-1}\) is recorded after 2.4 ka BP that rises even further up to 150, 200 and 250 cm ka\(^{-1}\) after 1.4, 0.7 and 0.3 ka BP, respectively.

Organic geochemical bulk parameters and biomarkers

The organic-geochemical and biomarker records of core GeoB19927-3, excluding CaCO\(_3\), have already been described and published by Saini et al. (2020). Thus, here only the new data of cores GeoB19948-3 and GeoB19905-1 are presented. In the discussion, however, data from all three cores are included. The results of all biomarkers contents are plotted in µg g\(^{-1}\) TOC and as accumulation rates in µg cm\(^{-2}\) ka\(^{-1}\) (Figs S3, S4). They are also presented in µg g\(^{-1}\) sediment in Figs S2–S4. The distribution of contents and accumulation rates of
biomarkers and bulk parameters reveal significant down-core variations.

**Core GeoB19948-3.** The GeoB19948-3 record presented here extends to ~8 ka BP. CaCO₃ contents vary between 3.6 to 1% over the last ~8 ka (Fig. 3B). Higher values of about 2.4% (mean) occur from c. 8 to 5.3 ka BP followed by lower contents of 1.9% (mean) towards the top of the core. TOC contents vary between about 0.8 to 0.4% over the last ~8 ka (Fig. 3D), and display higher values of about 0.7% (mean) from c. 8 to 5.3 ka BP, followed by lower contents of 0.6% (mean) towards the top of the core. HBI III contents were typically near zero over the last ~8 ka, except for two prominent peaks of 0.6 and 0.4 µg g⁻¹ TOC at about 6.5 and 1.6 ka BP, respectively (Fig. 3F). The phytoplankton sterols dinosterol and brassicasterol display higher contents from 8 to 5.3 ka BP, followed by a gradual decrease towards the top of the core (Fig. 3H, J). IP₂₅ contents vary between 0.2 and 2.0 µg g⁻¹ TOC and display higher values from 8 to 5.3 ka BP, followed by lower values towards the core top (Fig. 3L).

The accumulation rates of TOC range from 0.10 to 0.26 g cm⁻² ka⁻¹ and are marked by a gradual decrease from c. 8 to 1.2 ka BP (Fig. 3E). The CaCO₃ accumulation rates show a similar generally decreasing trend from 8 to 1.2 ka BP and vary between about 0.2 to 1 g cm⁻² ka⁻¹ (Fig. 3C). HBI III accumulation rates are typically low (near zero) throughout the Middle-to-Late Holocene (mean ~0.02 µg cm⁻² ka⁻¹; Fig. 3G), except for a prominent peak around 6.7–6.3 ka BP. This peak value, however, should be interpreted with caution due to the limited number of AMS ¹⁴C dates. The accumulation rates of the phytoplankton biomarkers dinosterol and brassicasterol vary between 0.7 and 5.7 µg cm⁻² ka⁻¹ and show a gradual decrease from c. 8–1.2 ka BP (Fig. 3I, K). IP₂₅ accumulations rates range between 0.03 and 0.5 µg cm⁻² ka⁻¹ and are relatively high in the early part of the Middle Holocene between 8–5.3 ka BP, followed by constantly low values between c. 5.3–1.2 ka BP.

![Figure 3](https://example.com/figure3.jpg)

**Figure 3.** Records of core GeoB19948-3 including bulk parameters and biomarkers and their accumulation rates. A. Bulk accumulation rate. B. CaCO₃ content. C. CaCO₃ accumulation rate. D. Total organic carbon (TOC) content. E. TOC accumulation rate. F. HBI III (µg g⁻¹ TOC). G. HBI III accumulation rate. H. Dinosterol (µg g⁻¹ TOC). I. Dinosterol accumulation rate. J. Brassicasterol (µg g⁻¹ TOC). K. Brassicasterol accumulation rate. L. IP₂₅ (µg g⁻¹ TOC). M. IP₂₅ accumulation rate. Black solid triangles mark the AMS ¹⁴C dates. All plots are shown vs. age in 1000 years before present (ka BP).
BP (Fig. 3M). The PIIIIP25 index varies in the range 0.1 to 0.8 and displays higher values in the Early Holocene prior to 7.6 ka BP; show decreased values in the Middle Holocene from c. 7.6 to 3 ka BP, followed by generally reduced values in the Late Holocene (Fig. 5H). Moreover, PIIIIP25 attains two minimum values of ~0.29 and ~0.15 at about 6.7–6.3 (Middle Holocene) and ~1.8–1.3 ka BP (Late Holocene), respectively.

Core GeoB19905-1. – TOC contents vary between 0.2 and 2.4%, showing the lowest values between ~11.5 to 8.2 ka BP followed by a sharp rise afterwards, and acquire their highest (mean 1.8%) values during the last c. 5 ka (Fig. 4B). CaCO3 contents range from 0.7 to 21.8% throughout the record and display continuously low values of about 3.3% (mean) between ~11.5 and 9.1 ka BP (Fig. 4D). They increase to maximum values of 21%, attaining the maximum at about 4.4 ka BP and drop to generally lower values (mean 3.3%) between 2.1 to 0.4 ka BP, followed by a minor rise to about 5% (mean) in the upper part of the core in the last c. 0.4 ka BP. The phytoplankton biomarkers dinosterol and brassicasterol co-vary between 3 and 43 µg g⁻¹ TOC throughout the Holocene (Fig. 4F, H). Phases with variable but high values (mean 11.4 µg g⁻¹ TOC) of dinosterol and brassicasterol were observed between ~11.5 and 8.6 ka BP; however, they show no increase in µg g⁻¹ sediment in this period (Fig. S4). A slight decrease in dinosterol contents
(mean 10.4 μg g⁻¹ TOC) occurs between 8.6–7.6 and 5.9–2.1 ka BP. Afterwards, a strong increase in dinosterol content is recorded from 2.1–0.4 ka BP followed by an even stronger rise to about 40 μg g⁻¹ TOC during the last c. 0.4 ka BP. IP₃ and HBI III are absent throughout the record of the last 11.5 ka BP (Fig. 4J).

The accumulation rates of TOC, CaCO₃ and phytoplankton biomarkers dinosterol and brassicasterol vary between about 0.2–3 g cm⁻² ka⁻¹, 1–14 g cm⁻² ka⁻¹, 1–116 μg cm⁻² ka⁻¹ and 2–74 μg cm⁻² ka⁻¹, respectively (Fig. 4C, E, G, I). IP₂₅ accumulation rates are zero (as IP₂₅ = 0) throughout the record (Fig. 4J). Overall, the accumulation rates of TOC and the phytoplankton biomarkers dinosterol and brassicasterol remain relatively low and constant until about 7.6 ka BP and show slightly increased values from 5.9 to about 2 ka BP. A period of a strong increase in the accumulation rates of dinosterol and brassicasterol followed during the last ~2 ka BP. However, TOC accumulation rates remain relatively low until about 1.5 ka BP and show a strong increase during the last 1.5 ka BP. During the last c. 0.4 ka BP, the highest accumulation rates of TOC, dinosterol and brassicasterol were recorded. CaCO₃ accumulation rates are low until 9.1 ka BP, followed by increased and maximum values from 9.1–7.6 and 5.9–3.0 ka BP and thereafter decrease during the last 3 ka, except for a minor rise during the last 0.4 ka BP.

Discussion

In order to investigate the palaeoceanographic changes along a N–S transect through eastern Baffin Bay to the Labrador Sea margin on the West Greenland Shelf, we have used a combination of sea ice, open-water phytoplankton productivity biomarkers and organic geochemical bulk parameters that are discussed together with relevant literature data. The records from this study show variability in sea-ice conditions along this north–south transect (Figs 5, 6) throughout the Holocene.

Deglacial to Early Holocene (11.5–7.6 ka BP)

Minimum occurrence or even absence of ice algae and phytoplankton biomarkers (cf. Saini et al. 2020) and related high P₁₁IP₂₅ indices (Fig. 5G) in core GeoB19927-3 in Baffin Bay suggest (eariy) spring conditions with extensive (spring) SIC prior to 7.6 ka BP. Primary productivity indicators in cores GeoB19927-3; AMD14-204C (Fig. 7E, F) show relatively low values in this period, suggesting a low productive environment (Limoges et al. 2020). Previous studies based on marine sediment cores from Baffin Bay have suggested that this region was controlled by cold deglacial conditions prior to c. 7.6 ka BP, especially during the initial part of the Early Holocene (Caron et al. 2019; Georgiadis et al. 2020; Hansen et al. 2020; Limoges et al. 2020). High P₁₁IP₂₅ index values (~0.9) prior to 7.6 ka BP from a nearby core AMD14-204C (Limoges et al. 2020) located in eastern Baffin Bay also suggest extended (early) spring sea-ice conditions (Fig. 5E). These generally harsh conditions were further supported by high IRD counts and extensive SIC (>10 months per year) as per reconstructions from eastern Baffin Bay (core CC70) in this period (Jennings et al. 2014; Gibb et al. 2015; Fig. 10A). Furthermore, extended sea-ice conditions (with generally reduced primary production; Fig. 7E, F) between ~8.8–7.6 ka BP in Baffin Bay (core GeoB19927-3; Saini et al. 2020) were reported and may be associated with the opening of Nares Strait and the inflow of cold Polar Water into Baffin Bay (Jennings 1993; Georgiadis et al. 2018; Jennings et al. 2019; Saini et al. 2020).

However, prior to ~9.1 ka BP, the surface and subsurface conditions in the NE Labrador Sea (core GeoB19905-1) are characterized by the absence of ice algae biomarkers as well as by extremely low accumulation rates of CaCO₃ and phytoplankton biomarkers (Fig. 8D, F, G). These (spring, summer and autumn) ice-free, albeit low primary productivity conditions might be related to the limited presence of warm Atlantic Water in the surface waters together with relatively high glacial meltwater discharge (Fig. 8C, H; Weiser et al. 2021) from the adjacent ice sheets. Such conditions might be attributed to the rapid retreat of the Greenland Ice Sheet (GIS) from the West Greenland Shelf (Andrews et al. 1999; Briner et al. 2016; Fig. 10A). Based on dinocyst reconstructions in the vicinity (core SA13-ST3), Allan et al. (2021) have reported generally low sea ice cover (<6 months per year) in the last 11 ka BP (Fig. 5B), possibly in winter months only. The occurrence of the dinocyst Pentapharsodinium dalei in core CC70 off Disko Bugt indicates large seasonal gradients from cold winters to mild summers and is possibly related to salinity fluctuations induced by meltwater discharge from the retreating GIS. This might indicate a weaker influence of the Atlantic Water (WGC) in the central Baffin Bay region (Gibb et al. 2015), in agreement with our reconstructions in the Labrador Sea (Figs 7, 8).

In the interval after ~9.1 ka BP, the NE Labrador Sea (core GeoB19905-1) is marked by an increase in the accumulation rates of CaCO₃ and the phytoplankton biomarker dinosterol as well as the absence of ice algae biomarkers (Fig. 4), all pointing towards increased marine productivity and ice-free conditions, which might correlate with the onset of the Holocene Thermal Maximum (HTM) conditions (Kaufman et al. 2004). Maximum occurrence of mixed benthic and planktic foraminifera specimens from the same core (GeoB19905-1) in this interval also points towards increased surface to subsurface productivity (Fig. 8). This might be also related to the disappearance of multi-year sea ice, which may have released an additional input of sympagic carbon into the water column further...
driving high marine productivity (Yunda-Guarin et al. 2020). This increase in primary productivity was most likely associated with a strong inflow of Atlantic Water and a decreased meltwater input (Fig. 8C, H; Weiser et al. 2021) and therefore, strengthened WGC influence from ~9.1 ka BP onwards (Fig. 10B) penetrating up to northern Baffin Bay (Lloyd et al. 2005; Saini et al. 2020).

Based on an increase in salinity and reduction in SIC from the outer shelf of Disko Bugt (core CC70), Jennings et al. (2014) and Gibb et al. (2015) suggested diminution of GIS meltwater input into the West Greenland areas. This rapid retreat of the GIS ice margin, thinning of ice sheets and spring–autumn ice-free conditions related to the strong oceanic forcing and high summer insolation in

Fig. 5. Comparison of different sea-ice records based on PIP$_25$ and dinocyst records along the eastern Baffin Bay–Labrador Sea (N–S) transect. A. Core GeoB19905-1 (this study). B. Core SA13-ST3 (Allan et al. 2021). C. Core MSM343310 (Kolling et al. 2018). D. Core MSM343300 (Ouellet-Bernier et al. 2014). E. F. Core AMD14-204C (Caron et al. 2019; Limoges et al. 2020). G. Core GeoB19927-3 (Saini et al. 2020). H. GeoB19948-3 (this study). I. J. Core AMD14-Kane2B (Caron et al. 2019; Georgiadis et al. 2020). Orange shading in PIP$_25$ records might be interpreted as ice-edge (IE?) situations (cf. Saini et al. 2020). The positions of these cores are indicated in Fig. 1. Calibrated AMS $^{14}$C ages of cores with varying age-resolution are shown as solid triangles for correct correlation of data.
the Early Holocene correspond well to the significant climate warming (Lloyd et al. 2005; Vinther et al. 2009; Renssen et al. 2012), widely recognized in the North Atlantic and western Arctic (Kaufman et al. 2004). Syring et al. (2020) and Müller & Stein (2014) also implied reduced SIC conditions (Fig. 9), as a result of strong inflow of warmer Atlantic Water in the Fram Strait area (Werner et al. 2016), in agreement with our reconstructions from the eastern margin of the Labrador Sea and Baffin Bay.

**Middle Holocene – transition to full interglacial conditions (7.6–3.0 ka BP)**

During the Middle Holocene ~7.6 ka BP, SIC started to decrease in Baffin Bay as shown by reduced values of P_{III}IP_{25} indices in cores GeoB19927-3 and GeoB19948-3 (Fig. 7D). This reduction in SIC was likely accompanied by strongly reduced meltwater input, causing increased surface salinity (Fig. 9), despite a decrease in subsurface Atlantic Water inflow (Fig. 8H; Weiser et al. 2021). In sum, this can be explained by assuming an increased influence of Atlantic Water on the surface water (Fig. 10B). This resulted in increased primary productivity as displayed by high peaks in productivity indicators (Fig. 8F, G). Short-term (cyclic) changes of minima and maxima in SIC are evident after ~7.6 ka BP in Baffin Bay, displayed by the P_{III}IP_{25} index (Fig. 5) based on Baffin Bay sediment cores GeoB19927-3 and GeoB19948-3. These reoccurring spring ice-edge conditions based on simultaneous peaks in accumulation rates of IP_{25} and the phytoplankton biomarkers brassicasterol and dinosterol (Fig. S2) are possibly related to oscillations in the ice melt influx and the influence of the WGC in the surface waters (Caron et al. 2019; Hansen et al. 2020). Additionally, the P_{III}IP_{25} index values along this north–south transect in eastern Baffin Bay (core GeoB19927-3, GeoB19948-3, AMD14-Kane2B; Fig. 5) indicate a prominent decline in SIC starting from ~7.6–6.3 ka BP. Based on dinoflagellate cyst data from core CC70, Gibb et al. (2015) reported an increased winter SST at ~7.6 ka BP associated with the strengthening of the warmer Atlantic Water influence that might further support our reconstruction. However, increased HBI III fluxes in nearby cores AMD14-Kane2B and AMD14-204C during this interval might be indicative of increased mobile sea ice and freshwater outflow via Nares Strait (Georgiadis et al. 2020; Limoges et al. 2020). Interestingly, dinocysts based sea-ice reconstructions from Disko Bugt (core MSM343310) in eastern Baffin Bay exhibit a similar reduction in SIC from 7.5–7.1 ka BP (cores GeoB19927-3, GeoB19948-3, AMD14-Kane2B). However, different reservoir ages (a few hundred to sometimes a thousand years) may limit the
exact age correlation (Fig. 5; Caron et al. 2019; Georgiadis et al. 2020; Hansen et al. 2020; Saini et al. 2020). On the other hand, the reduced SIC may have occurred earlier in the southern areas of the Baffin Bay transect (7.5–7.1 ka BP) before penetrating northwards up to the Kane Basin. Based on diatom and biomarker data from core AMD14-204C, Limoges et al. (2020) also suggested decreased sea-ice conditions with increased primary production, possibly due to increased WGC influence during this interval. The decreased SIC (Fig. 7D) after ~7.6 ka BP, together with high Atlantic Water influence associated with a decreased inflow of meltwater, might suggest significant warming of the subsurface conditions corresponding to the HTM-like conditions observed in NE Baffin Bay throughout the Middle Holocene (Lloyd et al. 2005; Gibb et al. 2015; Fig. 10B). Several records from circum Greenland, the Canadian Arctic Archipelago, and Svalbard areas also suggested a similar increased influence of Atlantic Water (WGC, IC) in the Middle Holocene since ~7.6 ka BP (Dyke et al. 1996; Justwan et al. 2008; Jennings et al. 2011; Ouellet-Bernier et al. 2014). Note, in contrast to sea-ice records from the Fram Strait area (cores MSM5/5-712 and PS93/025), which seem to follow predominantly the summer insolation trend (Müller et al. 2012; Syring et al. 2020), our records from eastern Baffin Bay indicate a dominant influence of meltwater influx on sea-ice formation throughout the Holocene (Fig. 9).

Furthermore, a general northwards increase in SIC shown by the PIP25 indices along the studied N–S transect (Figs 5, 6) might be related to the fading Atlantic Water inflow (WGC) northwards and/or due to interaction with southwards flowing Polar Currents (PC) along the Baffin Islands (Canadian Arctic). This trend may also be related to the seasonal differences in light and sea-ice conditions as a function of latitude as the timing and extent of the ice and phytoplankton productivity vary from April to late summer from the south towards higher latitudes (cf. Wassmann et al. 2020 for a detailed explanation).
For the Middle Holocene between 5.9–3.0 ka BP, i.e. after the hiatus, our records from core GeoB19905-1 off NE Labrador Sea show ice-free conditions in spring–autumn as displayed by the continued absence of ice algae biomarkers (IP_{25} = 0; Fig. 4J). Dinosterol accumulation rates show a slight increase in this interval (Fig. 8F); however, CaCO_{3} accumulation rates remained rather high until about 4 ka BP (Fig. 8F, G), suggesting subsurface conditions characterized by high biological (carbonate) productivity. This interpretation is supported by the continuously maximum occurrence of mixed planktic and benthic foraminifera specimens in core GeoB19905-1 (J. Weiser, pers. comm. 2021), which might correlate with the HTM-like conditions in the Middle Holocene (cf. Discussion for more explanation). I to III represent different environmental periods shown in the maps of Fig. 10, based on our proxy records.

Fig. 8. Comparison of selected proxies for sea ice with productivity, WGC current strengths in the NE Labrador Sea and/or inflow of warm Atlantic Water (ATW) as well as meltwater influence in the area. A. Summer insolation at 70°N (Laskar et al. 2004). B. Camp Century δ^{18}O record (Vinther et al. 2009). C. Core GeoB19905-1 fine grain size mode AR (EM1) accumulation rate (Weiser et al. 2021). D. Core GeoB19905-1 sea ice (IP_{25} = 0). E. Core GeoB19905-1 dinosterol (µg g^{-1} TOC). F, G. Core GeoB19905-1 productivity (dinosterol and CaCO_{3}; accumulation rates). H. Core GeoB19905-1, SS (mean sortable silt) (Weiser et al. 2021). I. Qualitative assessment of the ATW inflow into the NE Labrador Sea. The green bar indicates the maximum occurrence of mixed planktic and benthic foraminifera specimens in core GeoB19905-1 (J. Weiser, pers. comm. 2021), which might correlate with the HTM-like conditions in the Middle Holocene (cf. Discussion for more explanation).
decrease in the meltwater discharge along the SW Greenland margin in this interval (Holland et al. 2008; Ren et al. 2009; Gibb et al. 2015) may have substantially increased subsurface salinity, thus maintaining a strong influence of Atlantic Water (WGC) to the ocean surface. Biomarker proxies, including sediment composition (grain-size) data, have already been previously successfully applied for subsurface and bottom water conditions in the High Northern Latitude areas, i.e. Denmark Strait and Labrador Sea (Andrews et al. 2020; Weiser et al. 2021).

Towards the final stage of the Middle Holocene (~4–3 ka BP) at core site GeoB19905-1 in the Labrador Sea, the accumulation rates of CaCO3 display a decrease whereas dinosterol accumulation rates remain unchanged (Fig. 8F, G). This suggests changes in subsurface conditions characterized by either a decrease in biological (carbonate) productivity and/or a decrease in preservation of organic carbon (i.e. CaCO3 and TOC). CaCO3 has been interpreted as productivity indicator very selectively as in core GeoB19905-1. J. Weiser (pers. comm. 2021) has analysed the foraminifera data showing well-preserved, mixed planktic and benthic specimens, coinciding with the increased CaCO3 accumulation rates. This might suggest that the CaCO3 is mainly of biogenic origin; hence, its accumulation rates could be used to infer surface and subsurface productivity. Here, this decrease in biological (carbonate) productivity could be linked to a decline in the WGC strengths, while the high dinosterol accumulation rates might be due to the increased deposition of fine-grained sediments (Fig. 8H; Weiser et al. 2021). The declined WGC strength temporally coincides with the onset of Neoglacial cooling, widely reported around Greenland areas (Levac et al. 2001; Long & Roberts 2003; Briner et al. 2010; Müller et al. 2012; Perner et al. 2012; Krawczyk et al. 2017; Schweinsberg et al. 2017). Dinocyst-based reconstructions in the nearby core SA13-ST3 also indicate a slight decrease in summer primary productivity in this interval (Allan et al. 2021), in agreement with our study. Moros et al. (2004) also argued for a period (4–3 ka BP) of climate instability and significant fluctuation in EGC and IC strengths in the East Greenland margin leading to cooling. Several lake sediment records in the vicinity of Southwest Greenland also showed renewed ice growth and enhanced glacier activity starting c. 4 ka BP (Larsen et al. 2021).
et al. 2017), supporting the onset of Neoglacial cooling in this time interval.

Late Holocene changes in sea-surface conditions (3–0 ka BP)

Surface water cooling along the West Greenland coast during the Late Holocene Neoglacial period is attributed to the variability in the strength of the Atlantic (IC) vs. Arctic (EGC, BC) currents (Moros et al. 2006; Seidenkrantz et al. 2007; Andersen et al. 2011; Ouellet-Bernier et al. 2014). Interestingly, our sea ice (IP_{25}) and phytoplankton biomarkers (dinosterol, HBI III), and related P_{IP_{25}} indices display reduced values (Figs 3, S2), and low accumulation rates of all biomarkers during this period at core sites GeoB19927-3 and GeoB19948-3, which may be attributed to stratified conditions associated with an enhanced input of cold corrosive Polar Water (Fig. 10C) hampering productivity. Based on diatom and biomarker data from core AMD14-204C, Limoges et al. (2020) point towards major oceanographic reorganization that may be related to the strengthening of the EGC and/or decreased contribution of warm IC into the WGC (Flatau et al. 2003; Morley et al. 2014). Based on the increased abundance of agglutinated foraminifera species in cores AMD14-204C and MSM43300, Hansen et al. (2020) and Perner et al. (2012) suggested enhanced Arctic Water influx into Baffin Bay and strongly stratified conditions during the Late Holocene. Based on diatom assemblages from nearby core AMD14-204C, Limoges et al. (2020) also reported enhanced drift/pack-ice in this interval, which might have contributed to increased multi-year sea-ice (IP_{25}-poor) presence and decreased primary production in this area. High peaks in the HBI III records from cores GeoB19927-3, GeoB19948-3 and AMD14-204C (Figs 3, S2) in this interval suggest an extended sea ice cover, which further indicates increased (multi-year) sea ice in Baffin Bay imported via Nares Strait (Fig. 10C; Limoges et al. 2020; Saini et al. 2020). This is further supported by a decrease in winter snowfall in western Greenland areas (Thomas et al. 2016), implying a more extensive SIC that might have limited moisture availability during this interval. Afterwards, low to moderate P_{IP_{25}} values (0.2–0.6) between 1.8 and 1.2 ka BP at core sites GeoB19927-3 and GeoB19948-3 (Fig. 5) might indicate a short interval of reduced sea ice cover associated with a positive NAO mode and related warmer subsurface conditions in eastern Baffin Bay (Moros et al. 2006; Trouet et al. 2009; Andersen et al. 2017).
Increased SSTSs were also suggested near the East Greenland shelves at c. 1.8–1.2 ka BP (Knudsen et al. 2004; Roncaglia & Kuippers 2004). The reduction in SIC might be also attributed to strong sea-ice interactions with the local fjords, which can also hamper ice growth (Ribeiro et al. 2017) as a result of poor light availability due to higher snow conditions as well as low salinity levels resulting from high local meltwater runoff. The short-term warming in the North Atlantic and the in-phase relationship between sea ice and NAO mode have been previously reported (Trouet et al. 2009; Ljungqvist 2010; Limoges et al. 2020; Saini et al. 2020). An increasing trend in SIC is more clearly evidenced after about 1.4 ka BP as shown by PIP25 indices in cores GeoB19927-3, GeoB19948-3 and AMD14-204C (Limoges et al. 2020; Fig. 5), coinciding with Neoglacial cooling observed elsewhere (Fig. 10C and references therein). Based on the increase in Arctic Water indicator dinocysts (Islandinium minutum; core AMD14-204C) during the last ~1.4 ka, Caron et al. (2019) have described an increase in SIC and colder conditions in eastern Baffin Bay related to Neoglacial cooling. Based on IP25 and diatom records from core AMD14-204C, Limoges et al. (2020) also suggested an increased SIC during the last 1.4 ka BP and colder Neoglacial conditions. This is further supported by enhanced driftwood findings during the Late Holocene in the Canadian Arctic Archipelago, possibly carried along by the strengthened Polar Water (i.e. BC) masses (Dyke et al. 1999). This increase in SIC during the Neoglacial period is in agreement with the reconstructions from the northern Baffin Bay (Levac et al. 2001) and the eastern Fram Strait area (Fig. 9) (Müller & Stein 2014; Syring et al. 2020).

However, in the NE Labrador Sea, after 3 ka BP, HTM-like conditions were followed by surface and subsurface conditions characterized by the absence of ice algae biomarkers and significantly decreased CaCO3 accumulation rates (Fig. 4E, J), albeit very high accumulation of the phytoplankton biomarker dinosterol (Fig. 8F; core GeoB19905-1). Based on the end-member grain-size analysis of the same core (GeoB19905-1), Weiser et al. (2021) documented a strong increase in the accumulation of poorly-sorted fine-grained sediments (Fig. 8C) presumably from the Neoglacial ice-margin advances. Due to increased flocculation/coagulation of fine-grained siliciclastic material (‘ballast effect’; cf. Ittekkot et al. 1992; Knies & Stein 1998; Fahl & Stein 2007; Iversen & Robert 2015), organic matter may become enriched in such finer-grained sediments as shown by elevated TOC values and high accumulation rates of organic carbon (Fig. 4). This may have resulted in more labile organic matter preservation in the sediments. Another explanation for the increased accumulation rates of the phytoplankton biomarkers in the Late Holocene (Fig. 8F) might be the increased nutrient supply (e.g. Fe, Si) associated with the enhanced local meltwater discharge (Bhatia et al. 2013; Hawkins et al. 2015; Arrigo et al. 2017; Cape et al. 2019). Based on exploration studies near the Sinarsuk deposit, Secher (1980) and Grammatikopoulos et al. (2002) found a significant amount of magnetite in the host rock, which might be a potential source of high Fe contents recorded in the Late Holocene section of the core GeoB19905-1 (Weiser et al. 2021).

The last 0.4 ka (upper ~100 cm, core GeoB19905-1, Fig. 4) are characterized by very high phytoplankton biomarker accumulation rates that might be explained on one hand by increased phytoplankton productivity/ increased preservation. On the other hand, the topmost maximum value and its down-core decrease may represent diagenetic alterations in the uppermost centimetres (cf. Fahl & Stein 2012; Belt & Müller 2013). The drop in phytoplankton biomarkers (Fig. 8F, G) evident at about 0.3 ka BP (~AD 1700) may be correlated with the Little Ice Age (LIA), widely reported in the Western Europe and North Atlantic regions (Jones & Mann 2004; Ljungqvist 2010; Spielhagen et al. 2011). However, more high-resolution reconstructions are needed to resolve the decadal to centennial-scale climate events.

Conclusions

Organic geochemical and biomarker investigation of AMS 14C-dated sediment cores (GeoB19948-3, GeoB19927-3, GeoB19905-1) from a N-S transect of the eastern Baffin Bay-Labrador Sea margin was used for environmental and palaeoceanographic reconstruction covering the Holocene period. Our data based on cores GeoB19948-3 and GeoB19927-3 suggest that Baffin Bay was seasonally covered by sea ice during the last ~10.1 ka. In contrast, predominantly ice-free (spring and/or autumn) conditions were observed during the last 11.5 ka in the NE Labrador Sea (core GeoB19905-1). We record a general decrease in SIC conditions along the N–S transect throughout the Holocene.

Extended (early) spring SIC conditions occurred prior to ~7.6 ka BP in the Early Holocene in eastern Baffin Bay, possibly linked to the limited subsurface presence of Atlantic Water (WGC). However, in the NE Labrador Sea, conditions remained mostly ice-free in spring–autumn in the Early Holocene due to the increasing but limited influence of Atlantic Water (WGC).

A transition towards decreased SIC and reoccurring ice-edge conditions is evident in the Middle Holocene (~7.6–3 ka BP), possibly caused by the short-term (cyclic) changes in the WGC influence associated with ice melting and can be characterized by HTM-like conditions. This decrease in SIC is synchronous with the decrease in the meltwater inflow leading to the salinization of surface waters in the Baffin Bay area. The HTM-like conditions in the Middle Holocene, in response to the final retreat of the LIS and GIS, are also recorded in the NE Labrador Sea, displayed by spring–autumn ice-free (PIP25 = 0) conditions together with an increased accumulation of productivity indicators.
Our combined sea-ice records from Baffin Bay indicate low in-situ sea-ice (IP25) algae production; however, enhanced drifted (multi-year) sea ice during the Late Holocene (last 3 ka BP) associated with the enhanced influx of Polar Water masses might have caused stratified conditions resulting in reduced primary productivity related to the Neoglacial cooling trend observed elsewhere. However, in the NE Labrador Sea, surface conditions were characterized by (spring–autumn) ice-free conditions with a strong decrease in carbonate productivity during the last 4 ka, in line with decreased WGC strength and/or enhanced advection of cold Arctic Water masses and increased meltwater inflow linked to the Neoglacial cooling. Our sea-ice records from the Baffin Bay-Labrador Sea on the West Greenland side indicate that meltwater input has mainly controlled sea-ice formation during the Holocene, in contrast to the East Greenland side (i.e. Fram Strait), where the Holocene sea-ice history follows predominantly the summer insolation trend.

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Supporting Information

Additional Supporting Information to this article is available at http://www.boreas.dk.

**Fig. S1.** Comparison of GeoB19927-3 age-model determined by BACON (Saini *et al.* 2020) with the updated version using PDV (Langner & Mulitza 2019), which uses modelled reservoir ages (Butzin *et al.* 2017) and BACON. Note, the updated age-model (shown in brown) used in this study is about 0.2 ka younger than previously shown by Saini *et al.* (2020).

**Fig. S2.** Down-core variations of core GeoB19927-3 based on Saini *et al.* (2020): (a) bulk accumulation rate, and accumulation rates of (b) total organic carbon, (c) CaCO3, (d) Z-HBI III, (e) dinosterol, (f) brassicasterol, and (g) sea-ice proxy IP25. The orange box marks the ice-edge (IE) positions. All plots are shown vs. age in 1000 years before present (ka BP).

**Fig. S3.** Combined record of core GeoB19948-3: (a) CaCO3 content, (b) total organic carbon (TOC), (c) HBI III (µg/g), (d) dinosterol (µg/g), (e) brassicasterol (µg/g), and (f) sea-ice proxy IP25. (g) Sea-ice proxy IP25.

**Fig. S4.** Combined record of core GeoB19905-1: (a) CaCO3 content, (b) total organic carbon (TOC), (c) dinosterol (µg/g), (d) brassicasterol (µg/g), and (e) sea-ice proxy IP25.