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1 A Late Glacial-early Holocene multiproxy record from the eastern Fram Strait,
2 Polar North Atlantic

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21 **Abstract**

22 The paleoceanographic development of the eastern Fram Strait during the transition from the
23 cold Late Glacial and into the warm early Holocene was elucidated via a multiproxy study of
24 a marine sediment record retrieved at the western Svalbard slope. The multiproxy study
25 includes analyses of planktic foraminiferal fauna, bulk sediment grain size and CaCO_3
26 content in addition to Mg/Ca ratios and stable isotopes ($\delta^{13}\text{C}$ and $\delta^{18}\text{O}$) measured on the
27 planktic foraminifer *Neogloboquadrina pachyderma*. Furthermore paleo subsurface water
28 temperatures were reconstructed via Mg/Ca ratios ($\text{sSST}_{\text{Mg/Ca}}$) and transfer functions
29 ($\text{sSST}_{\text{Transfer}}$) enabling comparison between the two proxies within a single record. The age
30 model was constrained by four accelerator mass spectrometry (AMS) ^{14}C dates.
31 From 14,000 to 10,300 cal yr B.P. *N. pachyderma* dominated the planktic fauna and cold
32 polar sea surface conditions existed. The period was characterized by extensive sea ice cover,
33 iceberg transport and low sub sea surface temperatures ($\text{sSST}_{\text{Transfer}} \sim 2.1^\circ\text{C}$; $\text{sSST}_{\text{Mg/Ca}}$
34 $\sim 3.5^\circ\text{C}$) resulting in restricted primary production. Atlantic Water inflow was reduced
35 compared to the present-day and likely existed as a subsurface current. At ca. 10,300 cal yr
36 B.P. Atlantic Water inflow increased and the Arctic Front retreated north-westward resulting
37 in increased primary productivity, higher foraminiferal fluxes and a reduction in sea ice cover
38 and iceberg transport. The fauna rapidly became dominated by the subpolar planktic
39 foraminifer *Turborotalita quinqueloba* and summer $\text{sSST}_{\text{Transfer}}$ increased by $\sim 3.5^\circ\text{C}$.
40 Concurrently, the $\text{sSST}_{\text{Mg/Ca}}$ recorded by *N. pachyderma* rose only $\sim 0.5^\circ\text{C}$. From ca. 10,300
41 to 8,600 cal yr B.P. the average $\text{sSST}_{\text{Mg/Ca}}$ and $\text{sSST}_{\text{Transfer}}$ were $\sim 4.0^\circ\text{C}$ and $\sim 5.5^\circ\text{C}$,
42 respectively. The relatively modest change in $\text{sSST}_{\text{Mg/Ca}}$ compared to $\text{sSST}_{\text{Transfer}}$ can
43 probably be tied to a change of the main habitat depth and/or shift in the calcification season
44 for *N. pachyderma* during this period.

45 Keywords: Planktic foraminifera, Sub sea surface temperature reconstruction, Trace
46 elements, Transfer functions, Stable isotopes, Late Glacial/Holocene transition, Fram Strait,
47 Polar North Atlantic

48 **1. Introduction**

49 The Arctic region has responded strongly to both modern and past global climate
50 changes (IPCC, 2007; Hald et al., 2007). One of the major components controlling the Arctic
51 environment, including sea ice distribution, is the influx and volume of relatively warm and
52 saline Atlantic Water flowing northwards into the Arctic Ocean (Schauer et al., 2004; IPCC,
53 2007). This influx primarily takes place through two gateways: the deep eastern Fram Strait
54 (Schauer et al., 2004) and the shallower Barents Sea (Schauer et al., 2002). In order to
55 improve our understanding of future climate changes in the Arctic region, it is a prerequisite
56 to understand and quantify past rapid oceanic changes with regard to water volume
57 transports, -temperature and -salinity.

58 Previous studies of the spatial and temporal oceanographic evolution in the Fram
59 Strait show rapid changes through the Late Glacial – early Holocene transition. Atlantic
60 Water masses were advected into the area during the Bølling-Allerød interstadial (Ślubowska
61 et al., 2005; Ślubowska-Woldengen et al., 2007; Ebbesen et al., 2007; Rasmussen et al.,
62 2007b), and during the Younger Dryas stadial the Atlantic Water advection continued
63 submerged under a layer of polar surface water (Rasmussen et al., 2007b). During the early
64 Holocene, advection of Atlantic Water was strong and it rapidly became dominant in the
65 surface water masses (Ebbesen et al., 2007; Hald et al., 2007). These paleoceanographic
66 reconstructions are based on a wide range of proxies, and with regard to estimation of past
67 sea surface temperatures (SST) primarily stable isotopes measured in planktic foraminifera
68 and transfer functions were used. Stable oxygen isotopes ($\delta^{18}\text{O}$) measured in foraminiferal

69 calcite inherently reflect the combined signal of salinity and temperature, while stable carbon
70 isotopes ($\delta^{13}\text{C}$) reflect the degree of ventilation and primary production in the ambient water
71 masses (e.g., Spielhagen and Erlenkeuser, 1994; Katz et al., 2010). Reconstructions of SST
72 using transfer functions in the Arctic have been encumbered by several factors, including a
73 restricted geographical distribution of the modern database (e.g., Kucera et al., 2005).
74 Furthermore, it has been found that polar and subpolar planktic foraminifera migrate through
75 the water column and have their maximum occurrence from 50 to 150 m water depth (e.g.,
76 Carstens et al., 1997; Volkman, 2000). Husum and Hald (2012) used both annual and
77 seasonal temperatures from different water depths and found the most robust statistical model
78 using summer temperatures from 100 m water depth.

79 The aim of the present study is to improve qualitative and quantitative estimates of the
80 paleoceanographic variability in the eastern Fram Strait during the transition from the Late
81 Glacial into the early Holocene (14,000 - 8,600 cal yr B.P.). A multi proxy analysis was
82 carried out using a sediment core retrieved from 1487 m water depth at the western Svalbard
83 slope, eastern Fram Strait. We used stable isotopes ($\delta^{18}\text{O}$ and $\delta^{13}\text{C}$) measured on tests of *N.*
84 *pachyderma*, planktic foraminiferal fauna distribution patterns, bulk sediment calcium
85 carbonate (CaCO_3) content and grain size distributions to interpret the paleoceanography. In
86 order to improve quantitative paleoceanographic reconstructions, we applied a transfer
87 function by Husum and Hald (2012) to the downcore planktic foraminiferal fauna
88 distribution, enabling reconstruction of summer subsurface temperatures ($\text{sSST}_{\text{Transfer}}$) (100 m
89 water depth). Furthermore, paleo-subsurface temperatures were calculated from Mg/Ca ratios
90 in *N. pachyderma* ($\text{sSST}_{\text{Mg/Ca}}$) which is an approach recently used in paleoceanographic
91 reconstructions in the eastern Fram Strait (Spielhagen et al., 2011; Aagaard-Sørensen et al.,
92 2013). Mg/Ca ratios of foraminifers primarily reflect water temperatures during test growth
93 (e.g., Elderfield and Ganssen, 2000), while salinity and pH are subordinate influences on test

94 Mg uptake (Nürnberg et al., 1996; Lea et al., 1999). In the present study we use the species
95 specific (*N. pachyderma*) Mg/Ca calibration by Kozdon et al. (2009) which enable
96 temperature reconstructions of subpolar and polar water masses ($>2.5^{\circ}\text{C}$). Foraminiferal test
97 fragmentation was used to assess the pre-analytical preservation state of carbonates (Le and
98 Shackleton, 1992; Pfuhl and Shackleton, 2004).

99 The investigated core site is situated under the axis of present-day inflow of Atlantic
100 Water close to the Arctic Front dividing Atlantic and Arctic water masses (Fig. 1A) and is
101 therefore well suited to record changes of the oceanic parameters within this main conduit of
102 heat and salt to the Arctic. The time period from 14,000 to 8,600 cal yr B.P. was chosen for
103 investigation as the rapid oceanographic changes previously documented across the Late
104 Glacial-Holocene boundary (e.g., Ebbesen et al., 2007; Hald et al., 2007) may be construed as
105 the most recent analogue to the rapid changes happening in the Arctic today (e.g. IPCC,
106 2007).

107

108 **2. Oceanographic setting**

109 The Fram Strait is a deep passage (2600 m) between Svalbard and Greenland that connects
110 the north-eastern North Atlantic to the Arctic Ocean (Fig. 1A). Warm and saline Atlantic
111 Water (T: 3 to 7°C; S: 34.9 to 35.2, Schauer et al., 2004; Walczowski et al., 2005) is
112 transported towards the Arctic Ocean via the West Spitsbergen Current, a meridional branch
113 of the North Atlantic Current (Loeng et al., 1997; Schauer et al., 2002) (Fig. 1A). The West
114 Spitsbergen Current is topographically steered along the western slope of Spitsbergen
115 through the eastern Fram Strait and into the Arctic Ocean (Blindheim and Rey, 2004;
116 Walczowski et al., 2005). This makes the eastern Fram Strait the main pathway for heat and

117 salt advection into the Arctic Ocean (Schauer and Beszczynska-Möller, 2009). At present the
118 water mass at the coring site is dominated by Atlantic Water occupying the upper 500 to 700
119 m of the water column below a ca. 25 m thick upper mixed layer (Fig. 1B). At ca. 78°N the
120 advected Atlantic Water begins to submerge (e.g., Aagaard and Carmack, 1989) and a major
121 part re-circulates in the Fram Strait creating a southward return flow, the Return Atlantic
122 Water (RAW) (T: >0°C; S: >34.90) (Bourke et al., 1988) (Fig. 1). North of Svalbard Atlantic
123 Water continues as a subsurface current into the Arctic Ocean northward as the Yermak
124 Slope Current (Manley, 1995) and eastward as the Svalbard Branch along the northern
125 continental slope of Svalbard (Aagaard et al., 1987; Manley, 1995) (Fig. 1A).

126 In the western part of the Fram Strait, the East Greenland Current occupies the upper
127 ca. 150 m of the water column carrying a cold, low salinity (T: 0 to -1.7°C; S: ca. 30 to 34)
128 polar water mass southward along with most (>90%) of the sea ice exported from the Arctic
129 Ocean (Woodgate et al., 1999; Rudels et al., 1999, 2005)(Fig. 1A). In the central Fram Strait
130 Polar and Atlantic Water mix and form Arctic Water masses with intermediate temperature
131 and salinity (e.g., Hop et al., 2006). The contrasting water masses are separated by transition
132 zones termed the Polar (separating Polar and Arctic water masses) and Arctic (separating
133 Arctic and Atlantic water masses) Fronts which roughly define the average summer sea ice
134 margin and the maximum limit of the winter sea ice margin, respectively (Swift and Aagaard,
135 1981; Hopkins, 1991) (Fig. 1A).

136

137 **3. Material and methods**

138 Kastenlot core MSM05/5-712-2 was retrieved from 1487 m water depth on the West
139 Spitsbergen Slope in the eastern Fram Strait (78°54.94' N, 06°46.04' E) during a cruise of the

140 RV *Maria S. Merian* in August 2007 (Fig. 1A). Water conductivity, temperature, and depth
141 were measured prior to coring (Fig. 1B). Proxy data from the 8.94 m long sediment core are
142 presented here at the core depth interval from 209 to 441cm. The interval is constrained by
143 four previously published accelerator mass spectrometry (AMS) radiocarbon date
144 measurements (Fig. 2) (Aagaard-Sørensen et al., 2013). Due to low abundance of planktic
145 foraminifera in the sediment below ca. 330 cm core depth it was only possible to collect
146 enough material for one AMS date. Therefore, the age model below this point is poorly
147 constrained. The AMS measurements were performed at the Leibniz Laboratory of Kiel
148 University, Germany, and Poznań Radiocarbon Laboratory, Poland (Table 1). Calibration of
149 the radiocarbon dates was performed using Calib version 6.0 (Reimer et al., 2004; Stuiver et
150 al., 2005) and the marine calibration curve Marine09 (Hughen et al., 2004; Reimer et al.,
151 2009). A total reservoir age of 551 ± 51 years was used. This value was reached using the
152 standard reservoir correction of 400 years and the modern reservoir age (ΔR) of 151 ± 51
153 years from the nearby Magdalenafjorden (Mangerud and Gulliksen, 1975; Mangerud et al.,
154 2006). The age model was established by linear interpolation between the calibrated
155 radiocarbon dates using the mean of the 2σ interval of highest probability as individual tie
156 points (Fig. 2, Table 1). In figures and text all dates will refer to calibrated years before
157 present, B.P. (present=1950). The Late Glacial - Holocene chronostratigraphic zones are
158 based on the most recent divisions defined on the basis of Greenland ice cores (Rasmussen et
159 al., 2006, 2007a; Steffensen et al., 2008; Walker et al., 2009): Bølling-Allerød interstadial
160 14,650 to 12,850 cal yr B.P., Younger Dryas 12,850 to 11,650 cal yr B.P. and Holocene
161 11,650 cal yr B.P. to present.

162 The lithology of the sediment core was visually described onboard after coring.
163 Sediment samples, at 6 cm-intervals, were freeze-dried and wet-sieved through 63 μm , 100
164 μm and 1 mm sieves. Dried sample fractions were weighed and used to determine the grain

165 size distribution. The >1 mm size fraction is considered as ice-rafted debris (IRD) (Fig. 3).
166 Total organic carbon (TOC) and total carbon (TC) were measured using a Leco CS 200
167 furnace at the University of Tromsø. The TC content (wt.%) was measured directly on bulk
168 sediment samples while the TOC content (wt.%) was measured on samples pre-treated with
169 HCl (10%) to remove CaCO₃ before combustion (1350°C). Subsequently bulk sediment
170 CaCO₃ content was calculated using the equation: $\text{CaCO}_3 = (\text{TC} - \text{TOC}) * 100 / 12$ (e.g., Knies et
171 al., 2003) (Fig. 3). CaCO₃ from 11,700 to 8,600 cal yr B.P. have previously been published in
172 Aagaard-Sørensen et al. (2013).

173 Approximately 300 planktic foraminifer specimens picked from the 100 µm-1 mm
174 size fraction were identified to species level and relative foraminiferal distributions and
175 fluxes were calculated (Fig. 3). Transfer function summer (July to September) sub sea surface
176 temperatures ($sSST_{\text{Transfer}}$) at 100 m water depth were reconstructed using the C.2 program
177 (ver 1.6) (Juggins, 2010) (Fig. 5). The applied transfer function is based on a training set
178 consisting of modern planktic foraminifera picked from the 100-1000 µm size fraction in
179 northern North Atlantic core top sediments and the Weighted Average Partial Least Square
180 (WAPLS) model (Husum and Hald, 2012). For the $sSST_{\text{Transfer}}$ reconstruction the three
181 component WAPLS model cross-validated by “jack knifing” was used. This model produced
182 low root mean squared error (RMSE) and low maximum bias in conjunction with relatively
183 high correlation between observed and estimated values (r^2) (e.g., Ter Braak and
184 Juggins, 1993; Birks, 1995; Husum and Hald, 2012) (Table 2). Planktic foraminiferal test
185 fragments picked from the 100-1000 µm size fraction were counted and a fragmentation
186 index was calculated using the equation of Pfuhl and Shackleton (2004): $\text{Fragmentation (\%)} = \# \text{ fragments} * 100 / (\# \text{ tests} * (1/3 * \# \text{ fragments}))$ (Fig. 4). Visual inspection of samples
187 and the structure of the broken fragments lead us to assume that tests broke down into
188

189 multiple fragments. Therefore, we applied a fragment-divisor of 3 (Le and Shackleton, 1992;
190 Pfuhl and Shackleton, 2004).

191 Stable isotope measurements were performed at the GEOMAR Helmholtz Centre for
192 Ocean Research, Kiel using a Finnigan MAT 253 mass spectrometer (reproducibility of
193 $\pm 0.03\text{‰}$ for $\delta^{13}\text{C}$ and $\pm 0.06\text{‰}$ for $\delta^{18}\text{O}$) and a Kiel IV Carbonate Preparation Device. Stable
194 oxygen and carbon isotopic ratios were measured on 20 to 30 tests of the planktic foraminifer
195 species *N. pachyderma* picked from the 125-250 μm sieve size fraction (Fig. 4). In order to
196 use well-mixed aliquots the tests were crunched and mingled. All measurements were
197 calibrated to Vienna Pee Dee Belemnite (VPDB) standard (NBS 19). Measurements were
198 carried out at 1 cm-intervals apart from at 373-380 cm core depth where scarcity of
199 foraminiferal fauna prevented analysis. The $\delta^{18}\text{O}$ isotope record was not corrected for vital
200 effects (Jonkers et al., 2010) but for the ice volume effect (Fairbanks, 1989) (Fig. 4). For
201 comparison we also show the uncorrected $\delta^{18}\text{O}$ isotope record and the record corrected for
202 the ice volume effect according to the ICE-5G(VM2) model (Peltier and Fairbanks, 2006)
203 (Fig. 4).

204 Trace element analysis was performed every 3 cm on ca. 50 tests of *N. pachyderma*.
205 Specimens were picked at a narrow size fraction (225-290 μm) to minimize possible size-
206 dependent bias (Elderfield et al., 2002). Foraminiferal tests were gently crushed between
207 glass plates to expose all test chambers to the reductive (anhydrous hydrazine) and oxidative
208 (H_2O_2) cleaning procedures following Boyle and Keigwin (1985) and Boyle and Rosenthal
209 (1996). Cleaned samples were analyzed for Mg/Ca, Mn/Ca and Fe/Ca by magnetic-sector
210 single-collector ICP-MS, on a Thermo-Finnigan Element2 at INSTAAR, University of
211 Colorado (Marchitto, 2006). Standards, with minor and trace element concentrations that
212 mimic the typical oceanic ranges of foraminifera, were prepared gravimetrically using 1000
213 ± 3 mg L11 stock solutions from High-Purity Standards and SPEX CertiPrep and made at 5

214 mM (200 ppm) Ca in 2% (v/v) HNO₃ (Fisher Optima) (Marchitto, 2006). ²⁶Mg, ⁵⁵Mn and ⁵⁶Fe
215 were measured in analog mode and ratioed to ⁴³Ca in low (Mg, Mn) and medium (Fe)
216 resolution (Marchitto, 2006). Replicate analysis was performed for every ca. 20th sample. The
217 average Mg/Ca reproducibility of sample splits was ±0.049 mmol/mol (n=3) which is below
218 4% difference between the average and recorded duplicate values. The system has long-term
219 1σ precisions for Mg/Ca of 0.5% (Marchitto, 2006). Samples with >100 μmol/mol in regards
220 to Fe/Ca (n=3) or Mn/Ca (n=4) were omitted (Fig. 4) due to possible contamination by
221 detrital material or secondary diagenetic coatings, which could lead to biased Mg/Ca values
222 (Barker et al., 2003). Likewise one sample with <5μg CaCO₃ postcleaning mass was omitted
223 (Marchitto, 2006) (Fig. 4). Foraminiferal Mg/Ca ratios measured from 11,700 to 8,600 cal yr
224 B.P. have previously been published in Aagaard-Sørensen et al. (2013).

225 Mg/Ca thermometry was used to obtain quantitative reconstructions of sea surface
226 temperatures (SST_{Mg/Ca}) by applying a species-specific (*N. pachyderma*) temperature
227 equation (Fig. 5). The applied linear equation of Kozdon et al. (2009) is based on cross
228 calibrated Mg/Ca and δ^{44/40}Ca proxy signals of *N. pachyderma* in Holocene core top samples
229 from the Nordic Seas: Mg/Ca (mmol/mol) = 0.13(±0.037) * T (°C) + 0.35(±0.17), where
230 numbers in parentheses show the error margins. Despite exponential thermodynamic control
231 on Mg uptake in calcareous foraminiferal tests this equation assumes that linearity adequately
232 depicts the Mg uptake/temperature relation in the narrow temperature range inhabited by *N.*
233 *pachyderma* and works for reconstructed temperatures above ca. 2.5°C (Mg/Ca >0.74
234 mmol/mol) (Kozdon et al., 2009). However, when temperatures are lower than 2.5°C, which
235 is typically associated with salinities less than 34.5, the method loses its precision (Kozdon et
236 al., 2009). The reductive cleaning of foraminiferal calcite utilized in the present study has
237 been shown to potentially decrease the Mg/Ca ratio by up to 15 % (Barker et al., 2003). For
238 comparison and in order to evaluate the potential impact of the reductive cleaning on the

239 temperature reconstructions we artificially increased the Mg/Ca ratio by 15% before
240 recalculating $sSST_{Mg/Ca}$ (Fig. 5) (See Discussion).

241

242 **4. Results**

243 **4.1 Lithology**

244 The sediment comprises two different lithological units. Unit B (14,000 - 10,300 cal
245 yr B.P.; 442 – 323 cm) holds dark olive grey monosulphide-rich silty clay with ca. 1 cm thick
246 olive black laminations. The overlying unit A (10,300 - 8,600 cal yr B.P.; 323 – 208 cm)
247 holds dark homogeneous olive grey, monosulphide-rich bioturbated silty clay (Fig. 3). The
248 sedimentation rate of unit B is 32 cm/kyr vs. 59 to 108 cm/kyr in unit A (Fig. 2). Unit B holds
249 2 to 7 wt.% material $>63 \mu\text{m}$ (Fig. 3) and up to 4 wt.% material $>1 \text{ mm}$ (IRD) (Fig. 3). Two
250 periods at 13,300 - 12,200 and 11,500 - 10,900 cal yr B.P. hold higher concentrations of
251 material $>1 \text{ mm}$ averaging 2 wt.%. Unit A is very fine grained with ~ 98 wt.% of material <63
252 μm . A diatom rich layer was identified at 10,500 - 9,800 cal yr B.P. (329 – 281 cm) (Fig. 3).
253 CaCO_3 values were low, averaging 4.5 wt.% prior to 10,000 cal yr B.P., and increased from
254 ca. 5 to 13 wt.% during the succeeding 1400 years (Fig. 3).

255 **4.2 Planktic foraminifera, Transfer function sSST and shell fragmentation**

256 Polar and subpolar species *N. pachyderma* and *Turborotalita quinqueloba* dominate
257 the planktic foraminiferal fauna (Figs. 3, 5). The clear dominance of these two species is
258 characteristic for Arctic marine environments (e.g., Johannessen et al., 1994; Carstens et al.,
259 1997; Volkman, 2000). Between 14,000 to 10,500 cal yr B.P. the faunal fluxes were low
260 (ca. 2 specimens/ $\text{cm}^2 \cdot \text{yr}$) (Fig. 3) and the foraminiferal fauna was dominated by *N.*
261 *pachyderma* ($>86\%$) with *T. quinqueloba* and *Neogloboquadrina incompta* (formerly denoted
262 *N. pachyderma* (dextral coiling); Darling et al., 2006) as secondary species ($<10\%$).

263 Between 10,500 and 10,100 cal yr B.P. *T. quinqueloba* became the dominant species
264 coincident with increasing planktic foraminifer fluxes (Figs. 3, 5). Relative abundance of *T.*
265 *quinqueloba* averaged 63% from 10,100 to 8,600 cal yr B.P., while overall planktic fluxes
266 remained high averaging 50 specimens/cm²*yr. In this period *N. pachyderma* constituted ca.
267 25% of the total fauna except at ca. 8,800 cal yr B.P. where it reached 68% (Fig. 3). After
268 10,500 cal yr B.P. other commonly found species were *N. incompta* (ca. 6%), *Globigerina*
269 *bulloides* (>3.1%), *Globigerinita calida* (>2.4%), *Globigerinita glutinata* (>2.2%) and
270 *Globigerinita uvula* (up to ca. 10%). *Globigerinita uvula* was primarily present from 10,000
271 to 9,200 cal yr B.P. with a peak in relative abundance of ca. 10% at 9,900 cal yr B.P. (Fig. 3).

272 The sub sea surface temperatures obtained by transfer functions (sSST_{Transfer}) show
273 low values ranging from 1.9 to 2.7°C (average 2.1°C) prior to ca. 10,300 cal yr B.P. A rapid
274 temperature increase occurred between ca. 10,500 to 10,100 cal yr B.P. followed by relatively
275 high, but slowly declining, temperatures ranging from 3.3 to 6.5°C (average 5.5°C) from
276 10,100 to 8,600 cal yr B.P. (Fig 5).

277 Low fragmentation of planktic foraminiferal tests, averaging 3.4% was found from
278 14,000 to 12,850 cal yr B.P. (Fig. 4). From 12,850 to 10,500 cal yr B.P. values were
279 generally higher, averaging 13.8% with values exceeding 45% around 12,000 cal yr B.P.
280 From 10,500 to 8,600 cal yr B.P. values averaged 8.9% with the highest value (32.2%) at ca.
281 8,800 cal yr B.P.

282 **4.3 Stable isotopes**

283 *N. pachyderma* δ¹³C values show two local maxima at ca. 12,700 (0.25‰) and ca.
284 8,800 cal yr B.P. (0.45‰) in addition to a broad maximum at 11,600 - 10,200 cal yr B.P.
285 (average ca. 0.25‰) (Fig. 4). Low average values of ca. 0.1‰ were found at 10,200 - 9,000
286 cal yr B.P.

287 *N. pachyderma* $\delta^{18}\text{O}$ values increased from ca. 3.1‰ at 14,000 to 3.7‰ at 12,100 cal
288 yr B.P. (Fig. 4). Low average values (around 2.9‰) from 11,900 to 11,500 cal yr B.P. were
289 followed by an increase leading to values of ca. 3.5‰ at ca. 11,400 - 11,200 cal yr B.P. From
290 ca. 11,000 cal yr B.P. the values declined and reached a minimum at ca. 9,500 cal yr B.P. (ca.
291 2.5‰) with one marked excursion towards heavier values at ca. 10,000 cal yr B.P. A slightly
292 increasing trend was observed after 9,500 cal yr B.P. (Fig. 4).

293 **4.5 Mg/Ca ratios and reconstructed SST_{Mg/Ca}**

294 The Mg/Ca ratios show values ranging from ca. 0.6 to 1.07 mmol/mol in the analysed
295 interval. An overall slightly increasing trend is noticed towards the youngest part of the
296 record (Fig. 4). Highest values were found at ca. 13,200, ca. 11,600 and after ca. 10,300 cal
297 yr B.P. Average Mg/Ca values were 0.81 mmol/mol prior to and 0.86 mmol/mol after 10,300
298 cal yr B.P. (Fig. 4). Sea surface temperatures based on Mg/Ca ratios (SST_{Mg/Ca}) (equation by
299 Kozdon et al., 2009) show temperature fluctuations between ca. 2.5 and 5.5°C throughout the
300 record. Average temperatures are 3.5°C before and 4.0°C after 10,300 cal yr B.P. (Fig. 5).

301

302 **5. Discussion**

303 **5.1. Evaluation of proxies**

304 In the present study paleoceanography and paleo-water temperature in the eastern Fram Strait
305 across the Late Glacial – early Holocene transition was estimated and reconstructed via
306 Mg/Ca ratios and stable isotopes in *N. pachyderma*, fossil foraminiferal fauna, and chemical
307 and physical properties of the sediment (Figs. 3, 4, 5).

308 *N. pachyderma* reflects/records ambient water properties at varying depths below the
309 thermocline from ca. 25-250 m water depth (e.g., Simstich et al., 2003) where modern day
310 temperature (August 2007) is 3.7-6.7°C and salinity is ~35.1 (Fig. 1B). Therefore water mass

311 property reconstruction, via stable isotopes and trace elements on this biotic carrier, do not
312 reflect actual sea surface conditions and in the following discussion $SST_{Mg/Ca}$ is denoted as
313 sub SST ($sSST_{Mg/Ca}$).

314 Low fluxes of planktic foraminifera and dominance of *N. pachyderma* (>90%)
315 characterized the period prior to 10,300 cal yr B.P. (Fig. 3). From 14,000 to 12,800 cal yr
316 B.P., the fragmentation of tests was low indicating good preservation (Fig. 4). From ca.
317 12,800 to 10,500 cal yr B.P., fragmentation was elevated, indicating reduced test preservation
318 (Le and Shackleton, 1992; Pfuhl and Shackleton, 2004) (Fig. 4). Highest fragmentation was
319 observed around 12,000 cal yr B.P. indicating that the test preservation was at a minimum for
320 the record as a whole (Fig. 4). This may have contributed to the scarcity of planktic
321 foraminifera observed at this time, leading to a deficient number of specimens for stable
322 isotope analysis (Figs. 3, 4). In addition, the only sample from the trace element dataset
323 omitted due to low post-cleaning mass refers to this approximate age potentially suggesting
324 that the foraminiferal calcite had been exposed to pre-analytic dissolution (Fig. 4).
325 Barker et al. (2005) found that low test preservation (i.e., enhanced dissolution) generally
326 causes a lowering of Mg/Ca ratios in foraminiferal calcite and thereby lowers the temperature
327 estimates. Furthermore, reduced preservation of planktic foraminiferal tests could facilitate
328 selective removal/destruction of dissolution-prone species (e.g. *T. quinqueloba*, *G. uvula*)
329 from the fossil assemblages (e.g., Conan et al., 2002). This can result in a relative increase in
330 dissolution-resistant species (e.g., *N. pachyderma*) and modification of the faunal
331 composition (e.g., Conan et al., 2002; Zamelczyk et al., 2012). Despite the difference in
332 fragmentation between the two periods (14,000 - 12,800 and 12,800 - 10,500 cal yr B.P.), no
333 distinct change in Mg/Ca ratios or fauna composition was observed (Figs. 3, 4). This
334 indicates that the preservation of tests had little or no impact on the present $sSST$
335 reconstructions (Fig. 5).

336 Barker et al. (2003) found that the reductive cleaning of foraminiferal calcite used in
337 the trace element analysis (see Material and methods section for details) potentially decreases
338 the Mg/Ca by up to 15%, thereby lowering the reconstructed $sSST_{Mg/Ca}$. Prior to ~10,500 cal
339 yr B.P. the $sSST_{Mg/Ca}$ reconstruction shows temperatures that were larger than the $sSST_{Transfer}$
340 (Fig. 5). By artificially increasing the Mg/Ca ratios by 15% the resulting temperature
341 estimates become even higher (~1°C) tentatively suggesting that the reductive cleaning
342 procedure is of minor importance for the reconstructed $sSST_{Mg/Ca}$ (Fig. 5). This tentative
343 conclusion is supported by earlier studies from the same core site that include Mg/Ca
344 temperature reconstructions based on the same biotic carrier (*N. pachyderma*) and the same
345 exact cleaning procedure as the material in the present study (Spielhagen et al., 2011;
346 Aagaard-Sørensen et al., 2013). Spielhagen et al. (2011) showed minor difference between
347 measured modern water temperatures and core-top $sSST_{Transfer}$ and $sSST_{Mg/Ca}$, while
348 comparison of reconstructed summer $sSST_{Transfer}$ published by Werner et al. (2013) with
349 $sSST_{Mg/Ca}$ from 8,800 cal yr B.P. to the present within the upper part of core MSM05/5-712-2
350 showed similar temperature ranges (Aagaard-Sørensen et al., 2013).

351 Prior to the Holocene Arctic surface water masses were spread far to the south and
352 east in the northeastern North Atlantic (Koç et al., 1993) and thus could have influenced the
353 coring site. Kozdon et al. (2009) found that modern Mg/Ca-derived temperature estimates
354 from the cold (<3°C), low saline Arctic domain and Polar waters in the Nordic Seas were
355 higher than corresponding maximum SST in the areas. Therefore it must be cautioned that
356 our $sSST_{Mg/Ca}$ reconstruction, with values averaging 3.6°C prior to ~10,500 cal yr B.P., could
357 potentially be erroneously high (Fig. 5). The fact that the average value of $sSST_{Mg/Ca}$ is higher
358 than average value of $sSST_{Transfer}$ in this period could alternatively suggest that $sSST_{Mg/Ca}$
359 measured on *N. pachyderma* reflect another water depth than the water depth represented by
360 the $sSST_{Transfer}$ reconstruction. While the $sSST_{Transfer}$ values reflect temperatures at 100 m

361 water depth (Husum and Hald, 2012), $sSST_{Mg/Ca}$ may reflect shallower water depths where
362 temperatures presumably were higher (Fig. 5). The depth habitats of *N. pachyderma* have
363 been found to be shallower than 100 m water depth in Arctic areas influenced by sea ice
364 (Volkman, 2000; Simstich et al., 2003; Pados and Spielhagen, in press). These findings may
365 therefore suggest that the calcification depth of *N. pachyderma* was shallower than 100 m in
366 the period from ca. 14,000 - 10,500 cal yr B.P.

367 In addition, the observed difference between the $sSST_{Transfer}$ and $sSST_{Mg/Ca}$
368 reconstructions may be linked to seasonal differences between the proxy approaches. In the
369 Arctic region peak fluxes of planktic foraminifera are reported during summer at sea ice
370 margins with high primary production (Carstens et al., 1997). Thus, during periods with ice-
371 cover and abundant icebergs, like the earliest part of our record (see section 5.2 for further
372 discussion), the primary production bloom and calcification season of *N. pachyderma* (i.e.
373 $sSST_{Mg/Ca}$) was possibly delayed/shifted towards late summer where water temperatures
374 potentially were higher than the average summer (July to September) temperatures
375 reconstructed by the transfer functions ($sSST_{Transfer}$) (Fig. 5).

376 After ~10,300 cal yr B.P., the subpolar species *T. quinqueloba* became the dominating
377 species (average >60%) and the flux of planktic foraminifera increased markedly (Figs. 3, 5).
378 The species *T. quinqueloba* is prone to dissolution, so when it is found in high numbers
379 together with relatively low test fragmentation, the test preservation is good (Figs. 3, 4).
380 Average $sSST_{Mg/Ca}$ values from 10,300 to 8,600 cal yr B.P. were ~0.5°C higher compared to
381 the previous period (14,000 to 10,300 cal yr B.P.) (Fig. 5). This finding is in contrast to
382 summer $SST_{Transfer}$ reconstructions from the Norwegian–Svalbard margin where average
383 values at 10 m water depth rose ~5°C in the early Holocene (Hald and Aspeli, 1997; Ebbesen
384 et al., 2007; Hald et al., 2007). However, when applying the transfer function of Husum &

385 Hald (2012) reconstructing temperatures for subsurface water masses at 100 m water depth
386 for the new and the abovementioned records a temperature increase of $\sim 3.5^{\circ}\text{C}$ is generated in
387 this period (Fig. 5). During most of the early Holocene the $\text{sSST}_{\text{Transfer}}$ (100 m water depth)
388 remains higher than $\text{sSST}_{\text{Mg/Ca}}$ suggesting that *N. pachyderma* may have responded to
389 oceanographic changes by descending in the water column and calcifying at depths below
390 100 m (Fig. 5).

391 Kozdon et al. (2009) suggest that *N. pachyderma* throughout its life cycle is actively
392 or inactively linked to an isopycnal layer with densities (σ_t) ranging from 27.7 to 27.8,
393 corresponding to a depth of 60-110 meters with the present day temperature and salinity
394 profile (Fig. 1B). As a consequence the species prefers gradually deeper habitats with
395 increasing temperatures, thus counterbalancing absolute sea surface temperature variations
396 (Kozdon et al., 2009). This suggestion could explain the relatively narrow (ca. 2.5 to 5.5°C)
397 temperature range depicted by the $\text{sSST}_{\text{Mg/Ca}}$ reconstruction (Fig 5) during a period and in a
398 region where significant oceanographic changes have been described (e.g. Koç et al., 2002;
399 Hald et al., 2007; Rasmussen et al., 2007b; Ebbesen et al., 2007; Farmer et al., 2008).
400 However, numerous studies showed that factors such as sea ice cover, proximity to sea ice
401 margins and oceanic fronts, water column stratification, water mass distribution, and food
402 availability also can influence the habitat depth of planktic foraminifera (Carstens et al.,
403 1997; Volkman, 2000; Simstich et al., 2003; Jonkers et al., 2010). Furthermore, a recent
404 comparison of plankton tow and physical oceanography data from the Fram Strait has shown
405 that the link of the *N. pachyderma* habitat to a narrow isopycnal band is less distinct in ice-
406 covered than in ice-free waters (Pados and Spielhagen, in press).

407 As mentioned earlier seasonal changes in calcification of *N. pachyderma* can also
408 impact the $\text{sSST}_{\text{Mg/Ca}}$ values. In the central Irminger Sea ($\sim 59^{\circ}\text{N}$) Jonkers et al. (2010) reports
409 largest fluxes of *N. pachyderma* during spring and in late summer, when SSTs can be

410 relatively variable. The strongly fluctuating $sSST_{Mg/Ca}$ in the early Holocene could therefore
411 also result from shifts in calcification season of *N. pachyderma* (Fig. 5). Moreover, laboratory
412 experiments have shown fastest growth rates for *N. pachyderma* at temperatures around 5°C,
413 and growth rates progressively slowing with increasing temperatures (Lombard et al., 2009).
414 This suggests that *N. pachyderma*, during the early Holocene, may have calcified earlier in
415 the season, when temperatures were more favorable (Fig. 5).

416

417 **5.2. Paleoenvironmental reconstruction and correlation**

418 **5.2.1 Period: 14,000 to 12,850 cal yr B.P. (Bølling-Allerød interstadial)**

419 The Bølling-Allerød interstadial is characterized by a planktic foraminiferal fauna
420 assemblage dominated by *N. pachyderma* (>90%) that together with low faunal fluxes, and
421 low $CaCO_3$ indicate polar conditions with reduced productivity (Johannessen et al., 1994)
422 (Fig. 3). This is supported by the reconstructed $sSST_{Transfer}$ showing cold summer conditions
423 (averaging 2°C) throughout the period (Fig. 5). Previous studies have documented relatively
424 strong meridional advection of Atlantic Water through the Eastern Fram Strait during
425 Bølling-Allerød (Birgel and Hass, 2004; Ślubowska et al., 2005; Rasmussen et al., 2007b).
426 The concurrent higher $sSST_{Mg/Ca}$ (averaging 3.5°C) could therefore suggest that *N.*
427 *pachyderma* calcified in chilled Atlantic Water masses, likely at water depths above 100 m
428 and during peak (late) summer conditions during this period (Fig. 5) (see 5.1. Evaluation of
429 proxies for further discussion).

430 Relatively large amounts of sand (>63 μm) and IRD (>1 mm) show influence of
431 icebergs and coastal sea ice at the core site, while low $\delta^{13}C$ values indicate stratification (e.g.,
432 Spielhagen and Erlenkeuser, 1994) which probably was caused by sea ice and melt water
433 (Figs. 3, 4). Low $\delta^{13}C$ values are also recorded further south at ca. 77°N in the Fram Strait

434 (Ebbesen et al., 2007) and at ca. 75°N on the Barents Sea slope (Sarthein et al., 2003) during
435 this period. The two IRD peaks found at around 13,600 and 13,100 cal yr B.P. may correlate
436 with cold inter Bølling-Allerød periods recorded in the NGRIP ice core (Rasmussen et al.,
437 2006) (Fig. 5). This suggests increased calving, transport and melting of icebergs or sea ice
438 during periods with cold atmospheric conditions. Elevated coarse fraction and IRD
439 concentrations have previously been reported during Bølling-Allerød on the West
440 Spitsbergen Slope (Rasmussen et al., 2007b; Ebbesen et al., 2007; Jessen et al., 2010), while
441 concurrent decreased IRD concentrations were ascribed to more prolonged and severe sea ice
442 conditions north of Svalbard under the axis of the Svalbard Branch (Fig. 1) (Koç et al., 2002;
443 Ślubowska et al., 2005).

444 **5.2.2 Period: 12,850 to 11,650 cal yr B.P. (Younger Dryas)**

445 During Younger Dryas the planktic foraminiferal assemblage was dominated by *N.*
446 *pachyderma* (>90%) indicating cold polar sea surface conditions (Fig. 3) (Johannessen et al.,
447 1994). The sediment CaCO₃ content and flux of planktic foraminifera were also low, pointing
448 at continued reduced primary production in the water column (Fig. 3). Slightly lowered sand
449 (>63µm) and IRD concentrations during this period (Fig. 3) may be interpreted as either
450 increased sea ice cover that suppressed iceberg transport, decreased glacier calving on
451 Svalbard and/or surface waters that were too cold to allow melting of icebergs/sea ice. Low
452 IRD concentrations west and north of Svalbard have previously been linked to prolonged sea
453 ice coverage and reduced iceberg transport during the Younger Dryas (Koç et al., 2002;
454 Wollenburg et al., 2004; Ebbesen et al., 2007; Ślubowska-Woldengen et al., 2007). Cold
455 conditions during the Younger Dryas stadial have been documented in numerous proxy
456 records in the Arctic region including ice core records (e.g. Rasmussen et al., 2006) (Fig. 5),
457 terrestrial proxy records (e.g. Landvik et al., 1998), and marine proxy records (e.g. Koç et al.,
458 1993; Ślubowska-Woldengen et al., 2007; 2008).

459 During the early part of the Younger Dryas (~12.800 – 12.000 cal yr B.P.) continued
460 low $sSST_{transfer}$ (averaging 2°C) and lowered $sSST_{Mg/Ca}$ (averaging 3.2°C) combined with high
461 $\delta^{18}O$ show that cold summer conditions prevailed while chilled Atlantic Water remained
462 present during the foraminiferal growth season (Figs. 4, 5). This is supported by Rasmussen
463 et al. (2007b) who found that subsurface advection of Atlantic Water to the Fram Strait
464 continued and that the water mass was colder and probably less saline than during the
465 preceding Bølling-Allerød.

466 Increased freshwater injections and sea ice expansion in the Nordic Seas have
467 previously been identified and linked to hampered meridional overturning circulation during
468 the Younger Dryas (e.g., Broecker et al., 1989; Koç et al., 1993; Sarnthein et al., 1995; Hald
469 and Aspeli, 1997; Jennings et al., 2006; Bradley and England, 2008). The almost
470 foraminiferal-barren/high fragmentation interval (~12,100 to 11,900 cal yr B.P) and
471 subsequent low $\delta^{18}O$ values may indicate increased freshwater influence in the surface and
472 subsurface water mass at ca. 12,100 to 11,500 cal yr B.P. (Fig. 3, 4). However, following the
473 period with high fragmentation/low test preservation at the end of Younger Dryas and into
474 the Preboreal (11,900 - 11,500 cal yr B.P.) the somewhat elevated average $sSST_{Mg/Ca}$ (~4°C)
475 could alternatively suggest that the lowered $\delta^{18}O$ may be interpreted as a temperature increase
476 during the season and/or at the depth of *N. pachyderma* calcification (Figs. 4, 5).

477 **5.2.3 Period: 11,650 to 8,600 cal yr B.P. (Early Holocene)**

478 In the earliest part of the Holocene, until ca. 10,500 cal yr B.P., cold surface water conditions,
479 with a low flux foraminiferal fauna dominated by *N. pachyderma*, continued (Fig. 3). The
480 sand and IRD content rose slightly, showing a continued, possibly enhanced, influence of sea
481 ice and iceberg melting (Fig. 3). Similar, cold sea and sea ice/iceberg-influenced surface
482 conditions are recorded both west and north of Svalbard during this period (Ebbesen et al.,

483 2007; Ślubowska-Woldengen et al., 2007). The marked IRD peak observed at ca. 11,000 cal
484 yr B.P. (Fig. 3) correlates to rapid ice retreat on Svalbard and in western Barents Sea
485 (Landvik et al., 1998). Despite the apparent continued influence from sea ice and icebergs in
486 the area, the average $\delta^{13}\text{C}$ value was higher than during Younger Dryas, suggesting improved
487 ventilation of the water mass (cf. Spielhagen and Erlenkeuser, 1994). Alternatively the higher
488 $\delta^{13}\text{C}$ values could reflect increased primary production (e.g., Katz et al., 2010) in the surface
489 waters, which is tentatively supported by the slightly elevated planktic foraminiferal fluxes
490 (Figs. 3, 4).

491 After ~11,500 cal yr B.P., the $\delta^{18}\text{O}$ values steadily decreased until ~9,700 cal yr B.P.
492 indicating gradually lowered salinity and/or increased temperatures within the sub surface
493 water mass (Fig. 4). The reconstructed $\text{sSST}_{\text{Mg/Ca}}$, although strongly fluctuating, show a
494 concurrent increasing trend, supporting the latter interpretation of the $\delta^{18}\text{O}$ signal (Fig. 5). A
495 similar declining $\delta^{18}\text{O}$ trend is also observed on the south-western Svalbard slope (76°N)
496 (Rasmussen et al., 2007b) and on the western Barents Sea shelf (75°N) (Sarnthein et al.,
497 2003) at this time, indicating increasing northward heat advection via the Norwegian Atlantic
498 Current (NwAC), suggested by Risebrobakken et al. (2011) to have culminated at ~10,000
499 cal yr B.P.

500 The generally decreasing $\delta^{18}\text{O}$ trend is punctuated by several excursions towards
501 heavier values with most pronounced increases $>0.5\text{‰}$ observed at ~11,300 and 10,000 cal yr
502 B.P. (Fig. 4) that are likely related to short term regional cooling events. The cooling at
503 ~11,300 cal yr B.P. correlates to the Preboreal Oscillation cooling event, which has been
504 observed in various marine proxy records in the Fram Strait, Nordic Seas and in northern
505 Norway (e.g., Björck et al., 1997; Hald and Hagen, 1998; Husum and Hald, 2002; Ślubowska
506 et al., 2005) and in Greenland ice cores (Rasmussen et al., 2006, 2007a) (Fig. 5) while the

507 latter (~10,000 cal yr B.P) approximately correlates to the so-called 9.95 ka anomaly
508 (presented on the b2k scale) in the Greenland ice cores (Rasmussen et al., 2007a). However,
509 the lower resolution of the $sSST_{Mg/Ca}$ and $sSST_{Transfer}$ records does not merit confirmation that
510 these $\delta^{18}O$ excursions represent coolings (Figs. 4, 5).

511 The transition from the cold Younger Dryas stadial to the warm Holocene interglacial
512 has been recorded in Greenland ice core records at 11,650 cal yr B.P. (Rasmussen et al.,
513 2006; Walker et al., 2009) (Fig. 5). Associated with a northward displacement of the Arctic
514 Front, separating Arctic and Atlantic water masses, Hald et al. (2007) showed that surface
515 water masses in the eastern part of the Nordic Seas experienced a time-transgressive
516 transition from cold Late Glacial into warm Holocene sea surface conditions. The transition
517 happened at ca. 11,800 cal yr B.P. at 60°N and at ca. 10,500 cal yr B.P. at 77°N (Hald et al.,
518 2007 and ref therein). In areas north of 71°N the rise in transfer function-reconstructed SST
519 was primarily driven by a rapid increase in *T. quinqueloba* abundances (Hald and Aspel, 2007;
520 1997; Ebbesen et al., 2007; Hald et al., 2007) (Fig. 5). A rapid increase in abundance and flux
521 of *T. quinqueloba* can also be observed in our record, translating into a $sSST_{Transfer}$ increase of
522 ~3.5°C between 10,500 to 10,200 cal yr B.P. (Fig. 5). The timing of this rapid increase is in
523 accordance with the northward delay in onset of warmer surface conditions observed by Hald
524 et al. (2007) in the Nordic Seas. The delayed Holocene oceanic warming at high latitudes has
525 been attributed to the lingering impact of a cold water and sea ice pool in high Arctic settings
526 (Hald et al., 2007) which at our core site is expressed as coarser and more IRD laden
527 sediment deposited before ~10,500 cal yr B.P. (Fig. 3). The delayed warming has furthermore
528 been connected to high albedo caused by sea ice/snow cover and the related asymmetry of
529 atmospheric and oceanic circulation patterns (e.g., Kaufman et al., 2004).

530

531 The faunal transition was also associated with increased abundance and flux of the relatively
532 thermophile species *N. incompta* and *G. bulloides* indicating stronger influence from Atlantic
533 Water (Johannessen et al., 1994; Risebrobakken et al., 2011) in the eastern Fram Strait with
534 summer sSST_{Transfer} averaging ca. 5.5°C after ca. 10,300 cal yr B.P. (Figs. 3, 5). This change
535 coincides with markedly increased bottom current speeds in the northern Fram Strait (Birgel
536 and Hass, 2004) and strong inflow of Atlantic Water to the western and northern Svalbard
537 margin and fjords (Ślubowska-Woldengen et al., 2007; Skirbekk et al., 2010). The increased
538 Atlantic Water influx likely also led to cessation of iceberg transport, as indicated by absence
539 of IRD in our record after ca. 10,500 cal yr B.P. (Fig. 4B) which occurred almost
540 concurrently all along the West Spitsbergen Slope (Ebbesen et al., 2007; Rasmussen et al.,
541 2007b; Jessen et al., 2010).

542 The high flux and abundance of *T. quinqueloba* combined with generally increased
543 planktic foraminiferal fluxes and increasing sediment CaCO₃ content after ca. 10,300 cal yr
544 B.P. (Figs. 3, 5) shows that the Arctic Front and associated productive water masses probably
545 were situated close to the west Spitsbergen and Barents Sea slopes during the early Holocene
546 (Johannessen et al., 1994; Hald and Aspeli, 1997; Sarnthein et al., 2003; Ebbesen et al.,
547 2007). A pronounced peak in relative percentage and flux of *G. uvula*, a species that tolerates
548 lowered salinities and is found in cold productive surface waters near oceanic fronts
549 (Boltovskoy et al., 1996; Husum and Hald, 2004) is observed at ca. 10,000 to 9,300 cal yr
550 B.P. (Fig. 3). High concentration of diatom frustules have been found in the sediments all
551 along the West Spitsbergen Slope during the Early Holocene (Jessen et al., 2010) and is also
552 observed in the present record between ca. 10,500 to 9,800 cal yr B.P. (Fig. 3). In conjunction
553 with the presence of *G. uvula* this finding further indicates the presence of highly productive
554 water masses in the area which likely was linked to proximity of the Arctic Front during the
555 Early Holocene (Fig. 3).

556 After ca. 10,000 cal yr B.P. minimum $\delta^{18}\text{O}$ values together with somewhat elevated
557 average $\text{sSST}_{\text{Mg/Ca}}$ ($\sim 4^\circ\text{C}$) indicate that ambient water temperatures during *N. pachyderma*
558 test formation were the highest within the record (Figs. 4, 5), while relatively low $\delta^{13}\text{C}$ values
559 indicate that the water mass was less ventilated at ca. 10,500 to 9,000 cal yr B.P. (Fig. 4). The
560 warm conditions are also reflected by the high summer $\text{sSST}_{\text{Transfer}}$ (averaging $\sim 5.5^\circ\text{C}$)
561 recorded after 10,300 cal yr B.P. (Fig. 5). However, the $\text{sSST}_{\text{Transfer}}$ shows a trend toward
562 lower average summer temperatures after ca. 9,300 cal yr B.P. with a pronounced low
563 ($\sim 3.3^\circ\text{C}$) observed at ca. 8,800 cal yr B.P. (Figs. 4, 5), which may indicate the first step
564 towards the termination of high early Holocene summer surface temperatures also observed
565 in other records along the Barents Sea and West Spitsbergen slopes at approximately this
566 time (Sarnthein et al., 2003; Ebbesen et al., 2007).

567

568 **6. Conclusions**

569 The paleoceanographic reconstruction shows polar surface conditions with faunal
570 dominance of the polar species *N. pachyderma* in the study area from 14,000 to 10,300 cal yr
571 B.P. The area was influenced by extensive sea ice cover and iceberg transport with low
572 foraminiferal fluxes and low primary production. The resulting summer $\text{sSST}_{\text{Transfer}}$ ranged
573 from 1.9 to 2.7°C with an average of 2.1°C . However, the quantitative reconstructions based
574 on Mg/Ca ratios show water temperatures ranging from 1.9 to 5.2°C with an average of 3.5°C
575 during this time interval pointing to warmer conditions in those water masses where *N.*
576 *pachyderma* calcified its test. This could possibly be ascribed to subsurface advection of
577 Atlantic Water masses combined with shifts in the calcification season and/or the habitat
578 depth of the foraminifera.

579 After ca. 10,300 cal yr B.P. ocean surface conditions ameliorated as *Turborotalita*
580 *quinqueloba* rapidly became the dominating species and Atlantic Water inflow intensified,
581 resulting in increased summer $sSST_{Transfer}$ ranging from 3.3 to 6.5°C (average 5.5°C).
582 Moreover the flux of planktic foraminifera increased and influence from sea ice and icebergs
583 diminished as the Arctic Front retreated north-westward. Concurrently the $sSST_{Mg/Ca}$
584 recorded by *N. pachyderma* showed values between 2.5 to 5.5°C with an average of 4.0°C
585 which is an increase in average $sSST_{Mg/Ca}$ of only ~0.5°C compared to the preceding period.
586 The relatively modest increase of $sSST_{Mg/Ca}$ compared to $sSST_{Transfer}$ was probably caused by
587 a deepening of the habitat depth and/or a shift in the main calcification season for *N.*
588 *pachyderma* during this period.

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874 **Figure captions**

875 **Table 1.** Radiocarbon dates and calibrations from core MSM5/5-712-2.

876 **Table 2.** Performance of transfer function model WA-PLS with components 1 to 5. Selected
877 model indicated in bold letters.

878 **Figure 1.** (A) Map of the north-eastern North Atlantic Ocean and adjoining seas showing the
879 major currents systems and average position of the Polar and Arctic fronts modified
880 from Marnela et al. (2008). Location of Kastenlot core MSM05/5-712-2 indicated by
881 a double circle and other cores (Hald and Aspeli, 1997 (T-88-2); Ebbesen et al., 2007
882 (MD99-2304)) by open circles. Abbreviations: NwASC: Norwegian Atlantic Slope
883 Current; NwAC: Norwegian Atlantic Current; WSC: West Spitsbergen Current;
884 NCaC: North Cape Current; RAW: Re-circulating Atlantic Water; SB: Svalbard
885 Branch; YSC: Yermak Slope Current; ESC: East Spitsbergen Current; EGC: East
886 Greenland Current. (B) Temperature and salinity profile from the core site measured
887 in August 2007.

888 **Figure 2.** Age model and sedimentation rate of core MSM05/5-712-2. Error bars show the 2σ
889 standard deviation of the calibrated ages.

890 **Figure 3.** Planktic foraminifera, sedimentological and geochemical data plotted against age
891 and depth in core MSM05/5-712-2. (A) Bulk sediment calcium carbonate (CaCO_3)
892 (wt.%) (Aagaard-Sørensen et al., 2013 and present study); (B) Total flux of planktic
893 foraminifera on logarithmic scale (grey shading); (C-G) Relative percentage of five
894 most abundant planktic foraminiferal species (black lines); (H-L) Flux of individual
895 species (grey shading); (M) Grain size $>1\text{mm}$ (IRD); (N) Selected grain sizes (weight
896 %) $>63\mu\text{m}$ and $<63\mu\text{m}$ and lithological units; (O) Diatom-rich layer (vertical light

897 grey shading). Chronostratigraphical zones follow Rasmussen et al. (2007a),
 898 Steffensen et al. (2008) and Walker et al. (2009). Diamonds on X-axis indicate
 899 radiocarbon dated levels.

900 **Figure 4.** Planktic foraminiferal stable isotopes, trace elements and fragmentation plotted
 901 against age and depth in core MSM05/5-712-2. (A) % fragmentation; (B) $\delta^{13}\text{C}$ in *N.*
 902 *pachyderma*. Error bar shows analytical precision ($\pm 0.03\text{‰}$); (C) Ice-volume
 903 corrected and uncorrected $\delta^{18}\text{O}$ (black line: Fairbanks, 1989; grey line: Peltier and
 904 Fairbanks, 2006; thin grey line: uncorrected data) in *N. pachyderma*. Error bar shows
 905 analytical precision ($\pm 0.06\text{‰}$); (D) Mg/Ca in *N. pachyderma* (Aagaard-Sørensen et
 906 al., 2013 and present study). The thin line shows the raw data, and the thick line is the
 907 five-point running mean. The crosses mark omitted data points. The filled circle
 908 shows the average reproducibility of sample splits (± 0.049 mmol/mol); (E) Post
 909 cleaning mass of CaCO_3 (μg). Chronostratigraphical zones follow Rasmussen et al.
 910 (2007a), Steffensen et al. (2008) and Walker et al. (2009). Diamonds on X-axis
 911 indicate radiocarbon dated levels.

912 **Figure 5.** Foraminiferal records and sSST reconstruction from core MSM05/5-712-2
 913 compared to other records from the Nordic Seas and the NorthGRIP ice core plotted
 914 against age. (A) Stable oxygen isotope ($\delta^{18}\text{O}$) record obtained from the NGRIP ice
 915 core (Rasmussen et al., 2006). (B) Reconstructed $\text{sSST}_{\text{Mg/Ca}}$ (black line) and
 916 $\text{sSST}_{\text{Mg/Ca}+15\%}$ based on Mg/Ca values that have been artificially increased 15 % (thin
 917 grey line) using the temperature equation of Kozdon et al. (2009). The $\text{sSST}_{\text{Transfer}}$ for
 918 100 m water depth (Husum and Hald, 2012) (thick grey line). Triangles on Y-axis
 919 indicate modern water temperature at 25 (6.7°C) and 250 (3.7°C) m water depth in the
 920 Fram Strait (See Fig. 1B). (C) $\text{sSST}_{\text{Transfer}}$ reconstruction at 100 m water depth in core

921 MD99-2304 (thick grey line) (Ebbesen et al., 2007; Husum and Hald, 2012). (D)
 922 sSST_{Transfer} reconstruction at 100 m water depth in core T-88-2 (thick grey line) (Hald
 923 and Aspeli, 1997; Husum and Hald, 2012). (E-G) Relative percentage of *T.*
 924 *quinqueloba* (grey shading) in core MSM05/5-712-2 (present study), MD99-2304
 925 (Ebbesen et al., 2007) and T-88-2 (Hald and Aspeli, 1997). Chronostratigraphical
 926 zones follow Rasmussen et al. (2007a), Steffensen et al. (2008) and Walker et al.
 927 (2009).

928 **Table 1**

Lab. code	Depth range (cm)	Material	¹⁴ C age	Calibrated age ± 2σ	2 σ max cal. age (cal. age intercepts) 2 σ min cal. age	Reservoir age (R=400 + ΔR)	δ ¹³ C(‰)
Poz-30723	214-215	<i>N. pachyderma</i>	8362±45	cal. BP 8749 ±209	cal. BP 8540 (8749) 8958	551±51	-0.6 ± 0.4
KIA 37423	280-281	<i>N. pachyderma</i>	9220±50	cal. BP 9797±252	cal. BP 9551 (9797) 10042	551±51	-2.99 ± 0.35
Poz-30725	322-323	<i>N. pachyderma</i>	9580±47	cal. BP 10310 ±158	cal. BP 10152 (10310) 10468	551±51	-2.6 ± 0.1
Poz-30726	428-431	<i>N. pachyderma</i>	12358±63	cal. BP 13629±197	cal. BP 13432 (13629) 13826	551±51	-1.1 ± 0.1

929

930

931 **Table 2**

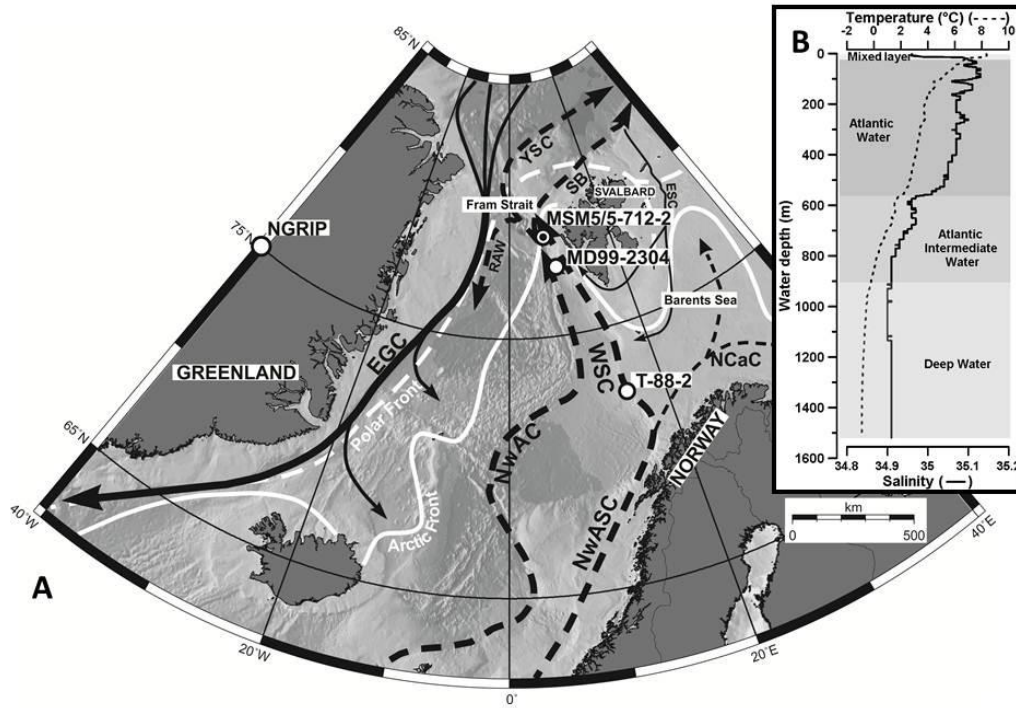
Model	RMSEP	r ²	Max Bias
WAPLS Component 1	0.56257	0.89493	0.63227
WAPLS Component 2	0.52205	0.90954	0.58702
WAPLS Component 3	0.51659	0.91141	0.56346
WAPLS Component 4	0.51718	0.91121	0.59419
WAPLS Component 5	0.52354	0.90902	0.59279

932

933

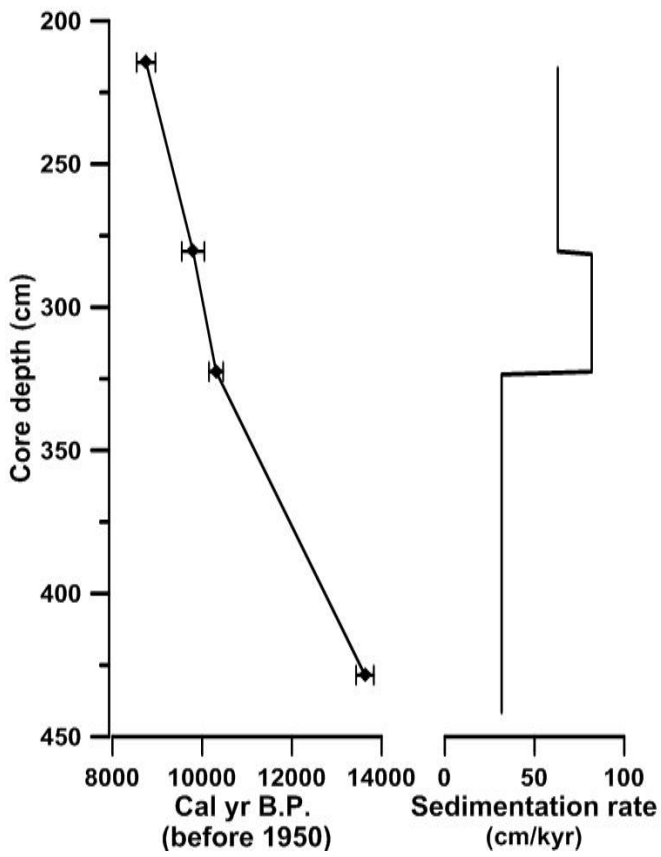
934

935 Fig 1



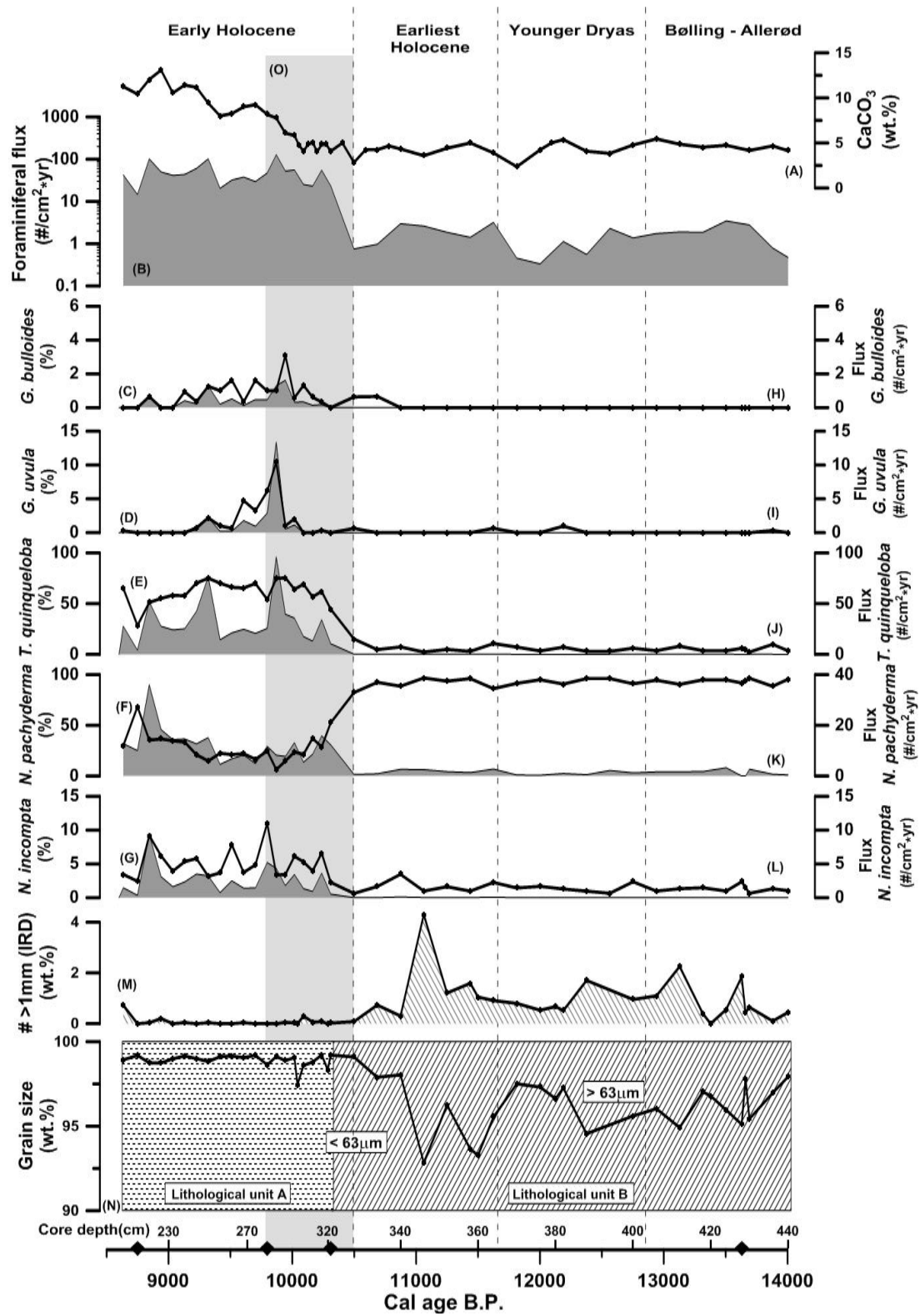
936

937 Fig 2



938

939 Fig 3

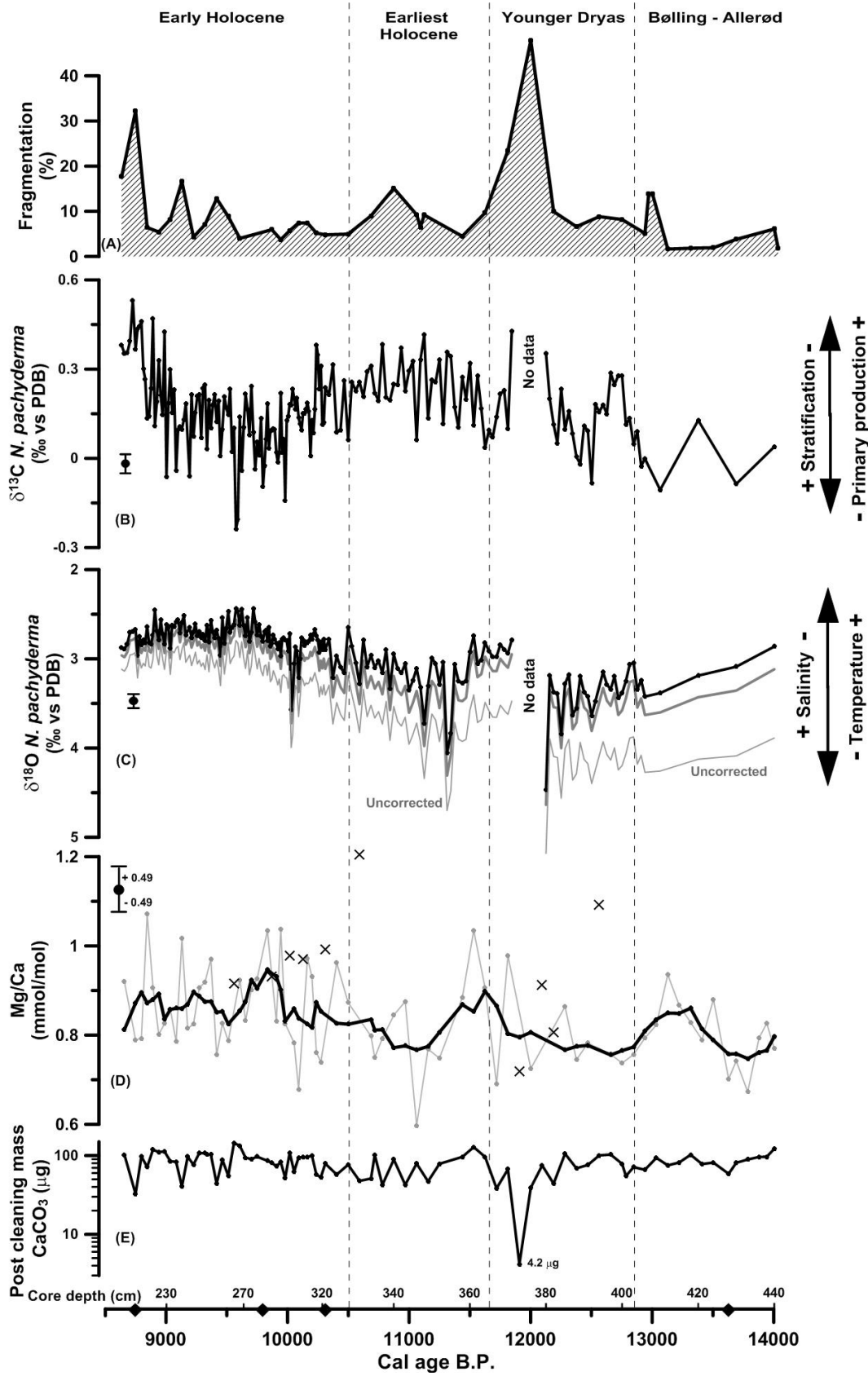


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942

943 Fig 4



945 Fig 5

