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► To cite this version:

J.T. T. Andrews, L. Smik, S.T. Belt, Marie-Alexandrine Sicre, I.N. Mccave. Ocean surface and bottom water conditions, iceberg drift and sediment transport on the North Iceland margin during MIS 3 and MIS 2. *Quaternary Science Reviews*, 2021, 252, pp.106722. 10.1016/j.quascirev.2020.106722 . hal-03414780

HAL Id: hal-03414780

<https://hal.science/hal-03414780v1>

Submitted on 4 Nov 2021

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1 Accepted, *Quaternary Science Reviews*, 13/11/2020. MS No.: JQSR-D-20-00139R3

2

3 **Ocean surface and bottom water conditions, iceberg drift and sediment transport on**

4 **the North Iceland margin during MIS 3 and MIS 2**

5

6 Andrews, J.T. ¹, Smik, L. ², Belt, S.T. ², Sicre, M-A ³, McCave, I. N. ⁴

7

8 1. Institute of Arctic and Alpine Research, and Department of Geological Sciences,

9 University of Colorado, Boulder, CO 80303

10 2. Biogeochemistry Research Centre, School of Geography, Earth and

11 Environmental Sciences, University of Plymouth, Drake Circus, Plymouth, PL4

12 8AA, UK.

13 3. LOCEAN, CNRS, Sorbonne Université, Campus Pierre et Marie Curie, 4 place

14 Jussieu, Paris, France.

15 4. Godwin Laboratory for Palaeoclimate Research, Department of Earth Sciences,

16 University of Cambridge, Downing Street, Cambridge CB2 3EQ, U.K.

17

18 **Orcid:** Andrews: 0000-0003-3169-5979

19 **Orcid:** Belt: 0000-0002-1570-2924

20 **Orcid:** Sicre: 0000-0002-5015-1400

21 **Orcid;** McCave: 0000-0002-4702-5489

22

23

24 corresponding: andrewsj@colorado.edu

25

26 **Key words:** Iceland Plateau, MIS 2 and 3, sea ice biomarkers, IP₂₅, alkenones,

27 sortable silt, sediment provenance

28 **Abstract**

29 Radiocarbon dates and marine tephra suggest that the upper 10 m of core MD99-2274,
30 off North Iceland, extends from ~0 to ~65 ka BP. A multi-proxy sediment and biomarker
31 study at a millennial-scale resolution is used to derive a paleoclimate scenario for this
32 area of the southwestern Nordic Seas, which during the Holocene had intermittent
33 excursions of icebergs and a seasonal cover of drifting sea ice across the site. The
34 sortable silt mean size (\overline{SS}) suggests a bottom current (1000 m depth) flow speed
35 maximum to minimum range of ~8 cm/s during Marine Isotope Stages 2 to 3, but the data
36 are unreliable for the Holocene. Slow-down in flow speeds may be associated with
37 massive ice and water discharges linked to the Hudson Strait ice stream (H-events) and to
38 melt of icebergs from Greenland in the Nordic seas where convection would have been
39 suppressed. Five pulses of sediment with a distinct felsic component are associated with
40 iceberg transport from E/NE Greenland. Sea ice, open water and sea surface temperature
41 (SST) biomarker proxies (i.e. IP₂₅, HBI III, brassicasterol and alkenones) all point
42 towards near-perennial sea ice cover during MIS 3 and 2, rather than seasonal sea ice or
43 open water conditions. Indeed, our biomarker and sediment data require that the seas
44 north of Iceland experienced a nearly continuous cover of sea ice, together with icebergs
45 calved from ice stream termini, which drifted southward. The cross-correlation of the
46 quartz % records between MD99-2274 and the well-dated core PS2644 in Blosseville
47 Basin indicates significant coherence in the records at a multi-millennial (~8 ky)
48 timescale. A transition to open ocean conditions is evident from the early Holocene
49 onwards, albeit with the occurrence of some drift ice and icebergs.

50

51 **1. Introduction**

52

53 *1.1 Aims of study*

54 In order to gain some understanding of the complex marine environments that prevailed
55 during the Late Quaternary we need to employ a multi-proxy approach that not only
56 characterizes ocean surface and bottom water conditions, but also provides direct
57 measurement of glacial influences on sediment supply. Several studies have been
58 reported from the North Iceland Shelf (Fig. 1) that document late glacial/Holocene
59 records (e.g. Andrews et al., 2018; McCave and Andrews, 2019a & b; Sicre et al., 2008;
60 Knudsen et al., 2003) but there are only limited references to conditions during Marine
61 Isotope Stages (MIS) 2, 3 or 4. Therefore, with the primary goal of establishing a
62 framework for environmental conditions in this sector of the Iceland Sea from MIS 2 to
63 MIS 4, we selected a previously unstudied core, MD99-2274 (Labeyrie et al., 2003) (Fig.
64 1), and sampled the upper 10 m. MD99-2274 (henceforth #2274) is a 10-cm diameter 26
65 m Calypso core retrieved from 67.582°N and 17.073°W at 1000 m water depth (Labeyrie
66 et al., 2003) during the IMAGES V cruise aboard the French RV *Marion Dufresne*. For
67 further context, we note that the core site is located 200 km east of the well-studied core
68 PS2644 (van Kreveld et al., 2000; Voelker, 1999; Voelker and Haflidason, 2015) and 163
69 km from core P57-7 (Sejrup et al., 1989) (Table 1, Fig. 1A). The main questions we
70 posed were: 1) what is an appropriate depth/age model, 2) is there evidence for either
71 pervasive sea ice or an ice shelf (Boers et al., 2018; Dokken et al., 2013; Petersen et al.,
72 2013), which have been called for to explain D-O cycles, 3) what were sea surface
73 temperatures (SSTs), and 4) are there substantial changes in grain-size and mineral
74 composition that can be associated with changes in bottom current flow speed and

75 changes in glacial sediment provenance? Given the location of the core (Fig. 1A) we
76 were particularly interested in whether we could discriminate between glacial sediments
77 derived from Iceland versus those from E/NE Greenland.

78 *1.2 Present-day oceanography*

79 The Iceland and Greenland Seas (Fig. 1A) are key areas for the formation of dense
80 overflow waters (Brakstad et al., 2019) that flow south through sills in the Scotland-
81 Greenland Ridge (Fig. 1C). The North Icelandic Jet (NIJ) flows southwestward along the
82 slope below ~1000 m (Fig. 1C) with a mean speed of $9.3 \text{ m} \pm 2.7 \text{ m/sec}$ towards Denmark
83 Strait (Mauritzen, 1996; Pickart et al., 2005) where it exits to form a major component of
84 North Atlantic Deep Water “....and points to the Iceland Sea as an important place for
85 *this water mass formation.*” (Jonsson and Valdimarsson, 2004). The study site lies in a
86 sensitive area with the surface flow being the East Icelandic Current (EIC), which brings
87 cold and relatively fresh surface water as a spin-off from the East Greenland Current
88 (EGC), whereas the North Icelandic Irminger Current (NIIC), sourced from the southern
89 warmer and saltier waters of the North Atlantic Drift (Stefansson, 1962), continues as an
90 eastward flow over the inner North Iceland Shelf (NIC) (Fig. 1C).

91 Sea ice in the form of drift ice has been noted to reach the area in modern times,
92 although the average position of the sea ice edge (30% sea ice cover by area) lies north of
93 our site (Divine and Dick, 2006) (Fig. 1C). Thirty years of observations on the presence
94 of icebergs (Andrews et al., 2019), their Fig. 7A) indicate that icebergs from E/NE
95 Greenland drift across the site.

96

97 *1.3 Background to study region*

98 Stein and colleagues (Nam et al., 1995; Stein, 2008; Stein et al., 1996) studied a
99 comprehensive suite of cores on the Scoresby Sund Trough Mouth Fan (Fig. 1 A, TMF)
100 and reported both ice-rafted debris (IRD) and $\delta^{18}\text{O}$ on the near-surface planktonic
101 foraminifera *Neoglobquadrina pachyderma* (Table 1). The cores included discrete IRD
102 peaks (counts $10\text{ cm}^3 > 500\ \mu\text{m}$), which they suggested may have been coeval with the
103 massive ice and water discharges of the Hudson Strait Heinrich (HS H-) events (Andrews
104 and Voelker, 2018b; Heinrich, 1988; Hemming, 2004; Hesse 2016). However, whether
105 the response of the Greenland, Iceland, and European ice sheets was synchronous or
106 asynchronous with the Laurentide Ice Sheet collapse events still requires clarification
107 (Dowdeswell et al., 1999; Elliot et al., 2001). Verplanck et al (2009) provided radiogenic
108 isotope data fingerprinting sediment sources from two cores on the Scoresby Sund TMF
109 (O'Cofaigh et al., 2002) (JR51-GC31 and -GC32) and another core (PS62/017-4) from
110 the Blossville Basin (Milo et al., 2005) (Table 1). Stein et al. (1996) and Verplanck et al.
111 (2009) described events in cores PS1730 and PS62/017-4 (Table 1, Fig. 1) that they
112 considered coeval with the HS H-events. Andrews and Voelker (2018) have argued that
113 the use of the term “Heinrich events” for locations such as the Nordic Seas is not
114 appropriate and should be modified. For example, the IRD-rich layer in PS2644
115 correlated with HS H-2 (Voelker et al., 1998) is now referred to as PS2644 IRD#2
116 (Andrews and Voelker, 2018). In our study, events that might correlate with HS H-events
117 will be termed #2274-IRD#.

118 There is no firm agreement on the extent and duration of sea ice cover in the
119 Nordic Seas during MIS 2 and MIS 3. The CLIMAP data showed an extensive cover
120 across the Nordic Seas (Ruddiman and McIntyre, 1981) whereas Sarnthein et al. (2003)

121 argue that the Nordic Seas during MIS 2 were “..largely ice free” during the summer
122 months. The presence of an ice shelf buttressing the East Greenland ice streams has also
123 triggered a debate especially as to a possible answer to the cause of D-O oscillations
124 (Pettersen et al., 2013; van Kreveld et al., 2000). However, other researchers working at
125 sites in the eastern Nordic Seas have rather focused on the role of sea ice (Dokken et al.,
126 2013; Hoff et al., 2016) and changes in the structure of the water column, and concluded
127 that during Greenland interstadials in MIS 3, sea ice was limited in extent and duration.

128 The presence of thick, pervasive sea ice could potentially limit the export of
129 icebergs from E and NE Greenland Ice Streams (Reeh et al., 1999), although the
130 sediment records from numerous sediment cores retrieved from the floor of the Arctic
131 Ocean clearly document that iceberg rafting occurred throughout the Pleistocene (Clark,
132 1990a,b; Stein, 2008; Phillips and Grantz, 2001; Stokes et al., 2005), with some evidence
133 that the timing of events in some cores were similar to those for HS H-events. For
134 example, IRD peaks in cores from the Arctic Ocean were linked to the McClure Ice
135 Stream in the NW sector of the Laurentide Ice Sheet and dated at 12.9, 15.6, ~22, and 30
136 ka BP (Stokes et al., 2005). Iceberg drift is primarily a function of the integrated current
137 direction and speed over depth, plus a component associated with wind forcing on the
138 exposed “sail” (Bigg, 2016). In many ways, sea ice protects icebergs as it inhibits wave
139 action, which is the greatest cause of iceberg disintegration (Bigg, 2016; Venkatesh et al.,
140 1994).

141 *1.4 Ice sheet extent MIS 1 to MIS 3*

142 #2274 lies only 60 km north of the LGM limit of the Iceland Ice Sheet (IIS) (Fig. 1)
143 (Andrews and Helgadóttir, 2003; Patton et al., 2017) with the onset of retreat associated

144 with calibrated radiocarbon dates of between 14 and 15 ka BP, depending on the ocean
145 reservoir correction (Andrews et al., 2018; Andrews and Helgadottir, 2003; Knudsen et
146 al., 2003). Retreat from the maximum position was rapid (Andrews et al., 2018; Norðdahl
147 and Ingolfsson, 2015; Patton et al., 2017), and the ice sheet was at or behind the present-
148 day coast by the time of the deposition of the Vedde tephra ~12.2 ka BP (Lohne et al.,
149 2013). Little detail is known about the history of this ice sheet during MIS 3 (e.g.
150 Andrews et al., 2017). Moles et al. (2019) argued that the North Atlantic Ash Zone II
151 (NAAZII) tephra, dated ca 54 ka BP (Austin and Hibbert, 2012), was erupted under >400
152 m of ice, thus indicating a reasonably extensive IIS during the Greenland ¹⁸O stadial 15.2
153 (Moles et al., 2019; Rasmussen et al., 2014), but no specific information is currently
154 available on the MIS 3 history of the ice sheet.

155 The Greenland Ice Sheet (GIS) extended to the shelf break during the LGM
156 (Funder et al., 2011b; Vasskog et al., 2015) but little is known about its history during
157 MIS 3 or MIS 4. Judging from the delivery of quartz-rich sediments to cores along
158 Denmark Strait, especially PS2644 and MD99-2323 (Andrews and Vogt, 2020a), it is
159 probable that the ice also reached a similar position at these times. Peterson et al. (2013)
160 suggested that an ice shelf may have extended out from the East Greenland Shelf across
161 Blosseville Basin, although the sedimentary evidence for this is scanty (Andrews and
162 Vogt, 2020a).

163

164 *1.5 Bedrock Geology and source signatures*

165 In terms of the mineral composition of #2274 sediments, the bedrock in glacial source
166 areas consists primarily of either mafic (basalt) or felsic (granites/gneisses/sandstones),

167 although finer source identification is possible (Andrews and Vogt, 2014; 2020) (Fig.
168 1A). Further, Andrews and Vogt (2014) demonstrated that the sediment mineral
169 signature of sediments offshore from the Caledonian Fold Belt was dominated by high
170 wt% of quartz, illite, and muscovite. Detrital carbonate sediments derived from the
171 Paleozoic outcrops of N Greenland and the Canadian Arctic are also recognized by color
172 and mineralogy. However, radiogenic isotopes (White et al., 2016; Verplanck et al.,
173 2009) allow more age-related differentiations, which in terms of our region (Fig. 1A and
174 B), consists of Archaean, Paleoproterozoic, Caledonian Fold Belt, and Tertiary volcanics
175 (Henriksen, 2008).

176 **2. Environmental proxies and age model**

177 *2.1 Data methods*

178
179 The proxies used in this paper are the sea ice biomarkers IP₂₅ and HBI II (Belt et al.,
180 2007; Belt and Müller, 2013; Belt, 2018), brassicasterol and HBI III as indicators of open
181 water primary production (Volkman, 1986; Belt et al., 2015), alkenones (for SST) (Sicre
182 et al., 2008a), % C37:4 alkenone to identify polar surface waters, grain-size indicators of
183 bottom flow and deposition (McCave and Andrews, 2019a, b; McCave et al., 2017),
184 magnetic susceptibility, and quantitative X-ray diffraction estimates of mineral wt%
185 (Andrews et al., 2017; Andrews and Vogt, 2014). The X-ray diffraction data for #2274
186 are available (Andrews and Vogt, 2020b) The full details of these methods are included
187 as Supplementary Material.

188

189 *2.2 Depth/age model*

190 The age model is based on radiocarbon dates and the occurrence of tephras (Table 2).
191 There are significant problems associated with obtaining and interpreting calibrated ages
192 because of the uncertainty of the ocean reservoir correction (ORC), which has varied
193 spatially and temporally, and might be as much as 1000 yr (Andrews et al., 2018; Skinner
194 et al., 2019; Voelker, 1999; Voelker et al., 1998). Three radiocarbon dates were obtained
195 on the near-surface planktonic foraminifera *Neogloboquadrina pachyderma* and the other
196 on lustrous shell fragments. Most tephras older than the Borrobol (ca 14.5 ka BP) (Lind
197 et al., 2016; Matthews et al., 2011) are dated by reference to GIS cores, which themselves
198 are based on a variety of assumptions and whose error increases with the estimated age
199 (Boers et al., 2017). The qXRD data (Andrews et al., 2013; 2018) suggest the presence of
200 high wt% of volcanic glass in two cores on the Iceland Shelf that might be coeval with
201 the Vedde and NAAZII tephras (Brendryen et al., 2011; Lohne et al., 2013). The tephra
202 bed at 607 cm in #2274 was identified by Haflidasson (person. commun. 2018) as being
203 similar to FMAZ IV dated at ~47.12 ka BP (Davies et al., 2008; Rasmussen et al., 2003;
204 Voelker and Haflidason, 2015) and that date is used in our depth/age models (see
205 Supplementary Material). Other discrete layers of black basaltic glass were noted in the
206 shipboard log at 99, 127.5, 717, and 740 cm (Labeyrie et al., 2003, p 477), and age
207 estimates were obtained from our depth/age model (see later).

208 We used the Bayesian radiocarbon calibration program “Bacon” (Blaauw and
209 Christen, 2005) to construct depth/age models, but we also acknowledge the many
210 problems associated with establishing accurate depth/age models (Telford et al., 2003;
211 Trachsel and Telford, 2017). The first model is based solely on the available ^{14}C dates
212 and the 607 cm tephra (Table 2A and B), while the second one is based on an assumed

213 age estimate for the core top of 500 ± 500 (i.e. little sediment loss) and the inferred
214 presence of the Vedde and NAAZII tephras. Given the uncertainty in the OCR, we used
215 a $\Delta R = 0$. In practice, there is relatively little difference in the median age estimates (Fig.
216 2A). The average sediment accumulation rate (SAR) is 68 yr/cm or 14.7 cm/ky, thus our
217 10 cm sampling density permits millennial-scale evaluations, with an average spacing
218 between samples of 0.5 cal ky. Remarkably, for MD cores of this vintage (1999), the
219 upper part of the core shows no evidence of piston-induced stretching (Skinner and
220 McCave, 2003). However, the spread between minimum and maximum age estimates is
221 often considerable given the relative paucity of dated levels, and the Bayesian approach
222 results in an age estimate for the core top of 3600 yr BP, although the estimated date of
223 500 ± 500 yr BP finds some support in our data. The estimated ages for the logged tephra
224 layers are: ~11, 13.2, 53, and 56 ka BP. A possible age for the 99 cm basaltic tephra is the
225 10.2 ka BP Saksunarvatn tephra (Lohne et al., 2013), which is widespread on the north
226 Iceland Shelf (Krisjansdottir et al., 2007; Eiriksson and Knudson, 2002). All our
227 subsequent data have been converted to a common depth/age model using the data in
228 Table 2B; thus, robust inter-proxy comparisons can be made. To ensure that we have not
229 forced our data into an existing framework we have not tuned our model to other records
230 (Blaauw, 2012).

231 We have also obtained radiocarbon dates on several *Vema* cores that were taken
232 to the north of Iceland and #2274 (Fig. 1; Table 3) (Manley and Jennings, 1996). The
233 calibrated radiocarbon dates range from ~13 to > 49 ka BP ($\Delta R = 0$) and were obtained
234 on relatively large samples of *N. pachyderma* (Table 3). Several tephras were noted in the
235 core description (Suppl. Data), thus indicating that conditions allowed for the deposition

236 of discrete tephras. The dates from these cores also provide additional information on the
237 presence of significant numbers of the planktonic foraminifera *N. pachyderma* (Greco et
238 al., 2019) and hence inferences about sea ice cover and light conditions.

239 3. Results

240

241 3.1 Lithology and Grain-size

242 The core log of core #2274 (Labeyrie et al., 2003. p. 477) described the sediment as
243 being principally mottled silty clay with colors ranging between 2.5Y4/2 to 5Y4/1.

244 Visible ice-rafted clasts occur but are not common. The grain-size measurements were
245 undertaken on sample splits from the qXRD samples and only 30 samples were
246 processed, resulting in a coarse resolution data set (on average one sample every 2300
247 yr). The sediments vary between a very coarse to a fine silt with average grain-sizes
248 varying between 54.3 to 6.05 μm . Sand $> 240 \mu\text{m}$ is considered to be ice-rafted (McCave
249 and Andrews, 2019a) and occurs in low % throughout the core, except for two notably
250 coarser intervals with $\text{IRD}_{240} > 5\%$, (Fig. 3).

251 We have also undertaken an analysis of the sortable silt mean size (\overline{SS}) and SS%
252 in the 63-10 μm fractions (McCave et al., 1995). The correlation coefficient between
253 these two variables is $r = 0.804$ indicating, relative to other cores (McCave and Andrews,
254 2019a,b), a somewhat noisy correlation, but a generally current-sorted signal
255 (Supplementary Fig. 1). Computation of the running correlation between SS% and \overline{SS}
256 yields high average values ($r > 0.9$) between ~ 11 and 42 ka BP but values unacceptable
257 for flow speed inference occur in the Holocene and during brief interval ~ 57 ka BP (Fig.
258 3). Variations in the flow speed of bottom currents (Fig. 1C) in this region reflect
259 changes in the vigour of the ocean overturning system because the NIJ feeds into the

260 Denmark Strait overflow, a key starting point for the North Atlantic western Boundary
261 Undercurrent.

262 The overall range (minimum-maximum) in flow speed indicated by this record is
263 ~ 8 cm/s. Calibration of the sortable silt proxy yields a sensitivity ($\text{cm s}^{-1}/\mu\text{m}$) rather than
264 an absolute speed-size relationship (McCave et al., 2017). In favourable circumstances
265 actual speeds may be estimated by matching core-top \overline{SS} data to nearby current meter
266 measurements and plotting the differences downcore. Unfortunately, because the
267 Holocene data are unreliable as a speed record, we cannot relate this to the present nearby
268 flow speed measurements of 9.3 cm/sec (Jonsson, 2004). Nevertheless, low speeds
269 correspond to HS H 1 (~ 15 ka), 4 (~ 40 ka), and 6 (~ 60 ka) (Fig. 3) as expected from
270 previous work on the impact of Heinrich and other cold events on N. Atlantic circulation
271 (e.g. Kleiven et al., 2011), on the basis of which, speeds of < 5 cm/s are probable.

272

273 *3.2 Mineral composition*

274 On Figure 4, we plot the changes in the weight % of key minerals as determined by
275 qXRD as well as the ratio quartz/pyroxene, which we use as a measure of felsic/mafic
276 bedrock (as opposed to quartz/plagioclase which was used by Moros et al. (2004)). The
277 quartz wt% in a surface grab from this site is $\sim 5\%$ (Andrews and Eberl, 2007), and the
278 median for the whole record is 5.3 % with a maximum of 16.8 %. The magnetic
279 susceptibility record for #2274 (Fig. 2A) is clearly inversely associated with the
280 variations in quartz (Fig. 2B), which, together with the K-feldspars, are diamagnetic
281 minerals (Robinson et al., 1995; Watkins and Maher, 2003). A similar inverse
282 relationship was noted in other cores from the area (Andrews and Vogt, 2020a). Hence

283 the magnetic susceptibility fluctuations support our interpretation that there are
284 substantial variations in the inputs of felsic- versus mafic-rich sediments.

285 The Holocene record mimics that from many sites on the North Icelandic Shelf
286 (NIS) in showing an increase in quartz toward the present-day (Andrews et al., 2019).
287 Quartz and pyroxene have an antiphase relationship ($r^2 = 0.47$), which in part is related to
288 the mineral data summing to 100% (i.e. a closed array), and which provides some
289 constraints on the interpretation (Aitchison, 1986; Chayes, 1971). There are five
290 sustained peaks in the quartz wt % (Fig. 4), and K-feldspar (not shown, K-feldspar values
291 track those of quartz (Andrews and Vogt, 2020a)) are therefore not included in this
292 figure), which we interpret as indicating the influx of sediment from NE Greenland and
293 possibly farther afield from Canada or Fennoscandia. Of these possible mechanisms,
294 icebergs alone carry basal and englacial debris that includes all size fractions from
295 cobbles to clay ($> 1 \mu\text{m}$). The variations in quartz are frequently matched by the sum of
296 calcite and dolomite (carbonate) (Fig. 4) ($r^2 = 0.13$, $p < 0.0001$) although the correlations
297 are much more significant for dolomite ($r^2 = 0.22$) than calcite ($r^2 = 0.07$). This probably
298 represents transport of glacially derived material from the carbonate bedrock of NE and
299 N Greenland and/or the Canadian Arctic Islands and Channels (Darby and Zimmerman,
300 2008; Lakeman et al., 2018; Phillips and Grantz, 2001). The estimated ages for the 5
301 peaks are 14.4, 31.5, 40, 54.7, and 61.8 ka BP (Fig. 4) with a possible smaller episode
302 ~ 22.8 ka BP. These age estimates are somewhat similar to the HS H-events (Andrews
303 and Voelker, 2018a; Heinrich, 1988; Hemming, 2004) (see Fig. 3) but their duration are
304 longer than the < 1 ky episodes of detrital carbonate deposition associated with the HS H-
305 events (Andrews and Voelker, 2018a).

306 Previous work on sediment sources in this area (Verplanck et al., 2009) provide
307 temporally limited but critical information using radiogenic isotopes on the < or > 63 μ m
308 fractions. Debris flow from the two Scoresby Sund TMF sites (JR51-GC31,-32, Table 1,
309 Fig. 1B) lay along the 1.7 Ga Paleoproterozoic isochron; the samples contained abundant
310 quartz and lesser amounts of basalt (Verplanck et al., 2009, p.53). However, the
311 sediments in the Blossville Basin (core 17-4, Fig. 1A, Table 1), some 150 km
312 downstream (Fig. 1A), and considered to be coeval with HS H events-1, -2, and -3, all
313 cluster along the 0.5 Ga isochron (Caledonide bedrock, that outcrops on the eastern edge
314 of NE Greenland north of Scoresby Sund (Fig. 1B)). The same isotopic signature
315 characterized the non-HS H sediments in this core. Pb systematics indicate that the
316 Holocene sediment samples at sites 907 (Table 1) and JR51-GC28 are dominated by the
317 0.5 Ga Caledonides (White et al., 2016). Given the sediment SedUnMix results (Fig. 5)
318 and the reported radiogenic isotopic data (Verplanck et. al., 2009; White et al., 2016), the
319 variations in quartz are most probably associated with sediment discharge events from
320 glacial erosion and transport in ice streams flowing through the numerous fiords north of
321 Scoresby Sund and primarily within the Caledonian Fold Belt outcrop (Evans et al., 2002,
322 2009; Stein, 2008).

323 The SedUnMix analysis included sediments from NE Greenland (Calendoides,
324 ~73N; Andrews et al., 2016), E Greenland (basalt), and Iceland. The analysis of possible
325 bedrock sources for the #2274 compositional changes indicated (as might be expected
326 given the bedrock geology of E and NE Greenland, and Iceland) that the NE Greenland
327 source had a granite and gneissic composition, whereas E Greenland and Iceland were
328 linked to basalt and also dolerite (Brooks and Nielsen, 1982; Henriksen, 2008; Higgins et

329 al., 2008; Kristjansson et al., 1979). The results (Fig. 5) indicate that felsic-rich sediments
330 from NE Greenland or farther afield (Arctic Canada, Fennoscandia) (Verplanck et al.,
331 2009) were deposited in a series of events that mimic the influx of quartz to the site (Figs.
332 2B and 4); the correlation between the NE Greenland Calendonides source estimates in
333 #2274 and the quartz wt% explains 79% of the variance. The average “unaccounted” or
334 “unexplained” composition averaged 20 ± 5 % and degree of fit or average absolute bias
335 is 2.3 ± 0.4 wt% indicating that the input mineral source regions provide a good fit to the
336 #2274 mineral compositions. Figure 5 highlights two periods when the mineral
337 composition indicates little deposition of sediment that could be ascribed to a felsic
338 source centered around 20 and 57 ka BP, the latter being a time of substantial deposition
339 of tephra at this site and also a time when glacial ice covered at least some of Iceland
340 (Moles et al., 2019). Source estimates from E. Greenland (sites seaward of the early
341 Tertiary basalt outcrop on the Geikie Plateau) and SW Iceland resulted in nearly identical
342 patterns over the last ~65 ka BP (Fig. 5), but the results from considering Icelandic basal
343 glacial marine diamictos (Dmm) as a source are different. The reasons for these two
344 differing estimates are presently unclear.

345 The provenance time-series thus suggests that we can identify four episodes in the
346 arrival of foreign sourced sediments; 1) from ~65 to 38 ka BP when distinct pulses of NE
347 Greenland sourced sediments arrived; 2) 38 to 17 ka BP when there was an overall
348 decrease in this source with virtually no quartz noted ~20 ka BP; 3) a large pulse of these
349 sediments centered ~ 15 ka BP; and 4) the last 10 ka or so that shows a steady increase in
350 this source. This latter event is also noted in MD99-2269 (Fig. 7) and is matched by
351 changes in the sea ice biomarker IP₂₅ (Cabedo-Sanz et al., 2016).

352 3.3 Biomarkers

353 The sea ice biomarkers IP₂₅ and HBI II were absent or below the limit of detection in the
354 majority of the sediment sections analyzed with only a few exceptions (Fig. 6). Of the
355 two, HBI II was always more abundant, consistent with findings from previous studies
356 from the study region and elsewhere in the Arctic (e.g. Massé et al., 2011; Xiao et al.,
357 2013; Bai et al., 2019). In some cases, only HBI II could be identified and quantified,
358 with IP₂₅ likely also present in such horizons but below the detection limit.

359 Alkenones and brassicasterol were found at very low concentrations in glacial
360 sediments contrasting with higher abundances in Holocene sediments. Further, the open
361 water biomarker HBI III was only detected in Holocene sediments (data not shown).
362 While alkenone-SSTs ranged from 7 to 9°C during the Holocene, they are unexpectedly
363 high in the glacial portion of the record, spanning from 8 to 16°C.

364

365 4. Discussion

366 4.1 Icebergs and IRD during MIS 3 and MIS 2

367 There is no general theory about the association of sea ice and icebergs and there is no
368 observational census of the icebergs being transported in the EGC as there is for the
369 Labrador Shelf off Newfoundland, apart from a 30-yr count of icebergs on the Iceland
370 shelves (Jónsdóttir *in* Andrews et al., 2019). However, Cabedo-Sanz et al. (2016) and
371 Darby et al. (2017) showed that Holocene variations in the wt% of quartz and the sea ice
372 biomarker IP₂₅ co-varied in cores to the west and south of #2274, yet this was not the
373 case at #2274 during MIS 2 and MIS 3 (Figs. 4 and 6). In N Greenland, semi-permanent
374 sea-ice conditions prevail today and did so intermittently during the Holocene (Funder et

375 al., 2011a) and it is reasonable to assume that sea-ice would have been more extensive
376 during MIS 2 and MIS 3 when the GIS may have reached the shelf break. However,
377 cosmogenic dates pertaining to the extent of the Northeast Greenland Ice Stream at
378 $\sim 78^{\circ}\text{N}$ (Larsen et al., 2018) have been used to argue that this ice stream was behind its
379 present margin "...41-26 ka."

380 Several authors have argued for the presence of an ice shelf fringing the E/NE
381 GIS (Boers et al., 2018; Petersen et al., 2013). However, sediments recovered from
382 beneath ice shelves are invariably fine-grained and lack ice-rafted debris (Domack et al.,
383 1999; Jennings et al., 2019; McKay et al., 2016), whereas the sediments from the
384 Scoresby Sund TMF (Fig. 1) and margin contain clear IRD (Stein et al., 1996) (Fig. 5;
385 see also Table 3). Radiocarbon dates in Stein et al., (1996a) were based on 2000 *N.*
386 *pachyderma* specimens per sample, and the numerous MIS 2 and MIS 3 radiocarbon
387 dates on *N. pachyderma* from PS2644 (Sarnthein et al., 2003; Voelker, 1999; Voelker et
388 al., 1998, 2000) were obtained on 10 mg samples of 800-2300 tests in 1-cm sediment
389 samples. Although the complete ecology of *N. pachyderma* is not well known, a study of
390 plankton hauls (Greco et al., 2019) indicates a relationship between sea ice cover and
391 chlorophyll, hence suggesting that "*light or light-dependent processes might influence*
392 *the ecology of this species.*" In addition, several of these cores have discrete tephra layers
393 indicating rapid accumulation of tephra by particles falling through the water column,
394 versus a more dispersed occurrence if the tephra was deposited on multi-year sea ice.
395 Together these data indicate that the sea ice, at times during MIS2 and 3 and probably
396 seasonally, must have had extensive leads and open-water areas.

397 Stein et al (1996) present detailed IRD data (counts $10 \text{ cm}^3 > 500 \mu\text{m}$) from a
398 series of radiocarbon dated cores on the Scoresby Sund TMF (Fig. 1; PS1726 and
399 PS1730, Fig. 1B) that reflect delivery of coarse sediments in a discrete series of episodes
400 (data from www.Pangaea.de). Stein (2008) noted coarse sediment intervals that were
401 attributed to iceberg-rafting at ~ 4 -15, 16, 17-18, 20-21, and 22-23 ka BP. There are no
402 mineral composition data for PS1730, but data exist for PS2644, which is 300 km away
403 (Table 1, Fig. 1B) (Andrews and Vogt, 2020a; Vogt, 2017). A comparison between
404 PS2644 and #2274 (Fig. 8A) indicates that PS2644, closer to the Scoresby Sund Ice
405 Stream, has more quartz wt% but there are some notable corresponding peaks in both
406 series. However, we note that the quartz wt% were obtained via two different but
407 comparable quantitative methods (Andrews and Vogt, 2020a; Vogt, 2017; Zou, 2016). To
408 evaluate similarities and differences between these two sites we used cross-wavelet
409 analysis (Roesch and Schmidbauer, 2018; Hammer et al., 2001) (Fig. 8). The wavelet
410 analysis of the two quartz records (Fig. 8A) demonstrates both important coeval events as
411 well as obvious differences. In addition, the overall match between these sites for the
412 earlier part of the record adds confidence to our age model, and also emphasizes the
413 important differences between 35 and 65 ka. The reconstructed wavelets for PS2644
414 show three major pulses of quartz at ~ 13 , 20, and 29 ka BP, and these are matched by
415 much lower peaks at #2274. Conversely, there are no distinct peaks during MIS 3 in
416 PS2644 but there are in #2274. The sense of the directional arrows in Figure 8B is that
417 PS2644 either leads or is in phase with #2274, and there is a hint of a significant shorter
418 period ~ 60 ka BP with the two records being anti-phase. The cross-wavelet power
419 spectrum (Fig. 8B) confirms the presence of a significant zone of coherence extending

420 from ~10-34 ka BP with the average cross-wavelet power peaking at ~8 ky (Fig. 8C); this
421 is of course similar to the periodicity of HS H-events (see Clark et al., 2007) (e.g. Fig. 3)
422 but lacks the diagnostic carbonate provenance indicators (Andrews and Voelker, 2018).
423 Possibly because of our 0.6 ky sample spacing (Fig. 8A), there is no obvious D-O signal
424 in the quartz PS2644 data, whereas it is evident in the $\delta^{18}\text{O}_{Np}$ data (Suppl. Fig. 3). The
425 difference in signals between #2274 and PS2644 during MIS 3 (Fig. 8A) suggests a
426 change in either the delivery of quartz-rich sediments or a dampening down of sediment
427 delivery.

428 The sortable silt evidence indicates that even at the glacial maximum there was
429 flow along the slope in the precursor to the NIJ. As this presently heads toward the
430 Denmark Strait outflow, we suggest that the Nordic Seas acted as a source of deep waters
431 (probably formed in the east where Atlantic inflow continued (Sarnthein et al, 1994)) that
432 overflowed to the North Atlantic where they formed a deep water mass (Howe et al.,
433 2016; Keigwin and Swift, 2017). The classical view of Nordic Sea behaviour during cold
434 periods is that freshwater from melting ice-sheets and -bergs suppresses convection
435 resulting in a severe reduction or even cessation of the AMOC inflow and overflow (e.g.
436 a recent model, including consideration of the EGC, analysing this is from Liu et al,
437 (2018)). However an emerging view is of a slowdown (not cessation) of Nordic Sea
438 overflows in cold periods (Howe et al., 2016; Keigwin and Swift, 2017). A very recent
439 view is that ice discharges in the North Pacific precede Heinrich events and may be
440 implicated as a triggering mechanism (Walczuk et al., 2020). In the Nordic Seas Atlantic
441 water inflow persisted throughout the Pleistocene glacials over the Norwegian slope
442 (Sarnthein et al., 1994; Newton et al., 2018). The evidence here indicates a persistent

443 outflow along the N Iceland Slope with reductions during HS H- events 1, 4, and 6. Flow
444 speed decreases have been noted for both shallow and deep flows in this region during
445 stadials and glacial intervals of the late and mid-Quaternary (Kleiven et al., 2011;
446 McCave and Andrews, 2019b). The Younger Dryas often shows speed decreases but
447 some cores record increased flow (McCave and Andrews, 2019b), as is seen here. These
448 disparities remain a puzzle.

449

450 *4.2 Rationalizing mineralogical and biomarker proxies for sea ice reconstruction*

451 When detected, the concentrations of IP₂₅ and HBI II were mainly much lower than those
452 reported previously for mid-late Holocene (ca. 6-0 cal. ka) and deglacial (ca. 15-11 ka)
453 sites from the NIS (Cabedo-Sanz et al., 2016; Xiao et al., 2017). However, the presence
454 and concentration of IP₂₅ at ca. 3.7 ka aligns with previous data reported from core JR51-
455 GC35 (located 76 km SW of #2274 (Figs. 1B and 7; Table 1)) for the mid-Holocene
456 (Cabedo-Sanz et al., 2016), consistent with the delivery of drift ice across the NIS at that
457 time (Fig. 7). The otherwise general absence of IP₂₅ and HBI II in #2274 points towards
458 an environment unfavorable for sea ice diatom growth, namely ice-free conditions or a
459 setting of near-permanent ice cover. To distinguish between these two scenarios, we
460 measured three other biomarker types indicative of open water conditions, i.e.
461 brassicasterol, HBI III and alkenones. In the case of brassicasterol, a phytosterol
462 characteristic of marine diatoms (Volkman, 1986), concentrations in selected sediments
463 from #2274 were relatively high in the Holocene section and typically two orders of
464 magnitude lower in older (>14 ka) intervals, indicative of much lower glacial primary
465 productivity reflecting near-perennial sea ice cover. Similarly, HBI III, a biomarker

466 derived from certain open water diatoms (Belt et al., 2017), was only detected in
467 Holocene sections (data not shown). Consistent with these findings, concentrations of
468 alkenones derived from coccolithophorid blooming in mid-late summers were also
469 substantially lower in the older sections compared to those in the Holocene (Fig. 6).
470 Further, the relatively high percentage contribution of the tetra-unsaturated alkenone $C_{37:4}$
471 prior to the Holocene (mean value 36% compared to 6% for the Holocene) is consistent
472 with a dominance of polar waters (Sicre et al., 2002; Bendle et al., 2005) potentially
473 laden with sea ice. Alkenone-derived SST estimates for the Holocene (ca. 7–9°C) are in
474 line with those reported from other high-resolution studies from the NIS (e.g. Bendle and
475 Rosell-Melé, 2007; Sicre et al., 2008b; Kristjansdottir et al. 2016). In contrast, SST
476 estimates prior to the Holocene were somewhat higher (ca. 8–16°C; mean 11.4°C)
477 although the accuracy of such estimates might be lower than for the Holocene owing to
478 the relatively high contributions from $C_{37:4}$ (Bendle and Rosell-Melé, 2004).
479 Anomalously warm SSTs associated with low alkenone concentrations during glacial
480 time have been reported in previous studies and attributed to advection of detrital
481 alkenones (Sicre et al., 2005; Knutz et al., 2011). Such advection by surface currents can
482 introduce significant bias in regions where there are large productivity and SST gradients,
483 thereby overprinting any local signal (Bendle and Rosell-Melé, 2004; Conte et al., 2006).
484 With extremely low alkenone production due to the presence of ice at #2274, transport of
485 allochthonous alkenones within the IC likely explains the deviation in SSTs towards
486 seemingly unrealistic warmer values. In any case, the most robust aspects of the
487 biomarker data point towards near-perennial sea ice cover prior to the Holocene, although
488 the presence of both phytosterols and alkenones (albeit at low concentrations) indicates

489 the occurrence of at least partial open water conditions, potentially restricted to leads or
490 regions of partial ice melt within otherwise heavily consolidated pack ice. Such
491 conditions would likely have led to short-term and reduced primary production during
492 relatively short summer seasons and limited to the near-surface layer due to a strongly-
493 stratified water column resulting from partial ice melt. Both such uppermost surface layer
494 production conditions in leads and advection of allochthonous alkenones within the IC
495 would account for the anomalously high glacial SSTs.

496 Our conclusion of near-perennial sea ice during MIS 3 and MIS 2 is broadly
497 consistent with outcomes from a recent 120,000 yr reconstruction of sea-ice conditions
498 for the North Atlantic (Maffezzoli et al., 2019) based on the analysis of enriched bromine
499 (Br_{enr}) in an ice core from the Renland Ice Cap (RIC) 560 km WNW from #2274 (Figs. 1
500 and 7 [RIC]). Albeit at a much broader spatial resolution (i.e. 50-85° N), Maffezzoli et al.
501 (2019) proposed that MIS 3 and MIS 2 experienced a (variable) mix of multi-year and
502 first-year sea ice, before transitioning to mainly first-year ice and open water conditions
503 following the termination of the LGM. Interestingly, the greater range of sea ice cover
504 inferred from the RIC Br_{enr} record is not at all clear in our #2274 record, but is evident in
505 a biomarker record from the eastern Nordic Seas, with extensive/near-perennial sea ice
506 cover during stadials and H-events (i.e. comparable to #2274) but ice-free conditions
507 during interstadials (since ca. 90 ka BP); such differences between marine sites in the
508 western and eastern Nordic Seas presumably reflects the variable influence of warm
509 Atlantic water, limited to the eastern Nordic Seas (Hoff et al., 2016). The most prominent
510 signature of first-year ice in the Br_{enr} records occurred during the Younger Dryas and it is
511 noteworthy that a transition from permanent to increasing seasonal sea ice at the NIS was

512 reported for this interval following a biomarker-based reconstruction of surface
513 oceanographic conditions from core #2272 (Fig. 1; 7; Xiao et al., 2017). Further, based
514 on relatively high concentrations of IP₂₅ in MD99-2272 during the Younger Dryas and
515 the preceding Bølling-Allerød, Xiao et al. (2017) concluded that biomarker production
516 was likely associated with locally formed first year ice rather than from advected drift
517 ice, the latter being a feature of modern-day oceanography. In contrast, our new data
518 from #2274 indicate still near-permanent sea ice cover at this time (Fig. 7). As such, we
519 interpret the combined ice core and marine sediment core data to suggest that as climate
520 conditions ameliorated at the end of the LGM, near-permanent sea ice cover transitioned
521 to first-year seasonal sea ice in the southern part of the region, especially during the
522 Bølling-Allerød and Younger Dryas, likely due to increasing influence of the IC (Xiao et
523 al., 2017); however, the spatial extent of this area of first year ice, located southward of
524 the near-permanent sea ice front that characterizes MIS 3 and MIS 2, remains uncertain at
525 this point (see Fig. 7 sub-panel). Large-scale sea ice reduction then characterized the
526 early Holocene (Fig. 7), with a marked increase in all open water primary productivity
527 biomarker proxies (Fig. 6). Increasing drift ice subsequently became a characteristic of
528 the NIS from the mid Holocene onwards (Fig. 7; Cabedo-Sanz et al., 2016).

529

Conclusions

530 The multi-proxy sediment data from core #2274 130 km off the north Iceland coast
531 appears at first sight to yield conflicting interpretations depending on whether sediment
532 mineral composition or biomarker proxy data are being considered; however, these can
533 be resolved through a more detailed consideration of the mode(s) of iceberg drift and
534 trajectory through largely consolidated and near-pervasive sea ice. The low- resolution

535 sampling for grain-size restricts detailed interpretation but the sediments are mostly
536 moderately sorted in the silt range allowing a valid record of bottom flow speed. This
537 shows low flow speeds during H-events 1, 4 and 6 related to decrease in Nordic Sea
538 overflow, but not cessation, and a peak in the Younger Dryas.

539 The mineral composition of the < 2 mm grain-size sediment samples shows 5
540 peaks with wt% of quartz values significantly higher than Holocene values. The
541 variations in the quartz wt% are also reflected in the estimated contributions of sediment
542 from Precambrian and Caledonian bedrock sources of NE Greenland. These data require
543 sediment transport to the #2274 site during MIS 3 and MIS 2. If the transport is by
544 icebergs then the sea ice cover had to allow icebergs to drift southward, as they do at
545 present (Figs. 1C, 7). A framework of near-permanent sea ice is confirmed from ultra-low
546 seasonal sea ice and open water biomarker concentrations. On the other hand, the
547 occurrence of non-zero concentrations of some phytoplanktic biomarkers, and numbers
548 of near-surface planktonic foraminifera (Table 3) points to some short-term open water
549 conditions, either from limited sea ice melt or following the opening of leads; the
550 presence of drifting icebergs may be significant in this respect (Fig. 7).

551 An underlying question for HS H-events is whether North Atlantic-wide glacial
552 marine sediment events were triggered as a response to events in Hudson Strait or
553 whether the events are part of a shared response to broader regional oceanographic
554 conditions (e.g. Marcott, et al., 2011; Bassis et al., 2017; Velay-Vitow et al.,
555 2019). Thus, were “coeval” HS H- events on the East Greenland margin (Stein et al.,
556 1996; Andrews et al., 1998; Voelker, 1999), or lagged events (e.g. Baffin Bay: Simon et
557 al., 2014 Jennings et al., 2018), triggered in response to events in the Hudson Strait ice

558 stream? If our quartz and IRD events (Figs. 3 and 8) are indeed coeval with HS H-
559 events, this implies that the stability of ice streams on the NE and E Greenland shelf (and
560 N Iceland) and Hudson Strait may all have been affected by basin-wide subsurface
561 warming in response to a reduction in the Atlantic meridional overturning circulation
562 (Shaffer et al., 2004; Clark et al., 2007; Marcott et al., 2011).

563

564 **Acknowledgements**

565 We thank Dr Anne E. Jennings for picking and providing the foraminifera for the
566 radiocarbon dates, and Dr Haflidi Haflidasson for the identification and geochemical
567 analysis of the tephra. Professor Grant Bigg provided guidance on the role of icebergs
568 and sea ice. We thank the Centre National de la Recherche Scientifique (CNRS) for
569 MAS salary and the crew of the *RV Marion Dufresne* for coring operations during the
570 IMAGESV cruise. Data from this study will be archived at: www.Pangaea.de, along
571 with other IMAGESV data for MD99-2274. We acknowledge the availability and our
572 use of data from PS1726, PS1730, and PS2644, which we accessed through
573 www.Pangaea.de. Finally, we thank three anonymous reviewers who provided critical
574 and helpful feedback on previous versions of the manuscript.

575 Tables

576 Table 1 Location of the cores referenced in this study and showing distance from MD99-
577 2274. Cores located on Fig. 1A and B unless noted as NA. The last 5 sites are cores that
578 specify sediment sources based on radiogenic isotopic data (Verplanck et al., 2009; White
579 et al., 2016).

580

581 Table 2 A and B: Data for two possible depth/age models for MD99-2274 used in the
582 Bayesian “Bacon” model—see text. $cc = 0$ when date derived from other sources and
583 does not require calibration; $cc = 2$ when ocean reservoir correction $\Delta R = 0$ is used
584 (marine IntelCal 13; Reimer et al., 2013).

585

586 Table 3: Depth/age data and calibrated ages for radiocarbon dates on near-surface
587 planktonic foraminifera (see Figs. 1 and 5). Ocean reservoir correction $\Delta R = 0$.

588

589 **Suppl. Table:** Geochemistry of the tephra layer (see text). Courtesy Dr. H. Haflidasson)

590

591 Figure Captions

592 Figure 1: A) location of MD99-2274 and some other cores noted in the paper (Table 1)
593 (ODV, Schlitzer, 2011). The shaded areas represent the late glacial maximum (LGM)
594 extent of the ice sheets north of Denmark Strait; the words “basalt” and “felsic” define
595 the primary sediment mineral sources and the arrows show probable flow paths for
596 icebergs. BB = Blosseville Basi; TMF = Scoresby Sund Trough Mouth Fan; B)
597 Additional cores referenced in the paper (see also Table 1). Note that “Cald” on this
598 figure references the southern outcrop of the Greenland Caledonides (Higgins et al.,

599 2008). SS = Scoresby Sund; RIC = Renland Ice Cap. C) Surface and bottom currents and
600 historical April sea-ice edge (1870-1920) (dashed white line; Divine and Dick, 2006).
601 NIIC = North Iceland Irminger Current; EGC = East Greenland Current; EIC = East
602 Iceland Current; Yellow lines: Bottom Currents DSOW = Denmark Strait Overflow
603 Water; NIJ = North Iceland Jet., S = Separated East Greenland Current; OC = Iceland Sea
604 Ocean Convection site (after Harden et al., 2016).

605

606 Figure 2: A) Downcore plot of magnetic susceptibility (SI^5) and Bayesian ((Blaauw and
607 Christen, 2016) depth age plots for MD99-2274 (see Table 2)---the red curve is for the
608 initial available data blue curve is for the estimated ages with the addition of an estimated
609 core top age and the presence of the Vedde and NAAZII tephras (see text). The Marine
610 Isotope Stage (MIS) boundaries are indicated. Location of radiocarbon dates and tephras
611 are noted. B) Plot of the departures from the median values of magnetic susceptibility
612 ($2.03 * 10^{-3} SI$) and quartz wt% (5.3). Note that the quartz axis is reversed.

613

614 Figure 3: Variation in the Sortable Silt mean size (3-point 1-2-1 weighted
615 smoothing with raw data dots) and IRD% >240 μm . Minima in \overline{SS} are seen at the time
616 of Hudson Strait H events -H6, -H4 and -H1 while -H4, -H2, early -H1 and the YD (-H0)
617 are marked by elevated IRD %. Blue bars are regions where the data are unreliable
618 indicators of flow speed according to the \overline{SS} -SS% correlation criterion of McCave and
619 Andrews, (2019a)

620

621 Figure 4: Plots of the variations in the weight% of minerals in MD99-2274, the
622 quartz/pyroxene ratio, and magnetic susceptibility. The green shaded areas represent
623 Holocene values, hence points above represent departures. Numbers 1 through 5 identify
624 IRD quartz peaks. The vertical blue shading areas represent times when the weight% of
625 quartz exceeds Holocene limits.

626

627 Figure 5: Plots of the sediment source percentages and the degree of fit (DOF), that is the
628 average absolute bias in the SedUnMix calculation of (observed mineral wt% - predicted
629 mineral weight%) for each sample. The top panel shows the location of measurable
630 quantities of gravel, and sites of tephra layers and the radiocarbon dates on near-surface
631 planktonic foraminifera (Table 3). Numbers on the NE Greenland panel represent the
632 peaks in that source and the yellow bars locate areas with minimal input from that area.

633

634 Figure 6: Biomarker data (A) IP₂₅ and HBI II concentrations; (B) $\sum C_{37:3} + C_{37:2}$ alkenone
635 and brassicasterol concentrations; C) SST° C estimates and the %C_{37:4}; and D) Weight %
636 quartz and different coarse sediment fractions.

637

638 Figure 7: Schematic presentation of changes in sea ice and iceberg distribution. The first
639 panel (upper left) shows core locations (see Table 1 and Fig. 1A and B) and the adjoining
640 panel the inferred conditions during MIS 3 and 2 with pervasive sea ice and embedded
641 icebergs. The remaining panels show the proposed evolution in the state of sea ice and
642 iceberg supply (red triangles) during deglaciation into the Holocene (adapted from

643 Cabedo-Sanz et al., 2016; Xiao et al., 2017). SS =Scoresby Sound, RIC=Renland Ice
644 Cap.

645

646

647 Figure 8: Analysis of the quartz wt% records from PS2644 (Vogt, 2017) and MD99-2274
648 at a common 0.6 ky spacing. A) Original quartz data (black line) and the wavelet
649 reconstructions for the two records; B) Cross-wavelet power spectrum of quartz wt% for
650 PS2644 and MD99-2274. The cone of confidence indicated by the light grey areas;
651 0.05% probability area demarcated by white line. Arrows pointing to the right mean that
652 the two records are in phase, arrows pointing down mean that x leads y, arrows pointing
653 to the left indicate the records are anti-phase and pointing up indicates that #2274 leads
654 PS2644. C) Cross-wavelet (Fig. 8B) average power. The 0.05 significance period is red
655 and delimited by the dashed slanting line. The horizontal dashed line indicates the peak
656 periodicity (~8.5 ky).

657

658

659 Suppl. Figure 1: Data for VM30-130 (see Fig. 1 and Table 3).

660

661 Suppl. Figure 2: Showing the reduced major axis association between sortable silt mean
662 size (\overline{SS}) and SS%.

663

664 Suppl. Figure 3: $\delta^{18}\text{O}$ *N. pachyderma* plots of cores from the Blosseville Basin/Scoresby
665 Sund Trough Mouth Fan (see Fig. 1 and 8) from cores PS1730 (Stein et al., 1996a,b,
666 and PS2644 (Voelker, 1999).

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Methods

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1179 **Magnetic susceptibility:** Magnetic susceptibility was measured on-board the *Marion*
1180 *Dufresne* (Labeyrie and Cort, 2005) in 2-cm increments (hence ~150yr sampling on
1181 average). Measurements were taken on the 1.5 m core sections. In this area of Iceland,
1182 the marine deposits are strongly affected by erosion and transport of basalt, which results
1183 in very high values of magnetic susceptibility. The export of sediments from the erosion
1184 of bedrock with much lower magnetic susceptibilities, such as granites and other felsic-
1185 rich bedrock in NE Greenland and from more distant sources (Verplanck et al., 2009;
1186 White et al., 2016) will lower the magnetic susceptibility readings. It is important to note
1187 that although magnetic susceptibility is straightforward to measure, data interpretation is
1188 complex, being a product of sediment density, grain-size, and mineralogy (Robinson et
1189 al., 1995; Stoner and Andrews, 1999; Watkins and Maher, 2003).

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Quantitative X-ray Diffraction (qXRD): The weight % (wt%) of the non-clay
and clay mineral composition of the < 2 mm sediment fractions is based on the US
Geological Survey method (Eberl, 2003), which has been used extensively in this region
(e.g. Andrews et al., 2017; Andrews and Eberl, 2007; Andrews and Vogt, 2014). One
gram of sediment (dry weight) is spiked with 0.111 g of zincite, prepared (Eberl, 2003),
run in the X-ray diffractometer, and the resulting intensity data processed in the Excel
macro-program Rockjock v6. We investigate the wt% and presence/absence of 34
minerals and reduced this number by combining individual mineral wt% into larger
groups, such as k-feldspars, plagioclase, dolomite, and amorphous minerals. Importantly
in the context of this paper we had earlier shown that qXRD can recognize the presence

1200 of tephra and volcanic glass, with some ability to distinguish between basaltic and
1201 rhyolitic glass (Andrews et al., 2013).

1202 To gain a better understanding of possible changes in the provenance of the
1203 mineral compositions we processed the mineral wt% data in a sediment unmixing
1204 program “SedUnMix” (Andrews and Eberl, 2012). Two models were considered, the first
1205 with qXRD results from #2274 with four appropriate bedrocks, namely: basalt, dolerite,
1206 gneiss, and granite; and secondly with the mineral compositions of glacial marine
1207 sediment samples from potential source areas, namely: NE Greenland, E. Greenland, and
1208 Iceland (Suppl. Table of bedrock and marine sediment sources). The program calculates a
1209 “degree of fit” and also derives error estimates on each source within a sample. Ideally,
1210 the sum of the sources should equal 100% but marked deviations from this suggest that
1211 one or more sources have not been included, and/or that the sources are not representative
1212 of the sediment samples.

1213 **Grain-size:** Sediment was wet-sieved at 2 mm and the grain-size volume
1214 percentages in 96 intervals between 0.01 and 2000 μm were obtained via a Malvern laser
1215 system. Comparisons between the Malvern and other grain-size systems have been
1216 documented and found comparable (McCave et al., 2006; McCave and Syvitski, 1991).
1217 However, the objections of McCave et al. (2006) to laser sizers on the grounds of grain
1218 shape (Konert and Vandenberghe, 1997) are not valid for equant grains such as those
1219 produced by glacial grinding, as pointed out by Piper (Marshall et al., 2014), and thus
1220 size data are believed valid in the setting of MD2274. Grain-size curves have provided
1221 vital information on sediment transport and deposition in this region, and methods have
1222 been developed to reconstruct variations in bottom current speed for sediments delivered

1223 to the ocean from dominantly glacial sources (McCave and Andrews, 2019a, b) The
1224 calibration of sortable silt mean (mean of 10-63 μm), a sensitivity, by McCave et al.,
1225 (2017) has been applied to changes in the grainsize record.

1226 **Biomarkers:** Biomarkers were extracted from freeze-dried subsamples (~2-4 g).
1227 Prior to extraction, samples were spiked with an internal standard (9-octylheptadec-8-ene,
1228 9-OHD, 10 μL ; 10 $\mu\text{g mL}^{-1}$) to permit quantification of the highly branched isoprenoid
1229 (HBI) biomarkers IP₂₅, HBI II and HBI III. 5 α -androstan-3 β -ol; (0.1 μg) was also added
1230 to permit quantification of brassicasterol in some cases. Samples were then saponified in
1231 a methanolic KOH solution (~5 mL H₂O:MeOH (1:9); 5% KOH) for 60 min (70 °C).
1232 Hexane (3 \times 2 mL) was added to the saponified mixtures, with supernatant solutions,
1233 containing non-saponifiable lipids (NSLs), transferred by glass pipettes to glass vials, and
1234 solvent removed using a gentle stream of N₂. Dried NSLs were re-suspended in hexane
1235 (0.5 mL) and fractionated using column chromatography (SiO₂; 0.5 g). Non-polar lipids,
1236 including IP₂₅ and HBI II, were eluted with hexane (6 mL), while more polar lipid
1237 fractions containing alkenones were eluted with MeOH (6 mL). For a few horizons,
1238 additional NSLs were fractionated to yield non-polar (hexane; 6 mL) and polar fractions
1239 containing sterols (hexane:methyl acetate 4:1; 6 mL). Each non-polar fraction was further
1240 purified to remove saturated components using silver-ion chromatography (Belt et al.,
1241 2015), with saturated compounds eluted with hexane (2 mL) and unsaturated compounds,
1242 including IP₂₅ and other HBIs, collected in a subsequent acetone fraction (3 mL).
1243 Analysis of fractions containing IP₂₅ and other HBIs was carried out using gas
1244 chromatography–mass spectrometry (GC–MS) following the methods and operating
1245 conditions described previously (Belt et al., 2012). Mass spectrometric analysis was

1246 carried out in total ion current (TIC) and selected ion monitoring (SIM) modes. The
1247 identification of IP₂₅ and HBI II was based on their characteristic GC retention indices
1248 (e.g. RI_{HP5MS} = 2081,2082 and 2044 for IP₂₅, HBI II and HBI III, respectively) and mass
1249 spectra (Belt, 2018). Quantification of all HBIs was achieved by comparison of mass
1250 spectral responses of selected ions (e.g. IP₂₅, *m/z* 350; HBI II, *m/z* 348; HBI III, *m/z* 346)
1251 in SIM mode with those of the internal standard (9-OHD, *m/z* 350) and normalized
1252 according to their respective instrumental response factors, derived from solutions of
1253 known biomarker concentration, and sediment masses (Belt et al., 2012). Fractions
1254 containing sterols were derivatized with N,O-bis(trimethylsilyl)trifluoroacetamide
1255 (BSTFA; 100 µL; 70°C for 60 min) immediately prior to analysis by GC–MS. Sterols
1256 were identified by comparison with GC–MS responses compared to those of standards.
1257 Sterol quantification was achieved as per the approach described above for HBIs.

1258 Polar fractions containing alkenones obtained from elution with MeOH (6 mL) were
1259 further purified with 2 mL of hexane:methyl acetate (95:5 v/v) and 2 mL of hexane:methyl
1260 acetate (90:10 v/v). Alkenones were analyzed using a Thermo Trace GC Ultra gas
1261 chromatograph equipped with a CPSil5 capillary column (50m length, 0.32 i.d. and 0.25
1262 mm film thickness), an FID detector and a septum programmable injector (SPI). Helium
1263 was used as carrier gas. 5α-cholestane was added as an external standard prior to GC
1264 injection. SST estimates were determined using the following equation (Prahl et al., 1988).

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$$U_{37}^{K'} = \frac{C_{37:2}}{C_{37:2} + C_{37:3}} = 0.034 T + 0.039$$

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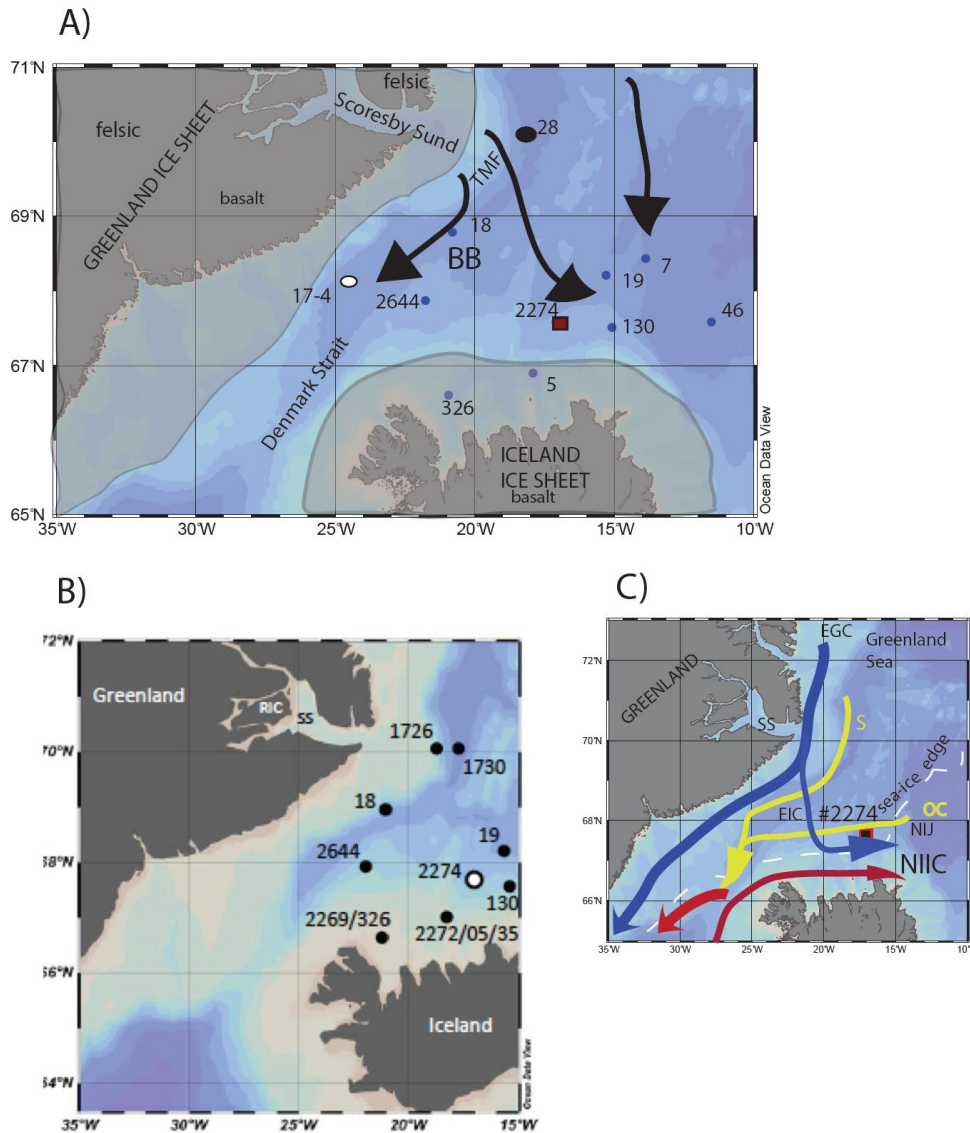
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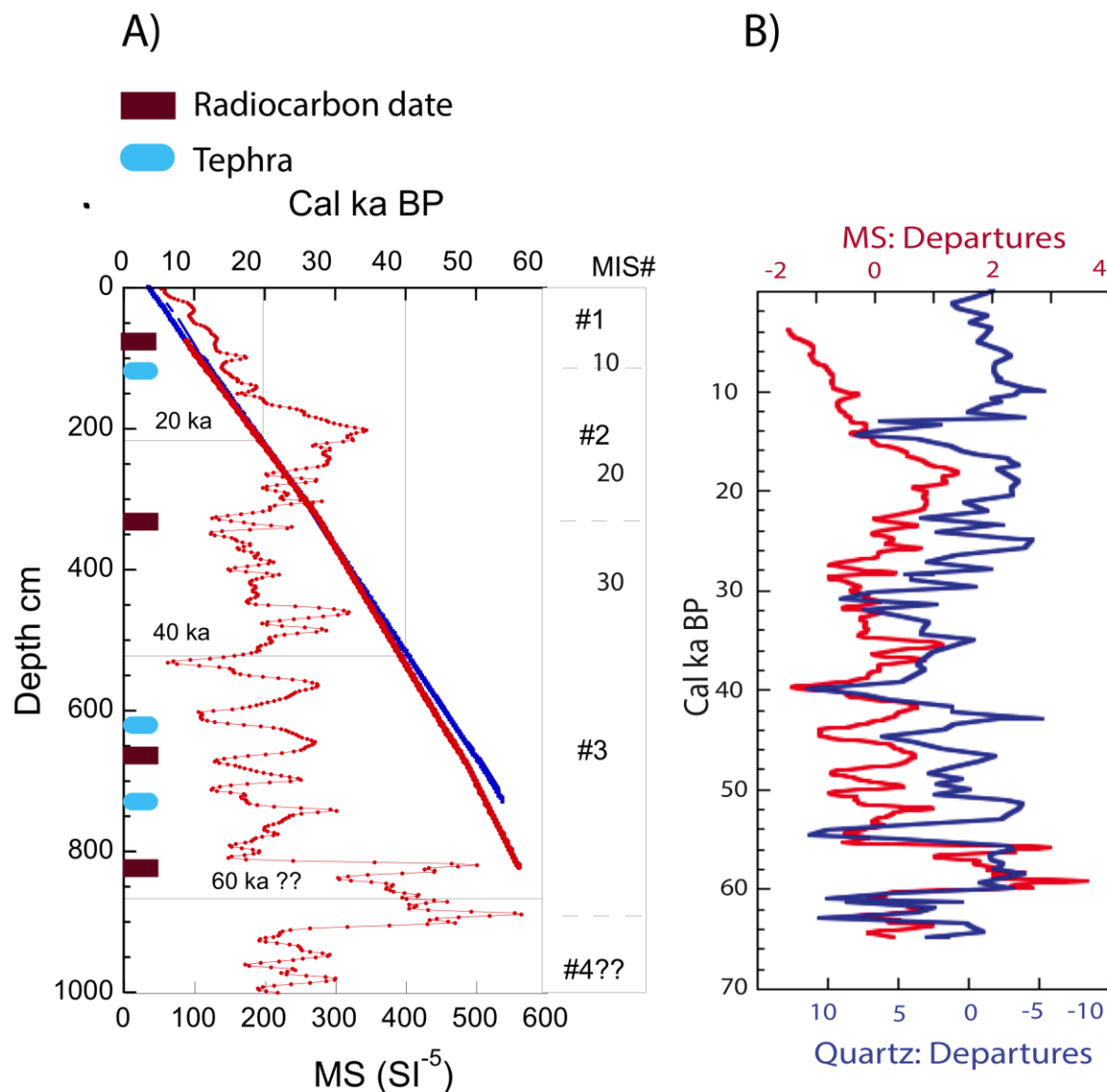
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1339 **Figures with Captions**
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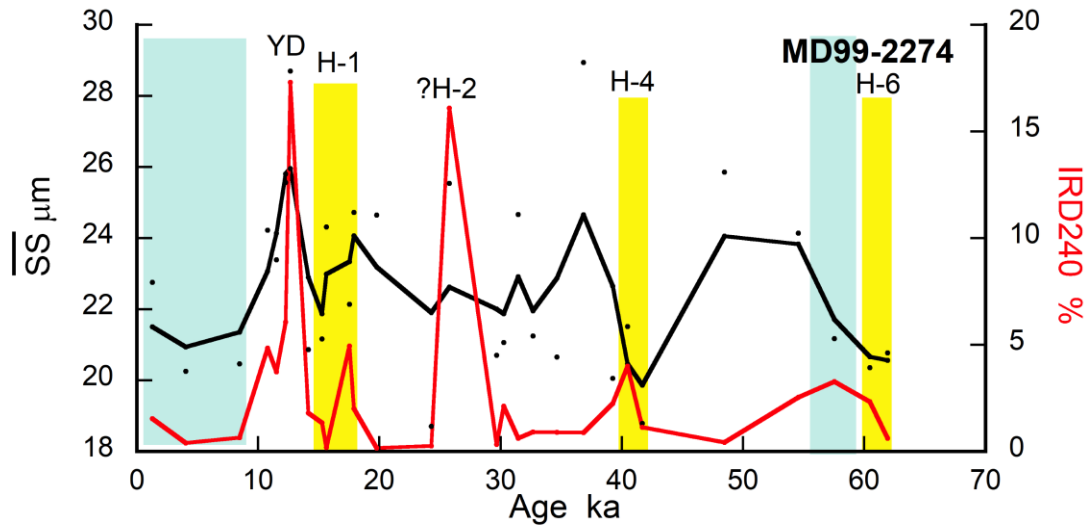


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 1342 Figure 1: A) location of MD99-2274 and some other cores noted in the paper (Table 1)
 1343 (ODV, Schlitzer, 2011). The shaded areas represent the late glacial maximum (LGM)
 1344 extent of the ice sheets north of Denmark Strait; the words “basalt” and “felsic” define
 1345 the primary sediment mineral sources and the arrows show probable flow paths for
 1346 icebergs. BB = Blosseville Basi; TMF = Scoresby Sund Trough Mouth Fan; B)
 1347 Additional cores referenced in the paper (see also Table 1). Note that “Cald” on this
 1348 figure references the southern outcrop of the Greenland Caledonides (Higgins et al.,
 1349 2008). SS = Scoresby Sund; RIC = Renland Ice Cap. C) Surface and bottom currents and
 1350 historical April sea-ice edge (1870-1920) (dashed white line; Divine and Dick, 2006).
 1351 NIIC = North Iceland Irminger Current; EGC = East Greenland Current; EIC = East
 1352 Iceland Current; Yellow lines: Bottom Currents DSOW = Denmark Strait Overflow
 1353 Water; NIJ = North Iceland Jet., S = Separated East Greenland Current; OC = Iceland Sea
 1354 Ocean Convection site (after Harden et al., 2016).



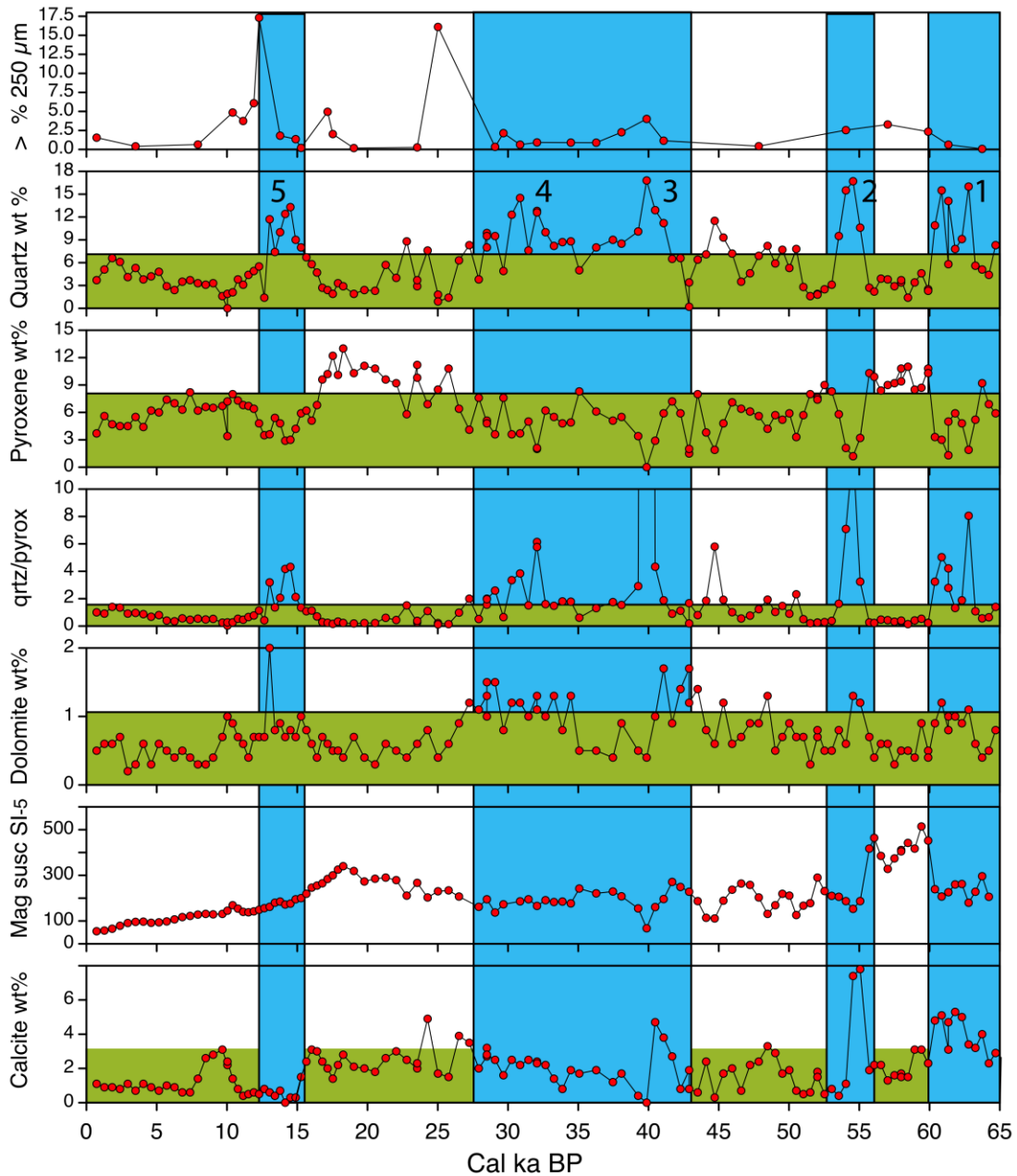
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Figure 2: A) Downcore plot of magnetic susceptibility (SI⁻⁵) and Bayesian ((Blaauw and Christen, 2016) depth age plots for MD99-2274 (see Table 2)---the red curve is for the initial available data blue curve is for the estimated ages with the addition of an estimated core top age and the presence of the Vedde and NAAZII tephras (see text). The Marine Isotope Stage (MIS) boundaries are indicated. Location of radiocarbon dates and tephras are noted. B) Plot of the departures from the median values of magnetic susceptibility (2.03×10^{-3} SI) and quartz wt% (5.3). Note that the quartz axis is reversed.



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 1365 Figure 3: Variation in the Sortable Silt mean size (3-point 1-2-1 weighted
 1366 smoothing with raw data dots) and IRD% >240 μm . Minima in \overline{SS} are seen at the time
 1367 of Hudson Strait H events -H6, -H4 and -H1 while -H4, -H2, early -H1 and the YD (-H0)
 1368 are marked by elevated IRD %. Blue bars are regions where the data are unreliable
 1369 indicators of flow speed according to the \overline{SS} -SS% correlation criterion of McCave and
 1370 Andrews, (2019a)

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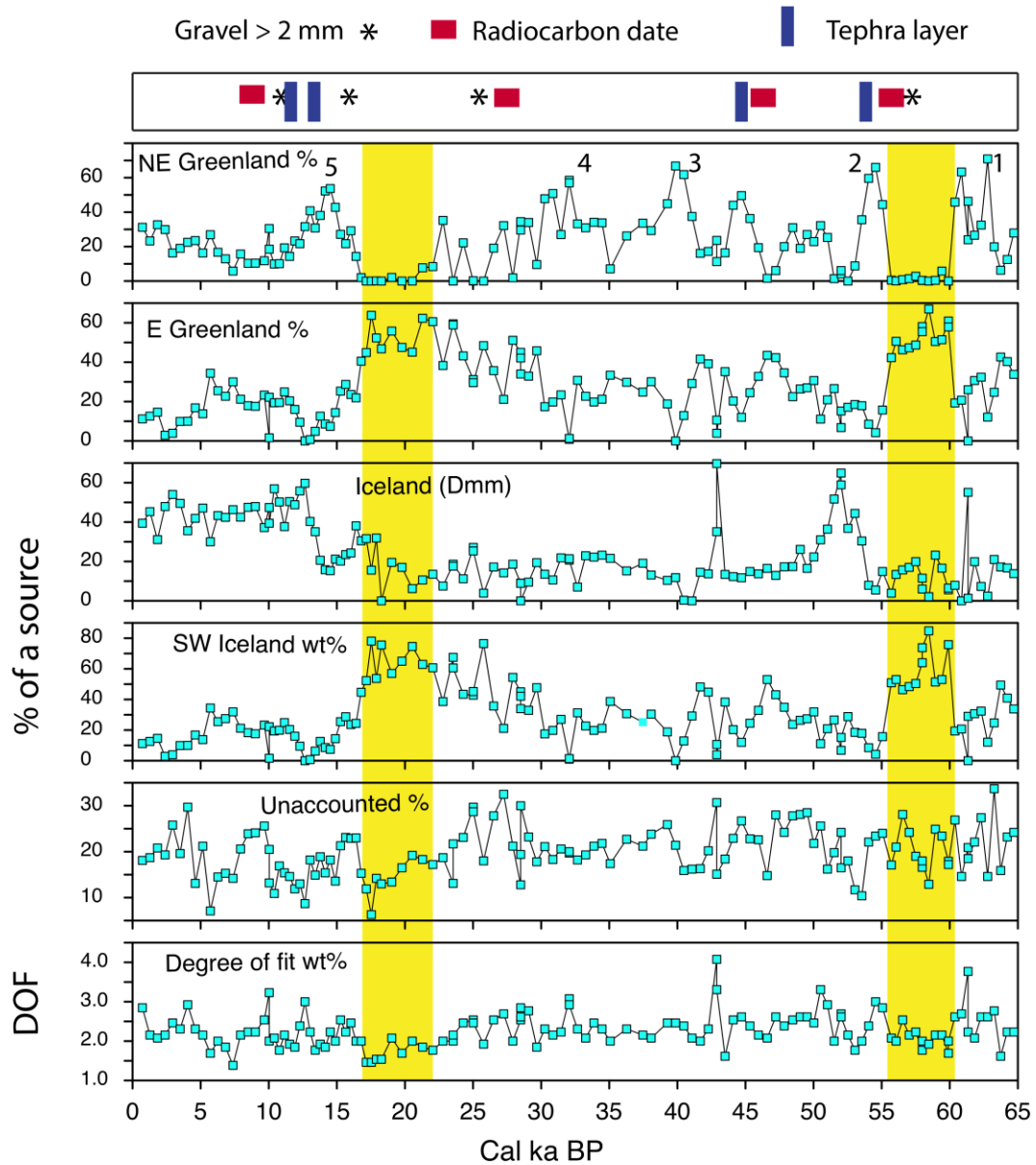
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Figure 4: Plots of the variations in the weight% of minerals in MD99-2274, the quartz/pyroxene ratio, and magnetic susceptibility. The green shaded areas represent Holocene values, hence points above represent departures. Numbers 1 through 5 identify IRD quartz peaks. The vertical blue shading areas represent times when the weight% of quartz exceeds Holocene limits.



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1380 Figure 5: Plots of the sediment source percentages and the degree of fit (DOF), that is the

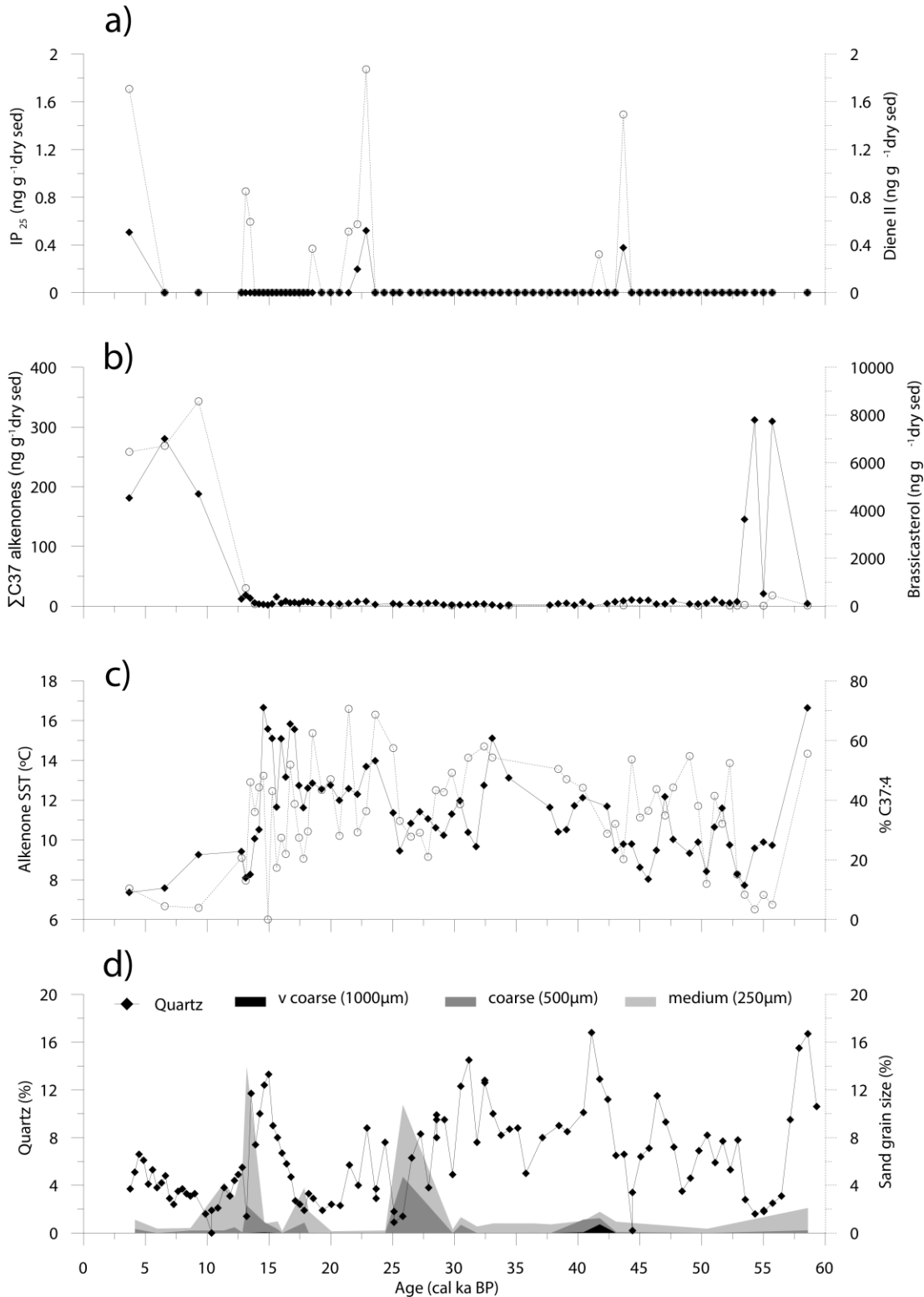
1381 average absolute bias in the SedUnMix calculation of (observed mineral wt% - predicted

1382 mineral weight%) for each sample. The top panel shows the location of measurable

1383 quantities of gravel, and sites of tephra layers and the radiocarbon dates on near-surface

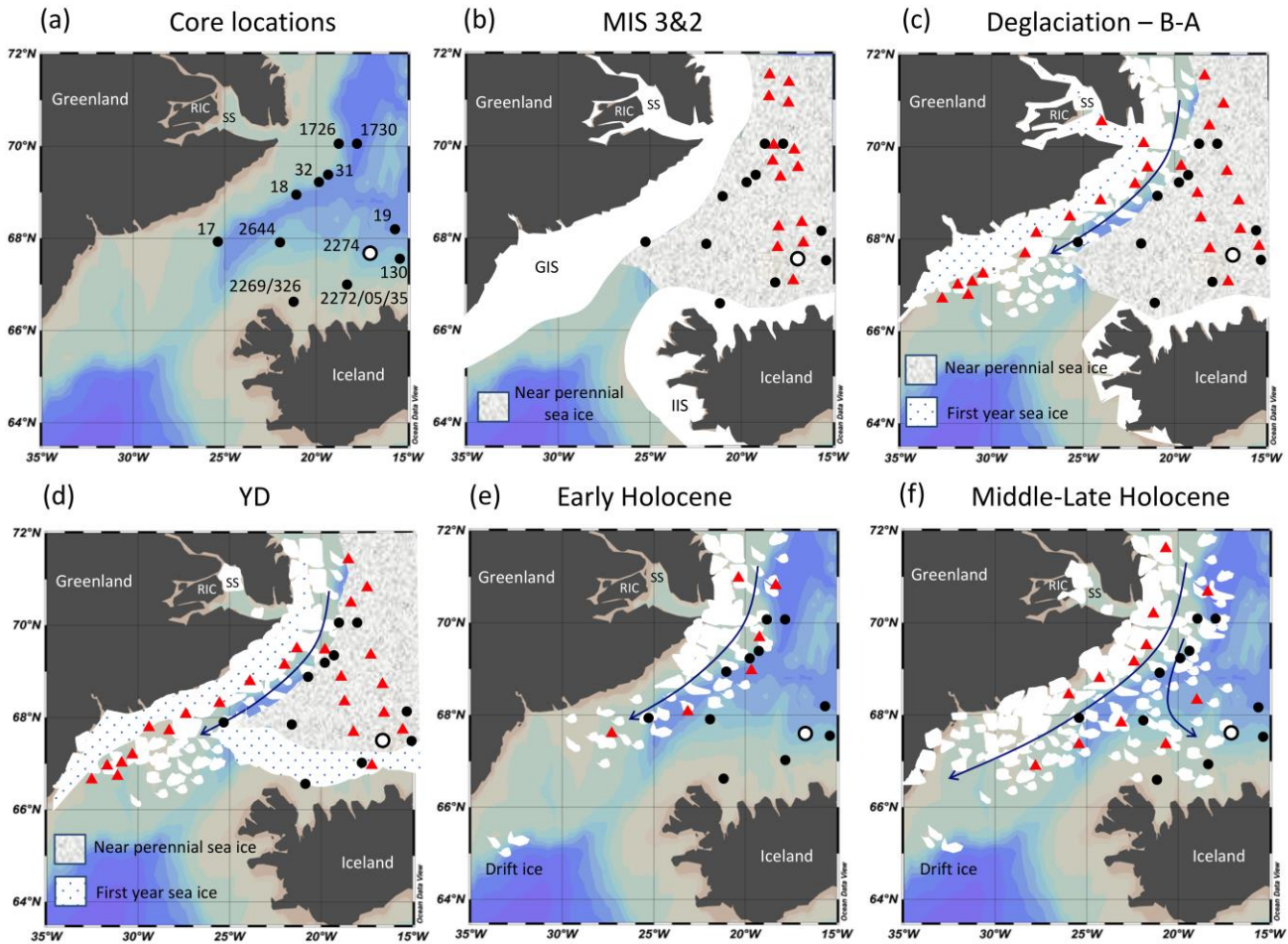
1384 planktonic foraminifera (Table 3). Numbers on the NE Greenland panel represent the

peaks in that source and the yellow bars locate areas with minimal input from that area.

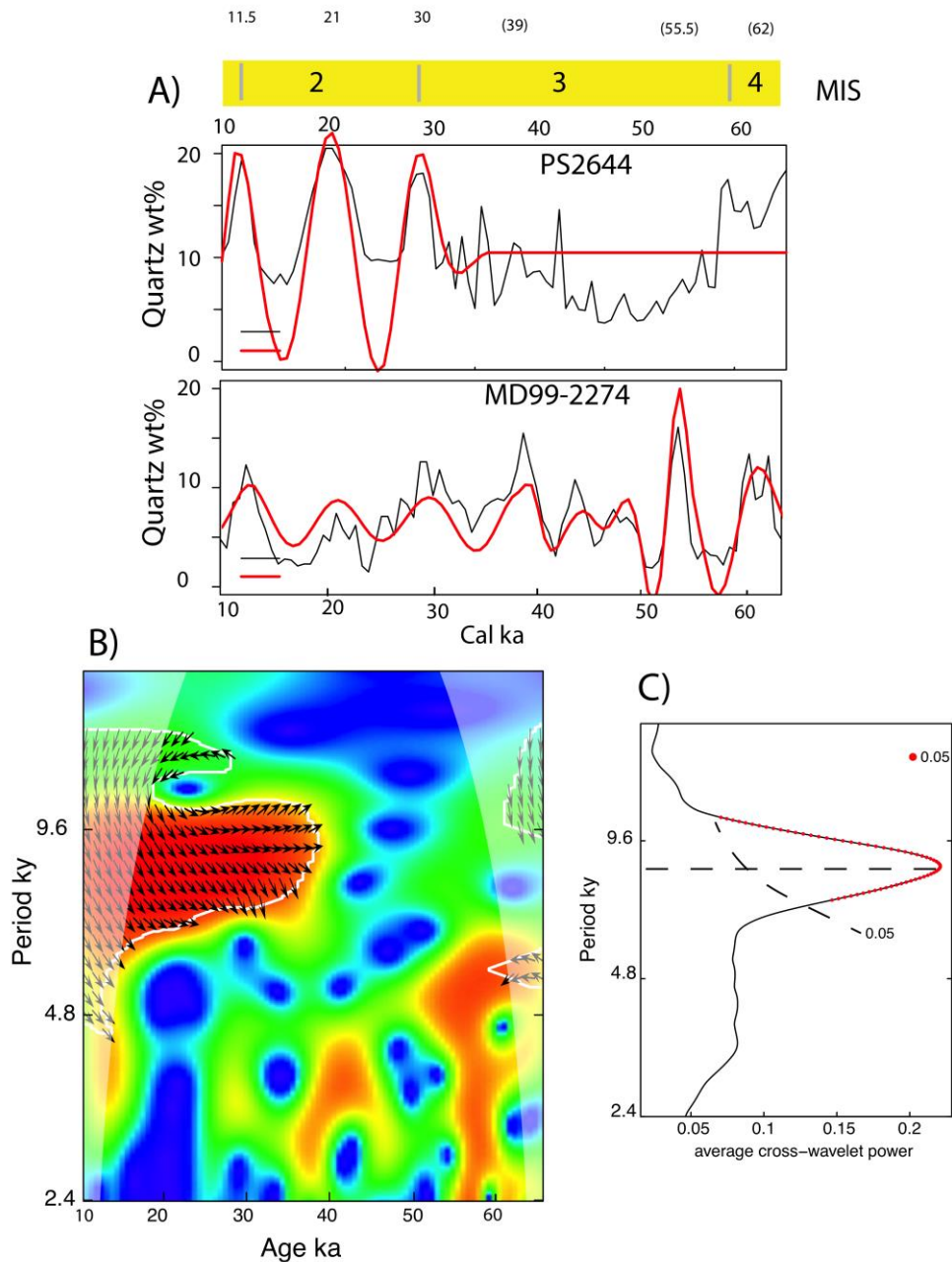


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Figure 6: Biomarker data (A) IP_{25} and HBI II concentrations; (B) $\Sigma C_{37:3} + C_{37:2}$ alkenone and brassicasterol concentrations; (C) SST °C estimates and the %C_{37:4}; and (D) Weight % quartz and different coarse sediment fractions.



1390 Figure 7: Schematic presentation of changes in sea ice and iceberg distribution. The first
 1391 panel (upper left) shows core locations (see Table 1 and Fig. 1A and B) and the adjoining
 1392 panel the inferred conditions during MIS 3 and 2 with pervasive sea ice and embedded
 1393 icebergs. The remaining panels show the proposed evolution in the state of sea ice and
 1394 iceberg supply (red triangles) during deglaciation into the Holocene (adapted from
 1395 Cabedo-Sanz et al., 2016; Xiao et al., 2017). SS = Scoresby Sound, RIC = Renland Ice
 1396 Cap.
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Figure 8: Analysis of the quartz wt% records from PS2644 (Vogt, 2017) and MD99-2274 at a common 0.6 ky spacing. A) Original quartz data (black line) and the wavelet reconstructions for the two records; B) Cross-wavelet power spectrum of quartz wt% for PS2644 and MD99-2274. The cone of confidence indicated by the light grey areas; 0.05% probability area demarcated by white line. Arrows pointing to the right mean that the two records are in phase, arrows pointing down mean that x leads y, arrows pointing to the left indicate the records are anti-phase and pointing up indicates that #2274 leads PS2644. C) Cross-wavelet (Fig. 8B) average power. The 0.05 significance period is red and delimited by the dashed slanting line. The horizontal dashed line indicates the peak periodicity (~8.5 ky).