



## Holocene cooling culminates in sea ice oscillations in Fram Strait

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### ABSTRACT

A reconstruction of Holocene sea ice conditions in the Fram Strait provides insight into the palaeoenvironmental and palaeoceanographic development of this climate sensitive area during the past 8500 years BP. Organic geochemical analyses of sediment cores from eastern and western Fram Strait enable the identification of variations in the ice coverage that can be linked to changes in the oceanic (and atmospheric) circulation system. By means of the sea ice proxy IP<sub>25</sub>, phytoplankton-derived biomarkers and ice rafted detritus (IRD) increasing sea ice occurrences are traced along the western continental margin of Spitsbergen throughout the Holocene, which supports previous palaeoenvironmental reconstructions that document a general cooling. A further significant ice advance during the Neoglacial is accompanied by distinct sea ice fluctuations, which point to short-term perturbations in either the Atlantic Water advection or Arctic Water outflow at this site. At the continental shelf of East Greenland, the general Holocene cooling, however, seems to be less pronounced and sea ice conditions remained rather stable. Here, a major Neoglacial increase in sea ice coverage did not occur before 1000 years BP. Phytoplankton-IP<sub>25</sub> indices (“PIP<sub>25</sub>-Index”) are used for more explicit sea ice estimates and display a Mid Holocene shift from a minor sea ice coverage to stable ice margin conditions in eastern Fram Strait, while the inner East Greenland shelf experienced less severe to marginal sea ice occurrences throughout the entire Holocene.

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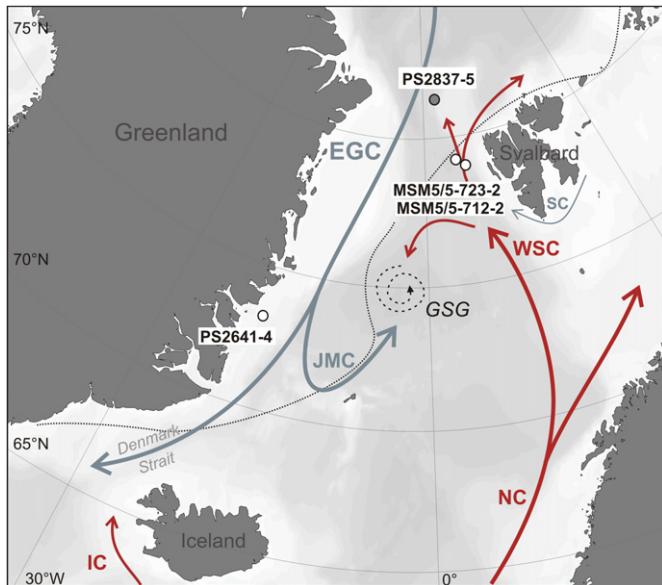
### 1. Introduction

The extent of sea ice coverage in Fram Strait, the major gateway connecting the Arctic with the Atlantic Ocean, is intrinsically tied to the advection of warm Atlantic Water along the continental margin of West Spitsbergen. As these temperate waters head to the north, they encounter polar water (and air) and sea ice from the Arctic Ocean, which causes cooling, freezing and thus brine rejection, and subsequent descent of Atlantic Water into the Nordic Sea's deep sea basins via the Greenland Sea Gyre (Fig. 1; Aagaard, 1982; Rudels and Quadfasel, 1991; Watson et al., 1999). These processes are of crucial importance for the so-called Nordic heat pump, which bestows a comparatively temperate climate upon Europe (e.g. Broecker, 1992). The climate-shaping impact of sea ice that exits Fram Strait became particularly evident during the “Great Salinity Anomaly” in the 1970s, when an enormous discharge of Arctic sea ice hampered the thermohaline convective overturn in the North Atlantic (in

terms of a vast low salinity freshwater lense), which resulted in a significant cooling in the North Atlantic area (Dickson et al., 1988; Belkin et al., 1998; Dima and Lohmann, 2007; Sundby and Drinkwater, 2007). Recently, Spielhagen et al. (2011) identified and linked natural fluctuations in the advection of Atlantic Water towards Fram Strait with shifting warm and cold climate intervals like the Medieval Climate Anomaly or the Little Ice Age. Furthermore, an unprecedented warming of North Atlantic Water throughout the past 120 years is reconstructed that highlights the importance of the direct feedback mechanisms between the atmospheric (global) warming, the oceanic heat transport through Fram Strait and the sea ice decline in the Arctic realm (Spielhagen et al., 2011). These dynamic interactions account for the Arctic amplification, which impacts not only on the Arctic Ocean but also on adjacent terrestrial (permafrost) areas and finally the global climate system (Lawrence et al., 2008; Serreze and Barry, 2011). Overland and Wang (2010), for example, put emphasis on the loss of Arctic sea ice and the resulting changes in large-scale atmospheric circulation patterns and the consequences for mid-latitude weather (wind) regimes.

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**Fig. 1.** Oceanographic setting and location of core sites in the study area. Red arrows refer to warm Atlantic Water carried by the Norwegian Current (NC), the Irminger Current (IC), and the West Spitsbergen Current (WSC). Blue arrows refer to polar water and sea ice carried by the East Greenland Current (EGC), the Jan Mayen Current (JMC), and the minor Sørkapp Current (SC). The Greenland Sea Gyre (GSG) and the modern winter sea ice margin (dotted line) are indicated as well. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

The finding of past variations in the sea ice distribution in Fram Strait thus supports the identification of palaeo-fluctuations in the intensity of Atlantic Water inflow and may reveal periods of a strengthened or weakened thermohaline circulation and/or atmospheric (North Atlantic Oscillation; NAO-like) forcing. The influence of the NAO on climate and sea ice conditions in the (sub) Arctic realm frequently has been appreciated as fundamental, though hardly assessable or predictable due to its highly variable temporal evolution (e.g. Dickson et al., 2000; Hurrell and Deser, 2010). In short, positive NAO phases are accompanied by stronger westerlies carrying moist air over Europe and Siberia, an increased Atlantic Water inflow through Fram Strait, and warmer temperatures in the Arctic, which lead to a reduction in sea ice formation. During intervals of a negative NAO these phenomena occur to be reversed (Dickson et al., 2000; Kwok, 2000; Hurrell and Deser, 2010). Within their thorough review about Arctic sea ice and its interaction with the atmosphere Bader et al. (2011) illustrate comprehensively how the current sea ice reduction leads to a poleward shift and an intensification of storm tracks, while the immediate impact on the NAO itself remains undetermined. Vice versa, the distinct impact of the NAO on the sea ice extent (e.g. in the Nordic Seas) has been acknowledged and documented more often (Deser et al., 2000; Dickson et al., 2000; Vinje, 2001). Given the lack of instrumental records of the NAO variability prior to 1932 (the first calculation of the NAO index dates back to 1932; Walker and Bliss, 1932) the attempts to link atmospheric fluctuations with climate changes are restricted to proxy reconstructions or numerical model experiments (e.g. Rimbu et al., 2004; Lorenz et al., 2006; Trouet et al., 2012). For instance, an Early to Late Holocene decrease in North Atlantic SSTs is interpreted to reflect a general long-term weakening of the NAO-like atmospheric circulation pattern (Rimbu et al., 2003). On shorter time scales, however, it seems essential to distinguish between the intensity and the frequency of cyclones in the North Atlantic to reasonably relate proxy data to Late Holocene NAO shifts (Trouet et al., 2012).

Though Northern Hemisphere climate (boundary) conditions throughout the Holocene are generally considered as fairly stable (Grootes and Stuiver, 1997), variations in sea surface temperatures (SSTs), glacier growth or terrestrial vegetation communities are increasingly substantiated within marine and terrestrial Arctic palaeoclimate studies (Birks, 1991; Svendsen and Mangerud, 1997; Andersen et al., 2004; for recent review see; Müller et al., 2010). Recently, Risebrobakken et al. (2011) demonstrated reasonably that, when interpreting marine proxy derived climate information (e.g. SSTs) in the Nordic Seas, the individual impacts of orbital forcing (mainly affecting sea surface conditions) and oceanic heat advection (affecting deeper parts of the ocean and convective processes) requires careful consideration as these are different mechanisms of climate change. Thus, the partly contradictory Holocene SST reconstructions in the Nordic Seas, which are based on coccolithophore-derived alkenone or foraminifer data (Calvo et al., 2002; Risebrobakken et al., 2003) can be explained by the simple fact that different proxies may respond to different mechanisms (Risebrobakken et al., 2011).

Concerning sea ice conditions, Holocene changes in the ice coverage in the Nordic Seas, however, have been deduced mainly indirectly from microfossil or geochemical data (Andrews et al., 2001; Jennings et al., 2002; Bonnet et al., 2010). A quantitative approach using diatom transfer functions in the Nordic Seas has been presented by Justwan and Koç (2008). By means of a sediment core north off Iceland, they reconstruct relatively constant sea ice concentrations of ca 5%–10% for the Early Holocene and slightly higher sea ice concentrations of about 10%–20% during the Late Holocene (Justwan and Koç, 2008). The application of this promising approach, however, may be limited by the comparatively high silica dissolution rate in the High Northern Latitudes (Kohly, 1998; Schlüter and Sauter, 2000).

The molecular sea ice proxy IP<sub>25</sub> – a biomarker lipid associated with sea ice diatoms – seems to be a direct and thus valuable tool for the reconstruction of a previous spring sea ice cover in the Arctic (Belt et al., 2007; Brown, 2011). Besides the identification of highly branched C<sub>25</sub> and C<sub>30</sub> isoprenoids as diatom specific biomarkers (e.g. Rowland and Robson, 1990; Volkman et al., 1994; Massé et al., 2004) that even may be found in Cretaceous sediments (Damsté et al., 2004), the derivation of the monounsaturated C<sub>25</sub> highly branched isoprenoid (i.e. the IP<sub>25</sub> alkene) from diatoms that live within the Arctic sea ice has been strengthened in various studies (Belt et al., 2008; Brown et al., 2011; Brown and Belt, 2012). With regard to this distinct association of IP<sub>25</sub> with sea ice, the detection even of trace abundances of this molecule in a sediment sample, which indeed is a question of instrumental sensitivity, may directly serve as an indication of a previous ice cover. The increasing use of IP<sub>25</sub> for palaeo sea ice assessments and its agreement with other proxy (Massé et al., 2008; Müller et al., 2009; Vare et al., 2009, 2010; Belt et al., 2010) and instrumental data (Müller et al., 2011) on sea ice occurrences hence supports the applicability of this biomarker.

In 2009, Vare et al. and Müller et al. presented reconstructions of sea ice conditions based on the IP<sub>25</sub> content in sediment cores from the central Canadian Archipelago and northern Fram Strait, respectively, which cover the entire Holocene. Both studies suggest gradually increasing (spring) sea ice occurrences from the Mid to the Late Holocene, presumably as a response to the Neoglacial cooling (Müller et al., 2009), but do not provide an in-depth analysis of the palaeoenvironmental and palaeoceanographic setting. The Neoglaciation – the general use of this term was first suggested by Porter and Denton (1967) – covers the period characterised by glacier advances, southward migration of the northern treeline and colder sea surface conditions in different regions of the Northern Hemisphere that followed the warm Early to Mid Holocene (for

further review see Wanner et al., 2008; Miller et al., 2010 and references therein).

The main objective of this study is to estimate to what extent this Holocene cooling affected the sea ice distribution in the Fram Strait and the East Greenland shelf. For this purpose, organic geochemical and IRD analyses were performed on sediment cores from the western continental margin of Spitsbergen and the continental shelf of East Greenland. This provides for a reconstruction of the spatial and temporal evolution of the sea ice coverage within the two most important oceanic (in and outlet) pathways that characterise the Fram Strait and influence the Arctic Ocean heat budget. The findings are compared and contextualised with previous palaeoenvironmental reconstructions for the study area.

## 2. Regional setting

The environmental setting in Fram Strait is controlled by a dynamic ocean current system and, owing to the high latitude, a distinct seasonality. Warm and saline Atlantic Water is directed northwards towards Fram Strait by the Norwegian Current (NC) and the West Spitsbergen Current (WSC), thus constituting the northernmost area of open (ice-free) water in the Arctic during winter (Fig. 1; Vinje, 1977; Aagaard, 1982). South of Spitsbergen these temperate waters encounter cold water and sea ice, which is carried by the minor Sørkapp Current (SC) from the Barents Sea along the southern tip and west coast of Spitsbergen (Swerpel, 1985). Further to the north at about 79°N, the WSC splits in two current systems, with an eastern (Svalbard) branch flowing along the northeastern shelf of Spitsbergen and a western (Yermak) branch following the western flank of the Yermak Plateau where it is partly recirculated southward (Fig. 1; Bourke et al., 1988).

The western part of Fram Strait experiences a huge discharge of polar water and sea ice that originates from the Arctic Ocean (i.e. predominantly from the East Siberian and the Laptev Sea) and is exported along the continental shelf of East Greenland by the East Greenland Current (EGC; Aagaard and Coachman, 1968; Rudels et al., 1999). Currently, only the proximal (inner) shelf of East Greenland and northern Fram Strait remain ice-covered until early summer (NSIDC, Boulder, USA). During periods of extremely cold winter (and spring) months with severe temperature and sea ice conditions in the Arctic, the ice flux may extend towards the east (and south), such that also the eastern part of Fram Strait experiences an intensified (drift) sea ice coverage (for association with NAO variability see Dickson et al., 2000; Vinje, 2001). This is also substantiated through IRD studies of sediment trap material from the continental slope of West Spitsbergen by Hebbeln (2000), who shows that fine-grained lithic material may be released from sea ice originating not only from Spitsbergen but also from the north (i.e. the Arctic Ocean).

## 3. Sediment material and methodology

The sediment cores MSM5/5-712-2 and MSM5/5-723-2 were obtained from the western continental margin of Spitsbergen (at 78°54.94 N, 6°46.03 E; 1487 m water depth, and at 79°09.66 N, 5°20.27 E; 1349 m water depth, respectively) during a *Maria S. Merian* cruise in 2007 (Budéus, 2007). The core sites are both located in close vicinity to the modern winter sea ice margin (Fig. 1). Sediment cores were stored at -30 °C until further treatment. For organic geochemical analyses subsamples were taken each cm, freeze-dried and homogenised. Sedimentary total organic carbon (TOC) contents were determined by means of a carbon-sulfur determinator (CS-125, Leco) after the removal of carbonates by adding hydrochloric acid. Total carbon (TC) contents measured by a CNS analyser (Elementar III, Vario) were used to calculate

carbonate contents ( $\text{CaCO}_3 = (\text{TC} - \text{TOC}) \times 8.333$ ). Core MSM5/5-712-2 was further studied for ice rafted detritus (IRD). Lithic grains of freeze-dried subsamples were counted on a representative split (>100 grains) in the 150–250 µm size fraction. Further IRD grain size and mineralogy analyses of these cores are subject of a forthcoming study.

Sediment core PS2641-4 from the East Greenland shelf (73°9.3 N, 19°28.9 W; 469 m water depth) was obtained during *Polarstern* cruise ARK-X-2 (Hubberten, 1995). TC and TOC (and thus also carbonate) contents of freeze-dried and homogenised subsamples (5–10 cm sampling intervals) from this core were determined by means of a Heraeus CHN-O-Rapid Elementar Analyser. We note that the freeze-dried sediment material was stored at room temperature for ca 15 years before it was analysed for its biomarker composition. This probably promoted some chemical alteration of the organic matter, which needs to be considered when looking at the absolute concentration profiles of the biomarkers.

For lipid biomarker analyses ca 1–4 g of sediment were extracted by an Accelerated Solvent Extractor (DIONEX, ASE 200; 100 °C, 5 min, 1000 psi) using dichloromethane:methanol (2:1 v/v). Prior to this step, 7-hexylnonadecane, squalane and cholesterol-d6 (cholest-5-en-3β-ol-D6) were added as internal standards for quantification purposes. Further separation of alkanes and sterols was carried out via open-column chromatography using SiO<sub>2</sub> as the stationary phase. Hydrocarbons were eluted with *n*-hexane (5 ml) and sterols with methylacetate:*n*-hexane (20:80 v/v; 6 ml). The latter were silylated with 500 µl BSTFA (60 °C, 2 h). Compound analyses of both fractions were carried out on an Agilent 6850 GC (30 m HP-5MS column, 0.25 mm i.d., 0.25 µm film thickness) coupled to an Agilent 5975 C VL mass selective detector. The GC oven was heated from 60 °C to 150 °C at 15 °C min<sup>-1</sup>, and then at 10 °C min<sup>-1</sup> to 320 °C (held 15 min) for the analysis of hydrocarbons and at 3 °C min<sup>-1</sup> to 320 °C (held 20 min) for sterols, respectively. Operating conditions for the mass spectrometer were 70 eV and 230 °C (ion source). Helium was used as carrier gas. The identification of individual biomarkers is based upon comparison of their retention times and mass spectra with published data (Boon et al., 1979; Volkman, 1986; Belt et al., 2007). Biomarker concentrations were calculated on the basis of their individual GC–MS ion responses compared with those of respective internal standards. For the quantification of IP<sub>25</sub> via its molecular ion a calibration factor was considered that was obtained from calibration experiments using a sediment with known IP<sub>25</sub> concentration, which thus serves as a substitute for synthetically produced IP<sub>25</sub>. The IP<sub>25</sub> concentration of this sediment has been verified through GC analyses. Aliquots of the hydrocarbon fraction of this sediment were then used for GC–MS calibration experiments (using a serial dilution), which rest upon correlations of different (quantified via GC) IP<sub>25</sub> concentrations obtained from total ion current analyses with the respective IP<sub>25</sub> concentrations determined by selected ion monitoring analyses (m/z 350 for IP<sub>25</sub> and m/z 266 for 7-hexylnonadecane). Within this study hydrocarbon fractions from sediment core PS2837-5 (Müller et al., 2009) were re-analysed and their IP<sub>25</sub> contents accordingly calibrated, which eases the comparison with the results obtained from core PS2641-4 and the *Maria S. Merian* cores. The herein presented PIP<sub>25</sub> indices are calculated following the equation by Müller et al. (2011), where concentration balance factors are used to account for the disparity between the generally low concentrations of IP<sub>25</sub> and the high concentrations (due to multiple phytoplankton sources) of phytoplankton biomarkers in the sediments. Accordingly, the accumulation rates of IP<sub>25</sub>, brassicasterol, and dinosterol were averaged over the whole core sections and the mean values were then used to calculate the balance factors for the respective cores (PIP<sub>25</sub>-equations with individual balance factors are also given in Fig. 8).

#### 4. Core chronologies

The chronology of the sediment cores MSM5/5-712-2 and MSM5/5-723-2 is based upon AMS  $^{14}\text{C}$  ages obtained from tests of the polar planktic foraminifer *Neogloboquadrina pachyderma* (sin.), whereas the age model of the sediment core PS2641-4 is based upon AMS  $^{14}\text{C}$  ages that were obtained from tests of benthic foraminifera. Additionally, AMS  $^{14}\text{C}$  ages were derived from shells of the bivalve *Bathyrca glacialis* (Evans et al., 2002). For the *Maria S. Merian* cores a marine reservoir correction of 408 years has been assumed to convert radiocarbon ages into calibrated calendar years before present (cal years BP) using the calibration software CALIB 6 (see Table 1; Stuiver and Reimer, 1993; Stuiver et al., 1998; updated to CALIB 6.0 by Stuiver et al., 2005; see CALIB at <http://calib.qub.ac.uk/>). A reservoir age of 550 years has been assumed for the correction of radiocarbon ages of the PS2641-4 core according to Hjort (1973). For the age model of this core we omitted one  $^{14}\text{C}$  age at 90.5 cm ( $1705 \pm 110 = 1091$  cal years BP) because this sample contained only a very little amount of carbon (0.06 mg). Furthermore, this dating would imply an enormous sedimentation rate ( $>500$  cm/1000 years) compared to the adjacent intervals ( $<50$  cm/1000 years). Such an “event” of extreme sediment deposition, however, cannot be identified in the sediment structures of the respective core section. Anticipating linear sedimentation rates at the core sites, ages of sediment intervals between  $^{14}\text{C}$ -dated horizons are based on linear interpolation (Fig. 2). Mass accumulation rates ( $\text{g}/\text{cm}^2/1000$  years) were calculated on the base of these sedimentation rates, density and porosity data (Evans, 2000), and were finally used to convert absolute sedimentary biomarker contents into flux rates.

**Table 1**

AMS radiocarbon ages for *Maria S. Merian* cores obtained from tests of the planktic foraminifer *Neogloboquadrina pachyderma* sin. For these dates a marine reservoir age of 408 years has been assumed according to Hughen et al. (2004). AMS radiocarbon ages for the *Polarstern* core were obtained from tests of mixed benthic foraminifera. In addition, two AMS  $^{14}\text{C}$  dates (labelled with a star) determined in shells of the bivalve *Bathyrca glacialis* (Evans et al., 2002) were used. For radiocarbon ages of this core a reservoir age of 550 years has been assumed according to Hjort (1973). The age obtained from benthic foraminifera at 90.5 cm sediment depth in PS2641-4 has been ignored for the calculation of the age model. Superscript numbers in the lab reference indicate  $^{14}\text{C}$  dates provided by Robert Spielhagen (1), Jacques Giraudeau (2), Christian Hass (3).

Sediment core	Lab reference	Core depth (cm)	AMS $^{14}\text{C}$ age	Calibrated age BP (2 $\sigma$ )	
MSM5/5-712-2	KIA 45217 <sup>1</sup>	11	815 $\pm$ 25	459 $\pm$ 49.5	
	KIA 41024 <sup>1</sup>	21	1570 $\pm$ 25	1130 $\pm$ 82.5	
	KIA 45218 <sup>1</sup>	28	1985 $\pm$ 25	1544 $\pm$ 96.5	
	KIA 45219 <sup>1</sup>	41	2565 $\pm$ 25	2242 $\pm$ 83	
	SacA 19113 <sup>2</sup>	60.5	3365 $\pm$ 30	3240 $\pm$ 97.5	
	SacA 19114 <sup>2</sup>	94.5	4915 $\pm$ 30	5256 $\pm$ 104	
	SacA 19115 <sup>2</sup>	139	6440 $\pm$ 30	6927 $\pm$ 106.7	
	KIA 38080 <sup>1</sup>	169	7305 $\pm$ 35	7767 $\pm$ 93.5	
	KIA 41025 <sup>1</sup>	192	7815 $\pm$ 45	8285 $\pm$ 97	
	MSM5/5-723-2	KIA 38738 <sup>3</sup>	11.5	675 $\pm$ 25	319 $\pm$ 68
		KIA 38700 <sup>3</sup>	51.5	2125 $\pm$ 25	1714 $\pm$ 93
KIA 43851 <sup>3</sup>		102.5	3820 $\pm$ 30	3769 $\pm$ 99	
KIA 38739 <sup>3</sup>		131.5	4950 $\pm$ 35	5294 $\pm$ 119	
KIA 43853 <sup>3</sup>		181	6120 $\pm$ 40	6545 $\pm$ 108	
KIA 38740 <sup>3</sup>		231.5	7290 $\pm$ 40	7752 $\pm$ 98	
PS2641-4	LuS 8471	20	995 $\pm$ 60	451 $\pm$ 139	
	LuS 9500	43	1240 $\pm$ 90	644 $\pm$ 182	
	LuS 8469	58	1645 $\pm$ 60	1033 $\pm$ 176.5	
	LuS 9124	90.5	1705 $\pm$ 110	1091 $\pm$ 247	
	LuS 9125	128	2835 $\pm$ 100	2382 $\pm$ 296.5	
	LuS 9502	181.5	3775 $\pm$ 150	3519 $\pm$ 407	
	LuS 8468	230	4625 $\pm$ 60	4640 $\pm$ 198	
	LuS 8470	261.5	5400 $\pm$ 60	5598 $\pm$ 192	
	AAR-2422*	375	6980 $\pm$ 130	7327 $\pm$ 278.5	
	AAR-2688*	413	7600 $\pm$ 70	7893 $\pm$ 196	
	LuS 8467	461.5	8415 $\pm$ 80	8783 $\pm$ 256.5	

#### 5. Results

##### 5.1. West Spitsbergen continental margin (cores MSM5/5-712-2 and MSM5/5-723-2)

On the base of our organic geochemical and IRD records the sedimentary sequence of core MSM5/5-712-2 can be separated into three intervals covering the past 8500 years BP (Fig. 3). Results obtained on core MSM5/5-723-2 cover the past 7000 years BP (Fig. 4).

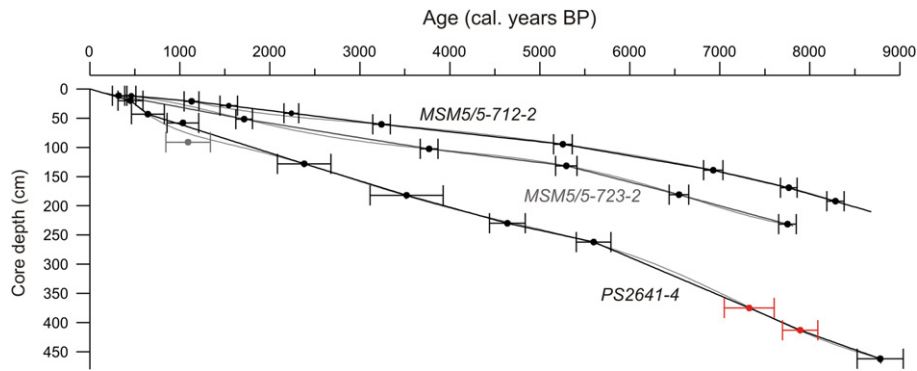
In core MSM5/5-712-2 the late Early Holocene (8500–7000 years BP) is characterised by lowest IRD counts ( $<20$  grains per gram sediment), reduced TOC (0.8–1 wt%) and moderate to maximum  $\text{CaCO}_3$  contents (10–16 wt%). Accumulation of phytoplankton-derived biomarkers (dinosterol and brassicasterol; Boon et al., 1979; Volkman et al., 1998) is at its maximum during this interval, whilst the accumulation of the sea ice proxy  $\text{IP}_{25}$  becomes significantly reduced after 8300 years BP (Fig. 3). Within this period, at about 8200 years BP, a short-term increase in the IRD content coincides with minimum TOC values and lowered phytoplankton biomarker flux rates. Meanwhile, an abrupt decline in previously high  $\text{IP}_{25}$  values is observed between 8300 and 8100 years BP. Further short-term lows in brassicasterol and dinosterol accumulation rates at 7600 and at 7100 years BP are not reflected in the IRD, TOC, or  $\text{IP}_{25}$  data (Fig. 3). Maximum  $\text{CaCO}_3$  contents peak at about 7400 years BP.

During the Mid Holocene (7000–3000 years BP), TOC contents of both *Maria S. Merian* cores reach slightly elevated values (ca 1 wt% in MSM-5/5-712-2; ca 1.2 wt% in MSM5/5-723-2) between 6400 and 5800 years BP and between 4200 and 3400 years BP (Figs. 3 and 4).  $\text{CaCO}_3$  contents of both cores decrease and maintain at minimum values (8–12 wt%). Continuously decreasing accumulation rates of phytoplankton biomarkers in core MSM5/5-712-2, however, are associated with consistently rising IRD and fluctuating  $\text{IP}_{25}$  contents throughout this period (Fig. 3). Considerable dinosterol minima occur at about 5000 and 3200 years BP. The  $\text{IP}_{25}$  record of core MSM5/5-723-2 shows that a period of slightly higher  $\text{IP}_{25}$  flux rates between ca 6200 and 5200 years BP is followed by an  $\text{IP}_{25}$  minimum at about 5000 years BP (Fig. 4). Thereafter, a gradual increase in the accumulation of  $\text{IP}_{25}$  is observed for the Mid Holocene.

The Late Holocene (3000–300 years BP) is marked by further increasing TOC and  $\text{CaCO}_3$  contents in both sediment cores (Figs. 3 and 4). The IRD content in core MSM5/5-712-2 increases as well and reaches maximum values at ca 500 years BP ( $>160$  grains per gram sediment; Fig. 3). We note that the accumulation of biomarkers in this core is highly variable throughout the past 3000 years (Fig. 3), whereas the increase in  $\text{IP}_{25}$  at core site MSM5/5-723-2 occurs to be rather gradual and maximum values are reached at about 300 years BP (Fig. 4). In core MSM5/5-712-2 we observe that during intervals of an elevated  $\text{IP}_{25}$  accumulation at ca 2800, 2300 and 1600 years BP, the flux rates of brassicasterol and dinosterol and also the TOC contents are increased as well (Fig. 3). Vice versa, intermediate periods of lowered  $\text{IP}_{25}$  accumulation correspond to periods of reduced phytoplankton marker contents. Divergent from these in-phase fluctuations, we find minimum phytoplankton marker flux rates that coincide with the youngest  $\text{IP}_{25}$  peak at about 500 years BP (Fig. 3). The TOC content, however, reaches maximum values at this time.

##### 5.2. Inner East Greenland shelf (core PS2641-4)

In comparison with the data from core MSM5/5-712-2, we obtained rather monotonous TOC and  $\text{CaCO}_3$  records from the Holocene section of core PS2641-4 (Fig. 5). TOC contents of



**Fig. 2.** Age model for sediment cores MSM5/5-712-2, MSM5/5-723-2, and PS2641-4 (thin grey lines refer to 5- and 7-point polynomial fits, respectively). Reservoir corrected and calibrated  $^{14}\text{C}$  ages with respective error bars are indicated for each age–depth correlation. Red  $^{14}\text{C}$  ages were determined on shells of the bivalve *Bathyarca glacialis* (Evans et al., 2002).  $^{14}\text{C}$  age at 90.5 cm core depth is not included in the age model of PS2641-4. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

0.8–1.3 wt% remain relatively stable until a distinct increase of similar magnitude than in the cores from the West Spitsbergen slope is observed during the Late Holocene. Notably low  $\text{CaCO}_3$  contents (0.3–1.6 wt%) contrast those of the sediment cores from eastern Fram Strait, and refer to the shift from an Atlantic Water influenced to an EGC dominated sea surface within Fram Strait (Hebbeln and Berner, 1993; Henrich, 1998; Huber et al., 2000). We note that the accumulation rates of biomarkers (in particular of  $\text{IP}_{25}$ ) are significantly higher at core site PS2641-4 than at the West Spitsbergen margin, which probably can be attributed to the basically higher mass accumulation at the East Greenland shelf due to ice rafting.

The late Early Holocene (8500–7000 years BP) is characterised by highest  $\text{CaCO}_3$  contents and elevated phytoplankton marker flux rates, whereas the accumulation of  $\text{IP}_{25}$  fluctuates from minimum to moderate values (Fig. 5). The most remarkable feature of the biomarker distribution of this core is a lack of dinosterol and brassicasterol between 7900 and 7600 years BP. This interval seems to interrupt the late Early Holocene section of relatively high phytoplankton marker contents (Fig. 5).

During the Mid Holocene (7000–3000 years BP), the accumulation of phytoplankton biomarkers decreases until ca 5300 years BP and thereafter increases again. Meanwhile, the  $\text{IP}_{25}$  flux rate first peaks at about 6000 years BP, then decreases and maintains relatively low values between 5200 and 2300 years BP.

For the Late Holocene, we note an increasing accumulation of brassicasterol, dinosterol, and particularly of  $\text{IP}_{25}$  since ca 3000 years BP (Fig. 5). The highest phytoplankton marker contents reach values similar to those seen between 7500 and 6500 years BP. Considerable maxima in all biomarker records and also TOC contents are observed for the past 1000 years BP.  $\text{CaCO}_3$  contents, in contrast, become successively reduced during the past 1000 years BP.

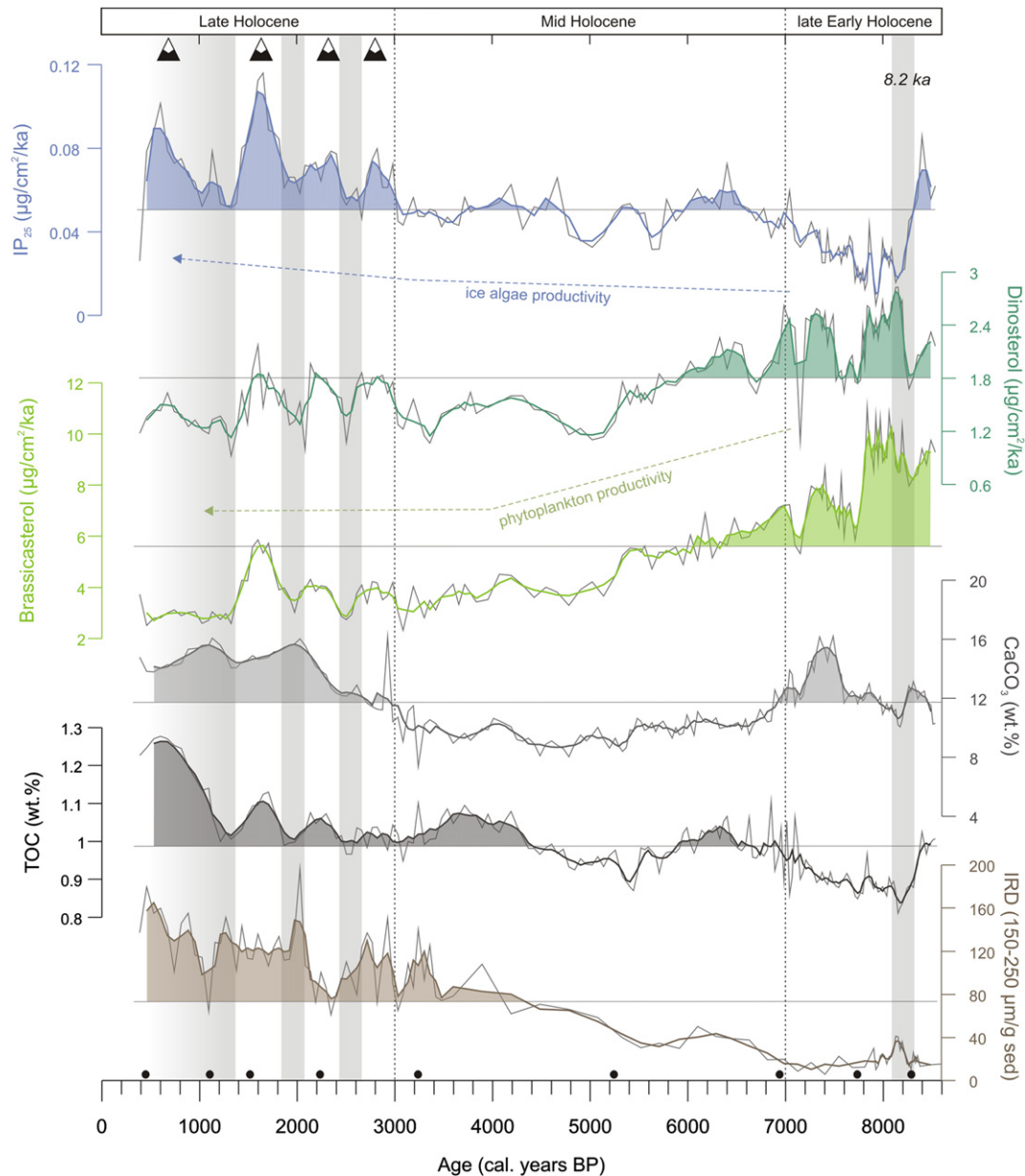
## 6. Discussion

With the identification of the sea ice biomarker  $\text{IP}_{25}$  in the sediment cores MSM5/5-712-2, MSM5/5-723-2, and PS2641-4 we yield novel and direct information about the development of the sea ice conditions along the West Spitsbergen continental margin and the continental shelf of East Greenland throughout the Holocene (Fig. 6). Coincident with the sustained cooling, which is inferred from decreasing  $\delta^{18}\text{O}$  values in the NGRIP Greenland ice core (NGRIP-Members, 2004) and a decline in Northern Hemisphere insolation (Laskar et al., 2004), a general upward trend in

$\text{IP}_{25}$  concentrations is observed in the *Maria S. Merian* cores and also in core PS2837-5 (Fig. 6) located on the Yermak Plateau close to the modern summer sea ice margin (Müller et al., 2009). This increase in  $\text{IP}_{25}$  concentrations – most pronounced during the past 3000 years BP – points to a successive (spatial and temporal) extension of the spring sea ice coverage in eastern Fram Strait, possibly due to a lowering of sea surface temperatures (SST). Previous studies from the Nordic Seas and adjacent areas that are based on diatom (e.g. Jiang et al., 2002; Andersen et al., 2004) or foraminifer assemblages (e.g. Slubowska-Woldengen et al., 2007) or alkenone temperatures (Marchal et al., 2002; Sicre et al., 2008) support cooler ocean temperatures during the Late Holocene. Low orbital forcing, reduced Atlantic water advection as well as a higher ice discharge from the Arctic Ocean towards eastern Fram Strait may have promoted such a cooling.  $\text{IP}_{25}$  concentrations in core PS2641-4 remain, within a certain range of variability, relatively constant throughout the Holocene until a remarkable increase occurs during the past 1000 years BP (Fig. 6). We thus suggest that a continuous export of Arctic sea ice towards this core site persisted throughout the Holocene but not all changes in this discharge system can be traced in sediments from the inner shelf of East Greenland. The consistent atmospheric cooling recorded in the nearby ice cores from Greenland possibly had no significant net effect on the intensity of the sea ice cover and  $\text{IP}_{25}$  sedimentation at core site PS2641-4 until a temperature threshold was reached at 1000 years BP. The previously reported lack of gravel IRD (>2 mm) in the Holocene section of this core (Evans et al., 2002) thus may relate to the generally low melting rates in the central EGC (preventing iceberg melt/debris release). Hence we assume that a strengthened ice discharge from the Arctic Ocean may have resulted in a broadening of the EGC towards the east, which would not have necessarily intensified the sea ice conditions near the coast of East Greenland. A notably weaker increase in sea ice coverage at the East Greenland shelf (compared to eastern Fram Strait), however, is also described in a palaeo-modelling experiment using NAOSIM (for model details see Kauker et al., 2003) between 6000 years BP and pre-industrial times (supplement to Müller et al., 2011). An influence of the nearby Kejsers Franz Joseph Fjord system on the sedimentation regime and hence the biomarker deposition at the inner shelf needs to be considered too.

### 6.1. The late Early Holocene (8500–7000 years BP)

Based on the high phytoplankton biomarker flux rates and minimum  $\text{IP}_{25}$  and IRD contents in core MSM5/5-712-2

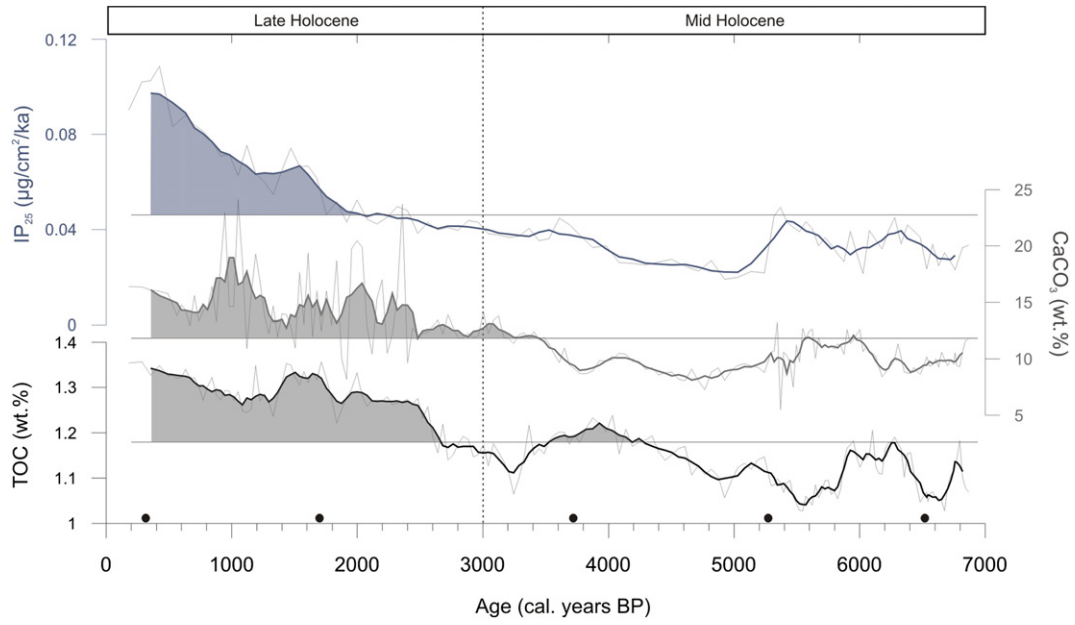


**Fig. 3.** IRD, TOC, and  $\text{CaCO}_3$  contents and biomarker accumulation rates of sediment core MSM5/5-712-2. Bold curves represent 5-point smoothed averages. Curve fillings highlight values above the calculated mean level. Grey vertical bars indicate cooling intervals. Black-white triangles denote glacier advances on Spitsbergen (Svendsen and Mangerud, 1997). Black dots refer to AMS datings.

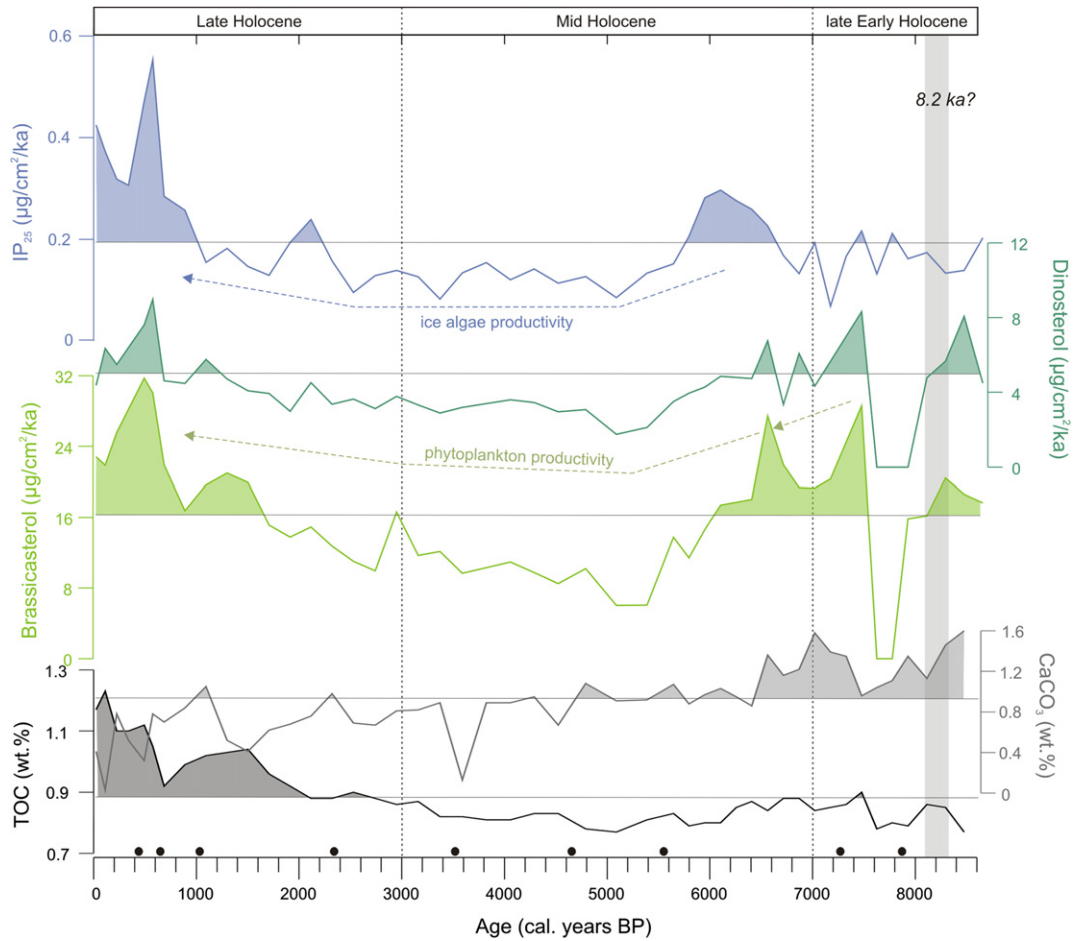
comparatively warm sea surface conditions along the West Spitsbergen shelf can be assumed for the late Early Holocene (Fig. 3). This is also supported by the relatively high (even maximum)  $\text{CaCO}_3$  contents, which, given the low IRD values, may result from a higher productivity of calcareous-walled organisms (e.g. foraminifers, coccoliths) rather than from detrital  $\text{CaCO}_3$  input. We suggest that this interval represents the latest phase of the Holocene Climate Optimum in eastern Fram Strait and consider that in addition to higher insolation values an intensified Atlantic Water inflow likely supported phytoplankton growth, whereas the (spring) sea ice margin was located further northward (i.e. the core site experienced only minor sea ice occurrences during the late winter/early spring months). This aligns with maximum SSTs reconstructed for the western continental margin of the Barents Sea by Sarnthein et al. (2003) and findings of Salvigsen et al. (1992),

who report optimum climate conditions for thermophilous molluscs on western Svalbard for the period between 8700 and 7700 years BP. Likewise, foraminifer-based reconstructions of ocean circulation changes along the West Spitsbergen shelf by Slubowska-Woldengen et al. (2008) reveal a strengthened inflow of Atlantic Water at that time.

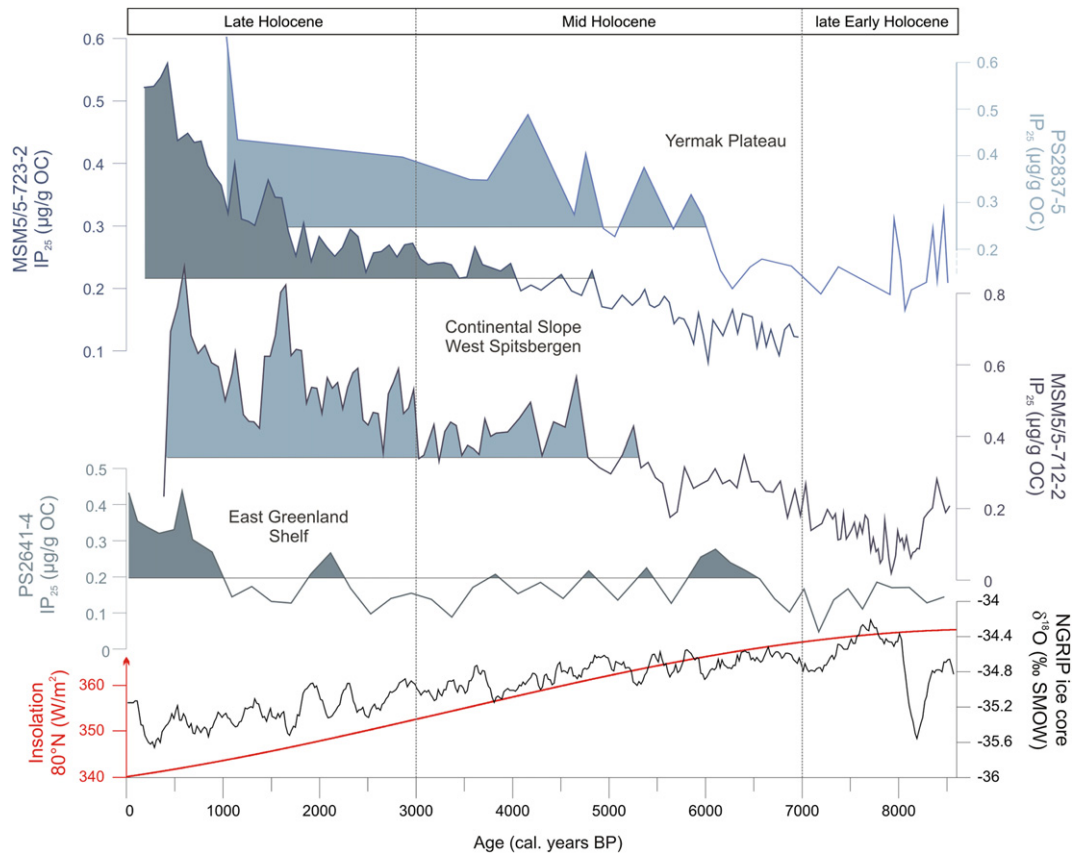
The abrupt reductions in the phytoplankton marker contents of core MSM5/5-712-2 at about 8300, 7600 and 7100 years BP that punctuate the late Early Holocene probably point to short-term deteriorations of the sea surface conditions. Since the decline at ca 8200 years BP – a prominent cooling event in the High Latitudes (Alley et al., 1997; Clarke et al., 2004; Kleiven et al., 2008) – coincides with a (somewhat retarded) minor short-term increase in IRD and a rapid decrease of previously high  $\text{IP}_{25}$  contents, we assume that the core site was affected by a massive ice discharge that



**Fig. 4.** TOC and CaCO<sub>3</sub> contents and IP<sub>25</sub> accumulation rates of sediment core MSM5/5-723-2. Bold curves represent 5-point smoothed averages. Curve fillings highlight values above the calculated mean level. Black dots refer to AMS datings.



**Fig. 5.** TOC and CaCO<sub>3</sub> contents and biomarker accumulation rates of sediment core PS2641-4. Curve fillings highlight values above the calculated mean level. Black dots refer to AMS dating points.



**Fig. 6.** Comparison of IP<sub>25</sub> concentrations (normalised to gram organic carbon) of sediment cores from northern, eastern, and western Fram Strait. Summer insolation for 80°N (red curve) taken from Laskar et al. (2004) and  $\delta^{18}\text{O}$  values from the NGRIP ice core (NGRIP Members, 2004) support a Holocene cooling. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

reduced not only the growth of phytoplankton but also that of ice algae. This agrees with reduced SSTs along the Barents Sea margin for this period (Sarnthein et al., 2003). Similarly, Müller et al. (2009) interpreted minimum fluxes of IP<sub>25</sub> and brassicasterol as indicative for a near-perennial sea ice cover at the western Yermak Plateau in northern Fram Strait at about 8200 years BP.

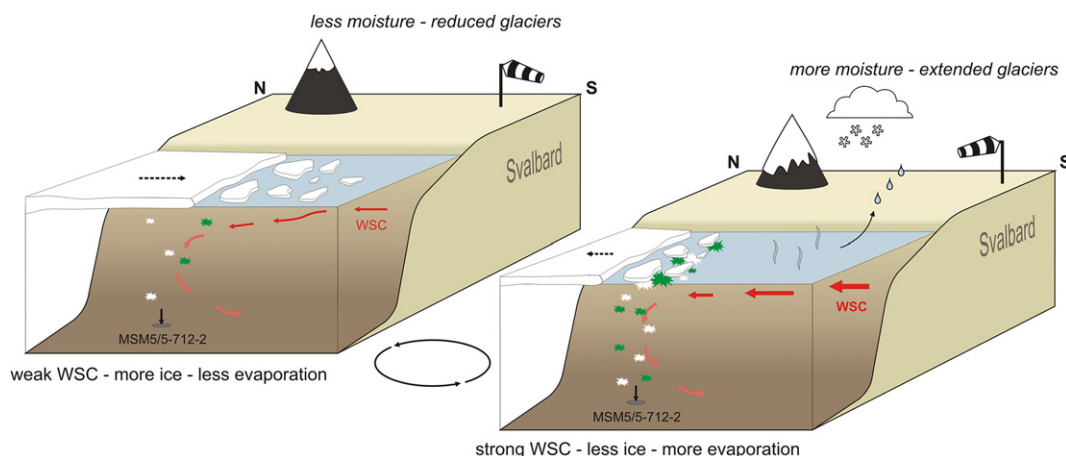
Relatively warm conditions probably also prevailed along the East Greenland shelf during the late Early Holocene since at core site PS2641-4 accumulation rates are relatively high for dinosterol and brassicasterol, and only moderate for IP<sub>25</sub> (Fig. 5), which may point to a reduced (not absent) ice cover at the inner East Greenland shelf. Minor sea ice cover, a higher release of nutrient rich freshwater from the melting Greenland ice sheet and changes in the local (Fjord) circulation system thus may have promoted the growth of phytoplankton. Similarly, e.g. Bauch et al. (2001) and Andersen et al. (2004) reconstruct rather warm sea surface conditions in the central Nordic Seas and at the East Greenland shelf for the late Early Holocene by means of foraminifer and diatom assemblages. This period likely corresponds to the retreat of the Greenland ice sheet from the inner shelf as is reconstructed by Evans et al. (2002) on base of sediment lithology and  $\delta^{18}\text{O}$  data from core PS2641-4. The sudden lack of the phytoplankton markers between 7900 and 7600 years BP could point to a short-term deterioration in sea surface conditions, which is not reflected in the IP<sub>25</sub> contents (Fig. 5). Increased meltwater discharge from the adjacent Kejser Franz Joseph Fjord either could have caused a strong stratification of the upper water layer, thus reducing the ventilation and availability of nutrients required for phytoplankton growth or, in contrast, it may have increased the turbidity, which

may suppress phytoplankton productivity as well. Another possibility would be that local sea ice formation benefited from the episodic release of cold and fresh meltwater from the Greenland ice sheet. If this was the case, however, we do not see a notable effect on the accumulation of IP<sub>25</sub>.

## 6.2. The Mid Holocene (7000–3000 years BP)

Lowered CaCO<sub>3</sub> contents, continuously decreasing accumulation rates of phytoplankton markers, the sustained increase in IRD and slightly higher IP<sub>25</sub> contents in sediment core MSM5/5-712-2 (and MSM5/5-723-2; Figs. 3 and 4) point to a gradually reduced phytoplankton productivity due to a cooling of the sea surface and a successive growth and extension of (winter/spring) sea ice at the continental slope of West Spitsbergen during the Mid Holocene. This agrees with a concurrent increase in IRD contents of sediment cores along the West Spitsbergen continental shelf and margin documented by Jessen et al. (2010) and Slubowska-Woldengen et al. (2007), who assume a Mid Holocene ice advance. Decreasing SSTs off and enhanced glaciation on West Spitsbergen (Hald et al., 2004) support this interpretation. To what extent this Holocene cooling trend observed in the subpolar North Atlantic domain (e.g. Andersen et al., 2004; Hald et al., 2007; Müller et al., 2010) may be related to the lowered Northern Hemisphere insolation and/or a reduced Atlantic Water advection and/or changes in the atmospheric circulation remains unknown. A general recovery of the Arctic's sea ice after its significant recession during the Holocene Climate Optimum (see e.g. Polyak et al., 2010 for review), however, occurs as a plausible and natural response to the





**Fig. 7.** Scheme illustrating periods of Neoglacial sea ice advance (left) and retreat (right) at the continental margin of West Spitsbergen due to variations in the inflow of warm Atlantic Water via the WSC and/or weakened/strengthened westerlies. Associated fluctuations in the productivity of ice algae (white patches) and phytoplankton (green patches) are indicated. Periods of ice retreat promote glacier growth on Spitsbergen due to an increased moisture supply from the sea. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

mitigation of orbital forcing. Sustained oceanic surface cooling that stimulated the sea ice formation during winter and retarded its retreat/melt during the late spring and early summer months is also supported by e.g. Rasmussen et al. (2007) and Jennings et al. (2002) who reconstruct increasingly cooler conditions along the West Spitsbergen shelf and an increased sea ice export through Fram Strait by means of benthic and planktic foraminifera and IRD records. Furthermore, on the bases of e.g. increasing  $\delta^{18}\text{O}$  values, increasing abundances of *N. pachyderma* (sin.) and diatom data, Rasmussen et al. (2007), Bauch et al. (2001) and Koç et al. (1993) strengthen that the southwestern continental margin of Spitsbergen and the Nordic Seas experienced an intensified water mass exchange with the Arctic Ocean for the period after 7000 years BP and continuous surface cooling (in-step with the lowered insolation) since ca 6000 years BP. Reduced winter SSTs are also reconstructed for the western continental margin of the Barents Sea and have been related to a weakened heat input with the WSC and a strengthened East Spitsbergen (Sørkapp) Current leading to periods of extended sea ice coverage (Sarnthein et al., 2003).

Successively reduced phytoplankton marker contents in core PS2641-4 indicate that also the East Greenland shelf became gradually cooled between ca 6600 and 3000 years BP (Fig. 5). A higher accumulation of  $\text{IP}_{25}$  around 6000 years BP may point to an increase in sea ice cover at this time. Similarly, Ran et al. (2006) interpret higher abundances of Arctic diatom taxa in a sediment core from the northern shelf of Iceland as indicative of a strengthening of the EGC between 6800 and 5500 years BP. The subsequent relatively low and uniform (i.e. with some minor fluctuations)  $\text{IP}_{25}$  and phytoplankton marker flux rates, however, suggest that largely constant sea ice conditions prevailed at the inner East Greenland shelf between 5000 and 2500 years BP. Recent reconstructions by Jennings et al. (2011) document that an intensified Irminger Current carried more warm Atlantic Water to Denmark Strait between 6800 and 3500 years BP. This could be related to a weakening of the EGC during this period, but is not reflected in the  $\text{IP}_{25}$  or phytoplankton marker data at our core site, which may reveal an influence of the nearby Fjord circulation and sedimentation processes. Moros et al. (2006), in turn, observe a long-term trend of increasing drift ice export via the EGC towards the North Atlantic since the past 5000 years. With respect to our biomarker data, we suggest that Mid Holocene variations in the strength or extent of the EGC, however, are not clearly recorded in sediments from the inner shelf but rather could be traced at the outer shelf of East

Greenland where the environment is less affected by Fjord conditions.

### 6.3. The Late Holocene (the past 3000 years BP)

Maximum IRD release and a sustained increase in the accumulation of  $\text{IP}_{25}$  during the past 3000 years – a period that is widely acknowledged as Neoglacial cooling phase (for recent review see Miller et al., 2010) – point to intensified sea ice occurrences at the West Spitsbergen continental margin. Forwick and Vorren (2009) and Forwick et al. (2010) assume an enhanced formation of shore-fast sea ice and/or a permanent sea ice cover along the West Spitsbergen coast that trapped IRD laden icebergs within the Isfjorden system during the past ca 4000 years BP. Thus, the elevated IRD contents at core site MSM5/5-712-2 may suggest a transport of lithic grains by sea ice rather than by icebergs originating from Spitsbergen glaciers. Further reconstructions of gradually cooled sea surface temperatures, lowered productivity and a higher polar water outflow to the Nordic Seas during the past 3000 years BP support this general increase in sea ice coverage (Koç et al., 1993; Andrews et al., 2001; Calvo et al., 2002; Jennings et al., 2002, 2011; Andersen et al., 2004).

We suggest that the in-phase fluctuations of  $\text{IP}_{25}$  and phytoplankton marker contents in core MSM5/5-712-2 (Fig. 3) can be attributed to periods of a rapidly advancing and retreating sea ice margin at this core site until ca 1200 years BP. Accordingly, the less variable though steadily rising accumulation of  $\text{IP}_{25}$  at core site MSM5/5-723-2 – ca 40 km further to the north of core site MSM5/5-712-2 – relates to a continuously increasing ice coverage during the past 3000 years BP. As marine primary productivity is demonstrably stimulated in the marginal ice zone (release of nutrients from the melting sea ice triggers the bloom of the phytoplankton algae in the proximity of the ice edge; Hebbeln and Wefer, 1991; Ramseier et al., 1999; Sakshaug, 2004; Smith et al., 1987), we conclude that the periods of peak  $\text{IP}_{25}$  and phytoplankton marker contents in core MSM5/5-712-2 at about 2800, 2300 and 1600 years BP reflect beneficial living conditions at the ice edge for both sea ice algae and plankton thriving in open water (Müller et al., 2009, 2011). Interestingly, Svendsen and Mangerud (1997) report concurrent (with minor temporal shifts) periods of abrupt glacier advances on West Spitsbergen. Since glacier growth requires a higher winter precipitation (and the main moisture source for the Svalbard archipelago is the subpolar North Atlantic;

Dickson et al., 2000; Humlum et al., 2005), we hence assume that these intervals of glacier advance and ice edge conditions at the core site may have been triggered by a temporarily strengthened WSC and/or changes in the atmospheric circulation pattern, which caused these recurrent northward retreats of the sea ice cover (Fig. 7). This is also supported by findings of Sarnthein et al. (2003) who identify two intervals of remarkably warmer SSTs and peak abundances of the foraminifer *Turborotalita quinqueloba* at the western continental margin of the Barents Sea at about 2200 and 1700 years BP, which they attribute to short-term pulses of warm Atlantic Water advection. Intermediate periods of lowered IP<sub>25</sub> and phytoplankton marker contents accordingly point to a weakened WSC – weakened through a lowered heat and/or volume transport – and probably an increased ice discharge from the Arctic Ocean that permitted sea ice advances beyond the core site (Fig. 7). Intensified sea ice advection via the Sørkapp Current is possible as well. Such a fluctuating ice margin (due to a variable Atlantic Water inflow) is also described in a recent study by Werner et al. (2011) on the bases of foraminifer and IRD data obtained from a box core that was recovered at the same core site. Both short-term anomalies in atmospheric pressure fields (controlling the intensity and strength of wind and oceanic sea surface current patterns) or even variations in the thermohaline circulation may have caused these sea ice fluctuations. Given the comparatively “sluggish” behaviour of thermohaline convection processes, the latter, however, appears to be a less probable explanation for such rapid oscillations.

The notably reduced accumulation of dinosterol and brassicasterol in core MSM5/5-712-2 during the past 1300 years BP may indicate a deterioration of the sea surface conditions that limited the productivity of phytoplankton at the West Spitsbergen margin. Meanwhile increasing IP<sub>25</sub> flux rates and highest IRD contents at about 600 years BP suggest an increase in sea ice coverage and the occurrence of icebergs that probably carried organic material, which could account for maximum TOC values. The possibility that a general increase in sea ice algal blooms during spring could account for this increase in TOC may be considered too. The Late Holocene maximum content of CaCO<sub>3</sub> thus could be related to a higher productivity of calcareous-walled plankton grazing on ice algae if it is not due to a higher input of calcareous IRD. Maximum IP<sub>25</sub> concentrations in sediment core MSM5/5-723-2 (Fig. 4), however, support this Late Holocene sea ice advance.

In contrast to the finding for eastern Fram Strait, the stepwise increase in IP<sub>25</sub> flux rates at the East Greenland shelf since 2500 years BP is accompanied by a continuously rising accumulation of brassicasterol and dinosterol (Fig. 5). The onset of this increase in IP<sub>25</sub> aligns with findings of Sicre et al. (2008), who document that at about 2500 years BP a trend of slightly warmer SSTs north off Iceland was reversed by a cooling tendency towards present. The short interval of warmer SSTs during the Medieval Warm Period (around 1000 years BP), however, is not clearly distinguishable in core PS2641-4. Instead, maximum fluxes of IP<sub>25</sub> and phytoplankton biomarkers point to favourable living conditions for both ice and phytoplankton algae at the East Greenland shelf during the past 1000 years BP. Further reduced CaCO<sub>3</sub> contents in this core hence may possibly be attributed to calcium carbonate dissolution due to an increased formation of corrosive bottom waters linked to higher sea ice production and/or organic matter (originating from phytoplankton blooms) remineralisation (Steinsund and Hald, 1994). Alternatively, this reduction in CaCO<sub>3</sub> may indicate a shift of the provenance of ice rafted (carbonate) material. The highest IP<sub>25</sub> contents determined for the past 600 years BP in all cores are tentatively attributable to the ‘Little Ice Age’ cooling that is also recorded in further marine (e.g. Andersson et al., 2003; Moros et al., 2006; Bendle and Rosell-Melé, 2007; Spielhagen et al., 2011) and also terrestrial archives from the subpolar North Atlantic (e.g. Nesje

et al., 2001; Seppä and Birks, 2002). The occurrence of a broad and severe (i.e. perennial) sea ice cover during this period, however, seems to be contradicted by the elevated phytoplankton marker contents in core PS2641-4. We hence conclude that during the past ca 600 years BP, stable marginal ice zone (probably polynya-like) conditions established at the inner East Greenland shelf, whereas eastern Fram Strait experienced an ice advance that reduced phytoplankton productivity. Assumptions that this ‘Little Ice Age’ North Atlantic cooling may result from a weakening of the Gulf Stream and a slowing of the thermohaline circulation (Broecker, 2000; Lund et al., 2006) would support the interpretation of high IP<sub>25</sub> and IRD contents in sediments from the West Spitsbergen slope as indicative of an extended ice cover at these sites. The preceding ‘Medieval Warm Period’, however, is not reflected (or resolved) in the biomarker records, though we note a short-term decrease in IRD in MSM5/5-712-2 at about 1100 years BP suggesting reduced iceberg occurrences.

#### 6.4. Sea ice fluctuations and North Atlantic Oscillation

With respect to the Late Holocene cooling that is observed at various sites in the subpolar North Atlantic and also in more remote areas, the hypothesis evolved that a continuous transition from a positive towards a negative NAO phase characterised the Holocene climate development (e.g. Andersen et al., 2004). Reduced Siberian river discharge during the past 2000 years BP and vegetation changes in northeast European Russia, for example, are interpreted to reflect the development of a colder and dryer climate in the Eurasian Arctic and could be related to negative NAO-like conditions (Stein et al., 2004; Salonen et al., 2011). And also Jessen et al. (2011) relate changes in Late Holocene pollen records from southern Greenland and the Labrador Sea to a distinct reduction of south-westerly air masses in favour of colder air originating from the north and thus conclude that the atmospheric circulation pattern in the subpolar North Atlantic likely shifted from a more positive to a more negative NAO. Sicre et al. (2008) attribute short-term SST changes off North Iceland during the Late Holocene to low frequency NAO forcing that partly seems to be associated with fluctuations in the meridional overturning circulation (Latif et al., 2006) and thus (large-scale) alterations in the Atlantic hydrological cycle. Regarding the comparatively rapid sea ice oscillations at core site MSM5/5-712-2 between 3000 and 1200 years BP, we consider that a relationship between sea ice extent and oceanic-atmospheric (i.e. NAO-like) forcing fields, in fact, could explain the observed fluctuations. As no (instrumental) records of the long-term (centennial- to millennial-scale) development of the NAO are available, its influence on Late Holocene environmental conditions in the subpolar North Atlantic remains elusive. Proxy- and model-based reconstructions of NAO conditions (e.g. Luterbacher et al., 2001; Trouet et al., 2009 and references therein) thus may provide valuable palaeoenvironmental information, though they are mainly confined to the past 1000 years BP – depending on the availability of respective proxy data for calibration. Trouet et al. (2009), for example, find that the ‘Medieval Warm Period’ was associated with a positive NAO mode, while a negative mode prevailed during the ‘Little Ice Age’. A genuine attempt to directly compare (and link) the biomarker fluctuations in core MSM5/5-712-2 during the past 3000 years with reconstructed NAO indices proves problematic as no NAO reconstructions are available for this time period. Shifting NAO conditions, however, could account for respective changes in the strength of westerly storm tracks and Atlantic Water advection to the continental margin of Spitsbergen (Kwok and Rothrock, 1999; Dickson et al., 2000; Hurrell and Deser, 2009). Neglecting the temporal resolution of this core (1 cm represents ca 40 years) and that IP<sub>25</sub> fluctuations follow

a rather multi-centennial than an annual to decadal (NAO-specific) pattern, we may link our findings with atmospheric circulation shifts. Positive NAO-like conditions could have prevailed during periods of elevated IP<sub>25</sub> and phytoplankton biomarker contents and glacier advances on Spitsbergen, while a negative NAO-like forcing may have promoted the recurrent southward advances of sea ice that punctuated these ice edge productivity intervals. Previously, the observation of fluctuating glacier extents in southwest Norway lead Nesje et al. (2001) and Imhof et al. (2012) to establish a relationship between Late Holocene glacier advances and positive NAO phases causing more humid and wet winter conditions over Scandinavia. Similarly, Giraudeau et al. (2010) credit both an increased advection of Atlantic Water into the Norwegian Sea and a coincidentally strengthened polar outflow towards the western Nordic Seas to positive NAO intervals during the Late Holocene. Such a seesaw pattern between warm water input through eastern and cold water output via western Fram Strait, however, is not observed in our records. With general regard to the Holocene climatic development, we notice that the sea ice conditions at core site PS2641-4 obviously were less prone to variations in the strength of the oceanic (and atmospheric) circulation system than the sites in eastern Fram Strait. The rapid sea ice fluctuations reflected in the record of core MSM5/5-712-2 during the past 3000 years BP and the likely associated changes in the advection of warm (WSC) and polar (EGC) water masses are not fully traceable in the record of core PS2641-4. Due to its location in the vicinity of the Arctic and Atlantic oceanic (and atmospheric) fronts, core MSM5/5-712-2 apparently experienced more significant palaeoceanographic and environmental changes, while the setting at the inner shelf of East Greenland remained relatively unaffected.

6.5. PIP<sub>25</sub> index and sea ice estimate

We recently demonstrated that the coupling of the environmental (sea surface) information carried by IP<sub>25</sub> and phytoplankton biomarkers by means of a phytoplankton-IP<sub>25</sub> index (PIP<sub>25</sub>) proves a valuable approach for quantitative reconstructions of (spring) sea ice coverage (Müller et al., 2011). A distinct connection between the

sea ice distribution and sedimentary IP<sub>25</sub> and phytoplankton marker contents is strengthened through correlation analyses of PIP<sub>25</sub> indices determined on surface sediments from the subpolar North Atlantic with sea ice concentrations derived from satellite and modelling data (Müller et al., 2011). According to the respective phytoplankton biomarker used for the calculation (brassicasterol or dinosterol), this index is specified as P<sub>B</sub>IP<sub>25</sub> or P<sub>D</sub>IP<sub>25</sub>, respectively. Highest PIP<sub>25</sub> (P<sub>B</sub>IP<sub>25</sub> and P<sub>D</sub>IP<sub>25</sub>) values in the range of 0.75–1 seem to reflect extended ice coverage throughout spring and summer, whilst minimum values refer to predominantly ice-free (spring/summer) sea surface conditions. Intermediate values (0.5–0.75) characterise sites within the productive marginal ice zone.

Throughout the past 8000 years BP, P<sub>B</sub>IP<sub>25</sub> and P<sub>D</sub>IP<sub>25</sub> indices calculated for core MSM5/5-712-2 sediments rise gradually (Fig. 8) and thus point to a general increase in (spring) sea ice coverage at the West Spitsbergen continental margin. Minimum and close to zero values denote a period of significantly reduced ice cover between 8200 and 7800 years BP. With reference to the findings of Müller et al. (2011) these values may refer to a sea ice concentration of less than 20%. Given the infancy of this approach, this interpretation of P<sub>B</sub>IP<sub>25</sub> and P<sub>D</sub>IP<sub>25</sub> values in terms of sea ice concentrations, however, needs to be considered as a very rough estimate. A sustained increase in sea ice occurrences at the core site is reflected by further rising P<sub>B</sub>IP<sub>25</sub> and P<sub>D</sub>IP<sub>25</sub> values, which finally pass the “threshold level” of 0.5 designating marginal ice zone conditions from ca 4800 years BP on. Maximum P<sub>B</sub>IP<sub>25</sub> values determined for the past ca 1000 years are in the range of 0.75–0.80 and thus indicate a shift towards severe ice coverage (presumably > 70% ice concentration).

P<sub>B</sub>IP<sub>25</sub> and P<sub>D</sub>IP<sub>25</sub> indices calculated for core PS2641-4 show minimum values during the late Early Holocene (8700–7200 years BP; Fig. 8) that point towards a variable or less pronounced sea ice coverage at the inner shelf of East Greenland. Given the lack of brassicasterol and dinosterol between 7900 and 7600 years BP, P<sub>B</sub>IP<sub>25</sub> and P<sub>D</sub>IP<sub>25</sub> values are 1 for this interval, which would refer to a significantly extended (spatially and temporally) ice cover. Throughout the past 7000 years BP, P<sub>B</sub>IP<sub>25</sub> and P<sub>D</sub>IP<sub>25</sub> indices of core PS2641-4 fluctuate between values of ca 0.4 to 0.6 (Fig. 8), which

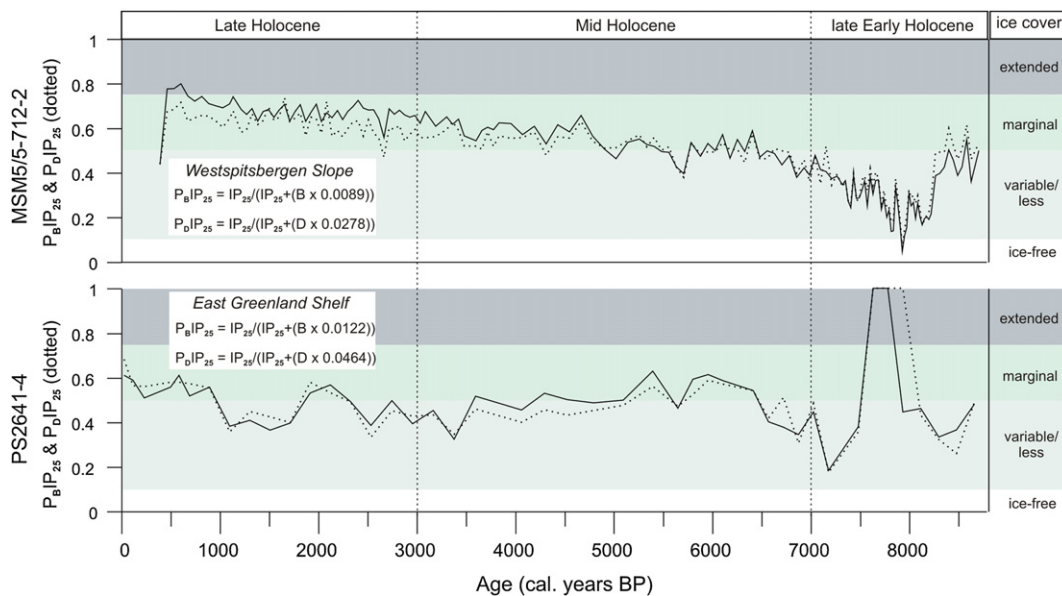


Fig. 8. P<sub>B</sub>IP<sub>25</sub> and P<sub>D</sub>IP<sub>25</sub> indices calculated for sediment cores MSM5/5-712-2 (top) and PS2641-4 (bottom). Indices were calculated using IP<sub>25</sub>, brassicasterol, and dinosterol accumulation rates and respective balance factors following Müller et al. (2011). Blue shadings refer to estimates of sea ice conditions (PIP<sub>25</sub> > 0.1 variable, >0.5 marginal, >0.75 extended ice cover) according to Müller et al. (2011). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

may denote variable to stable marginal sea ice conditions (20–50% ice concentration) in western Fram Strait throughout the Mid and Late Holocene. The observation that  $P_{BIP_{25}}$  and  $P_{DIP_{25}}$  indices obtained from this core are largely in the same range or even lower than those from core MSM5/5-712-2 occurs somewhat surprising as sea ice conditions at the East Greenland shelf are generally considered to be more severe due to the pronounced discharge of sea ice and icebergs via the EGC. In a previous study, the absence of coarse-grained IRD in the PS2641-4 sediments has been interpreted to reflect a sedimentation regime that is dominated by meltwater discharge rather than by iceberg rafting (Evans et al., 2002). This prompts the assumption that much of the brassicasterol and dinosterol identified in this core could originate from the adjacent Fjord systems and is advected through massive meltwater releases in summer. In fact, higher concentrations of algae-specific fatty acids, brassicasterol and dinosterol (pointing to a higher phytoplankton productivity) are observed in the surface sediments from different East Greenland Fjords compared to sediments from the East Greenland shelf at similar latitude (Kierdorf, 2006; Müller et al., 2011). While calculating the  $P_{BIP_{25}}$  and  $P_{DIP_{25}}$  indices, the amount of “in-situ” phytoplankton accordingly could be overestimated due to lateral advection of biomarkers, which would result in (too) low  $P_{BIP_{25}}$  and  $P_{DIP_{25}}$  values.

The short-term variability observed in the biomarker records of core MSM5/5-712-2 during the Late Holocene (Fig. 3), however, is not reproduced by the  $P_{BIP_{25}}$  and  $P_{DIP_{25}}$  indices of this core. The in-phase fluctuations of  $IP_{25}$  and phytoplankton marker contents referring to rapid advances and retreats of the sea ice margin seem to be counterbalanced through the index calculation. Coevally high amounts of  $IP_{25}$ , brassicasterol, and dinosterol (indicating beneficial ice-edge conditions) as well as coevally low biomarker contents (suggesting an unfavourable/severe ice cover) thus give the same  $PIP_{25}$  values – as was already suggested by Müller et al. (2011). We hence conclude that for the proper identification of different sea ice conditions individual biomarker concentrations need to be known and accordingly interpreted.

Nonetheless, we consider these  $PIP_{25}$  indices to describe sea ice conditions a promising means, as they may enable a more quantitative assessment of the ice coverage (once the modern analogue calibration data set has been extended) and thus provide valuable information that can be used for e.g. palaeo freshwater budget estimates.

## 7. Conclusions

The Holocene sea ice evolution in eastern and western Fram Strait is reconstructed by means of the sea ice proxy  $IP_{25}$ , IRD data and the phytoplankton-derived biomarkers brassicasterol and dinosterol. In line with a lowered Northern Hemisphere insolation and decreasing temperatures, the (spring) sea ice coverage along the western continental margin of Spitsbergen increased between 8500 and 1000 years BP. In contrast, sea ice conditions at the inner shelf of East Greenland probably remained relatively constant over this period. Estimates of the sea ice conditions based on phytoplankton- $IP_{25}$  indices ( $P_{BIP_{25}}$  and  $P_{DIP_{25}}$ ) reveal a Mid Holocene shift from a reduced ice cover to marginal ice zone conditions in eastern Fram Strait, whereas the inner East Greenland shelf – according to  $PIP_{25}$  estimates – would have experienced a predominantly marginal ice cover throughout the past 7000 years BP. During the Neoglacial, a rapidly advancing and retreating sea ice margin characterised the environmental setting at the West Spitsbergen slope. This points to a highly variable inflow of warm Atlantic Water, which, in turn, could be associated with short-term changes in the oceanic-atmospheric forcing pattern. This variability

is not clearly reflected in the core from the East Greenland shelf, where only a delayed Late Holocene ice advance (i.e. a pronounced increase in  $IP_{25}$  contents) is observed for the past 1000 years BP. The use of  $IP_{25}$  and phytoplankton markers, however, proves a valuable combinatory approach for the assessment of sea surface conditions. Regarding the  $P_{BIP_{25}}$  and  $P_{DIP_{25}}$  indices we recommend that the reconstruction of sea ice coverage should not be solely based on these ratios but that individual biomarker contents need to be considered as well.

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## Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1594/PANGAEA.779628.

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