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1 **Sea ice, ice-rafting, and ocean climate across Denmark Strait during rapid**
2 **deglaciation (~16 to 12 cal ka BP) of the Iceland and East Greenland shelves**
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Abstract

A suite of cores from the Northwest Iceland and East Greenland shelves sampled fossiliferous or unfossiliferous basal glacial diamictons. Radiocarbon dates above the diamictons are similar on both shelves, however, the value of the ocean reservoir correction, ΔR , is unknown. Deglaciation occurred either ~16 or 14 cal ka BP depending on the choice of ΔR . The ice sheets were behind the present coastline by 12.2 cal ka BP. We examine seven cores that record the glacial/deglacial transition and present new data on the sea-ice biomarkers IP_{25} and $C_{25:2}$ from four of the cores plus data on IRD counts, grain-size spectra, $\delta^{18}O$ on the near-surface planktonic foraminifera *Neoglobobulimina* *pachyderma* (s), foraminifera assemblages, and quartz %. IP_{25} concentrations are markedly higher for the East Greenland sites, while they are frequently below the limit of quantification off Iceland; observations that parallel the wt% quartz in the sediments. The $\delta^{18}O$ *N. pachyderma* (s) data show a strong gradient across Denmark Strait with lighter $\delta^{18}O$ values towards the East Greenland shelf indicative of a large freshwater flux. The presence of the chilled Atlantic Water benthic foraminifera, *Cassidulina neoteretis*, indicates that rapid ice sheet retreat was associated with ocean forcing, combined with other factors.

Keywords: Denmark Strait, Deglaciation, Ocean reservoir correction, IP_{25} biomarker, Sedimentology

Abstract: 254 words

Main text: 6455 words

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49 **Introduction**

50 A significant number of cores from around Iceland and E Greenland penetrate a basal
51 diamicton, often containing shells and foraminifera, overlain by glacial marine sediments
52 rich in ice rafted debris (IRD) (Jennings et al., 2000; Olafsdottir, 2004). At the Late
53 Glacial Maximum (LGM), reconstructions and marine core data (Andrews, 2008;
54 Andrews et al., 1998, 2000; Dunhill, 2005; Funder et al., 2004; Hubbard et al., 2006;
55 Vasskog et al., 2015) indicate that the Iceland and Greenland ice sheets were terminating
56 at their shelf breaks with deposition on the slopes above the Denmark Strait. In a
57 companion paper (Andrews et al., 2017a) we documented the variations of the Iceland
58 and Greenland ice sheets during Marine Isotope Stage (MIS) 3 and LGM as archived in
59 sediments at the northern and southern ends of Denmark Strait. Active sediment
60 deposition ceased on the Kangerlussuaq Trough Mouth Fan ca. 15.3 ka ¹⁴C BP (Andrews
61 et al., 1998; Dunhill, 2005) and retreat to the present outer coast occurred prior to
62 deposition of the Vedde tephra (Jennings et al., 2006). Geophysical data from the
63 Kangerlussuaq Trough, E Greenland (Dowdeswell et al., 2010; Stein, 1996), and from the
64 West Iceland shelf (Syvitski et al., 1999) indicate that there are sites where pre Last
65 Glacial Maximum (LGM) sediments exist, but no such sites have been successfully
66 cored. Syvitski et al (1999), Norddahl and Ingolfsson (2015), and Petursson et al. (2015)
67 argued that the Iceland Ice Sheet retreated rapidly driven by a rapid rise in relative sea
68 level. Jennings et al. (2006) presented evidence for a rapid retreat of the Greenland Ice
69 Sheet under the influence of subsurface Atlantic Water from marine cores along
70 Kangerlussuaq Trough (KT, Fig. 1A). Thus evidence for rapid ice retreat has been
71 presented for both the Greenland and Iceland ice sheets citing different reasons for the

72 retreat and modeling studies of deglaciation (Patton et al., 2017; Tarasov and Peltier,
73 2002) do not necessarily focus on all multi-proxy evidence. However, in order to
74 understand the forcing mechanism(s) driving ice sheet retreat, the radiocarbon dates have
75 to be adjusted by an ocean reservoir correction that probably varied both temporally and
76 spatially (Stern and Lisiecki, 2013).

77 The purpose of this paper is to use the paleoceanographic records from sites on
78 either side of the Denmark Strait for the time period between deglaciation and deposition
79 of the Vedde tephra ca $12.1 \pm$ cal ka BP in order to explore the potential drivers of ice
80 sheet retreat and to test for the synchronicity of Iceland and Greenland ice sheet retreat.
81 Our focus is on determining conditions during deglaciation, and especially the evidence
82 pertaining to the presence of sea ice and icebergs, and whether we can detect evidence for
83 ocean forcing, which may explain some of the differences between field and modeling
84 data (Lecavalier et al., 2014; Sinclair et al., 2016). The cores are from the N Iceland
85 Húnaflói and Djúpáll troughs, and the E Greenland Kangerlussuaq Trough (Fig. 1A,
86 Tables 1 and 2). Although evidence for ice retreat has been presented separately for both
87 the Greenland and Iceland ice sheets, here we examine the combined regional evidence
88 across Denmark Strait using multiple proxy data, with an emphasis on reconstructing
89 deglacial environmental conditions by the use of sea ice biomarkers, foraminifera, light
90 oxygen isotope data, and sediment properties. Biomarker and sediment mineralogy are
91 new here, whereas the other variables have been reported previously. Data from
92 PO175/1-5 (Mienert, 1990) have not been previously published. Of particular interest,
93 given the evidence for present-day ocean warming impacting the retreat of Greenland
94 tidewater glaciers (Andresen et al., 2011, 2012; Holland et al., 2008), is whether the rapid

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95 retreat of the ice sheets in our area can also be attributed to the incursion of Atlantic
96 Water (Jennings et al., 2006; Knudsen et al., 2003) or if there are other important factors.
97 Regardless of the exact chronology (see below) the sequence of events and regional
98 variations will remain intact.

99 **Background to study region**

100 Atlantic Water in the form of the Irminger Current flows as the surface current northward
101 along the West Iceland shelf and splits into two branches, one of which flows eastward
102 along the inner N Iceland shelf, whereas the other branch swings south and is present as
103 an intermediate water mass in the large E and SE Greenland troughs (Fig. 1A) (Andresen
104 et al., 2011; Jennings et al., 2011; Kraus, 1958; Syvitski et al., 1996). On the East
105 Greenland shelf the surface current, the East Greenland Current, consists of a cold,
106 relatively low salinity water mass derived from the Arctic Ocean: the East Iceland
107 Current is a branch of this current (Fig. 1A). Under present conditions there is a
108 pronounced tongue of cold surface water that extends to the east of Iceland (Siedov et al.,
109 2016). Sea ice can be present on the Iceland shelves (Divine and Dick, 2006; Ogilvie,
110 1992; Ogilvie and Jonsdottir, 2000) bringing in driftwood from Siberia (Eggertsson,
111 1993; Hellmann and al., 2015), while on the East Greenland shelf sea ice is present for
112 many months of the year (Hastings, 1960). There are no tidewater glaciers on Iceland
113 today whereas on E Greenland there are many tidewater glaciers contributing icebergs
114 onto the inner shelf (Bigg, 1999; Seale et al., 2011).

115 Our study area lies across the Iceland-Greenland Ridge (Larsen, 1983). Iceland is
116 composed of mid-Tertiary to Recent basalt and other extrusive volcanics (Hardarson et
117 al., 2008), in which the occurrence of quartz and high-pressure feldspars is non-existent

118 or rare. The occurrence of quartz on the Iceland shelf has been used as an indicator of the
119 incursion of drift ice (i.e. icebergs and sea-ice) (Eiriksson et al., 2000; Moros et al., 2006;
120 Thors, 1974). The bedrock geology on the East Greenland side of Denmark Strait is
121 more complex, with a large (60,000 km²) outcrop of early Tertiary basalt between
122 Kangerlussuaq Fjord and the southern coast of Scoresby Sund (Brooks and Nielsen,
123 1982; Brooks, 2008) (Fig. 1A). Cretaceous sandstones and shales exposed in the fjords
124 (Larsen et al., 1999), and Precambrian basement granite gneisses that form most of the
125 hinterland bedrock (Henriksen, 2008), are rich in quartz and k-feldspars which are
126 common components of late Quaternary marine sediments (Andrews et al., 2014, 2015).

127 Analysis of seafloor surface sediments from the N Iceland shelf indicated that
128 quartz and the sea ice biomarker IP₂₅ (Belt et al, 2007) are present and are well correlated
129 with each other, but both are absent in sediments from the W and SW Iceland shelf
130 (Cabedo-Sanz et al., 2016). A paper by Xiao et al. (2017) was recently published on the
131 variations in IP₂₅ in core MD99-2272 just to the east of our B997-326 site and at the same
132 site as HM107-05 (Fig. 1).

133 **Methods**

134 The cores (Table 1) were obtained on cruises of the *Poseidon* (1991) (Mienert, 1990), *Jan*
135 *Mayen* (1996) (Hald, 1996), *Bjarni Sæmundsson* (1997) (Helgadottir, 1997) and *Marion*
136 *Dufresne* (1999) (Labeyrie et al., 2003). Detailed descriptions of the methods are given
137 as Supplemental Material and the data are available from Pangeae (www.pangeae.de; see
138 Andrews, Cabedo Sanchez, et al., 2017b). Counts of the > 2 mm fraction (Grobe, 1987)
139 are a measure of ice-rafting by icebergs, and the Malvern grain-size spectra (Suppl.
140 Material) also provide evidence on the mode of sediment transport and deposition

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141 (Jonkers et al., 2012). The weight % of quartz on the Iceland Shelf provides a measure of
142 the importation of far-travelled ice-rafted sediments (Andrews, 2009; Eiriksson et al.,
143 2000) whereas calcite wt% is a measure of marine productivity (Thordardottir, 1984,
144 1986). Regional assessments of oceanographic controls on foraminifera species
145 (Jennings and Helgadottir, 1994; Jennings et al., 2004a; Rytter et al., 2002), stable
146 isotope values (Smith et al., 200; Azetsu-Scott and Tan, 1997), and the sea ice biomarker
147 IP₂₅ (Cabedo-Sanz et al., 2016) provide critical information on paleoceanography.

148 **Marine Isotope Stage 2 (MIS2 & 1) radiocarbon dates:** INSTAAR Date Lists
149 (Dunhill et al., 2004; Manley and Jennings, 1996; Quillmann et al., 2009; Smith and
150 Licht, 2000; Andrews et al., in prep) contain 82 dates on marine carbonates (shells and
151 foraminifera) ≥ 12.0 ¹⁴C ka BP from the region. Additional dates on near surface
152 planktonic foraminifera, benthic foraminifera, molluscs, and mixed benthic and
153 planktonic foraminifera have been obtained from other papers (e.g. Knudsen et al., 2003,
154 2004)(Suppl. Table 1). Many of the dates noted above and in Supplemental Table 1 were
155 used in the ice sheet modeling study (Patton et al., 2017, see their Table 1). Note that we
156 will refer to the ¹⁴C date without marine reservoir correction as “¹⁴C ka BP” and the
157 calibrated dates, after correction, as “cal ka BP.” We focus on cores that penetrated a
158 basal diamicton or have a robust chronology (Fig. 1, Tables 1 and Suppl. Table 1). We
159 do not attempt to “tune” our data to other records, such as the Renland Ice Cap (Fig. 1A),
160 as such a process can result in circular reasoning (Blaauw, 2012).

161 ¹⁴C dates from core tops on the Kangerlussuaq Trough Mouth Fan date ~15 to 16
162 ¹⁴C ka BP (Andrews et al., 1998; Dunhill, 2005) (Fig. 1B) and evidence from NE and W
163 Greenland indicates that the Greenland Ice Sheet started to retreat shortly after the LGM

(O'Cofaigh et al., 2004; Jennings et al., 2017). There are many dates on either side of Denmark Strait range between 12 and 15 ^{14}C ka BP (Fig. 1B) and bracket the dates for deglaciation. Cosmogenic dates on terrestrial outcrops from either side of Denmark Strait (Brynjolfsson et al., 2015; Dyke et al., 2014; Principato et al., 2006; Sinclair et al., 2016), tephra identifications (Andrews et al., 2008; Jennings et al., 2002, 2014), and ^{14}C dates on raised marine deposits (Norðdahl and Ingolfsson, 2015), indicate that the ice sheets retreated to the present coastlines by the time of the Vedde tephra ca 10.6 ^{14}C ka BP (Lecavalier et al., 2014; Lohne et al., 2013). On Figure 1B we plot the basal ^{14}C dates and whether the cores penetrated basal diamictons (Dmm) (Eyles et al., 1983). We also plot the location of cores JM96-1225 and -1228 where the *N. pachyderma* (s) $\delta^{18}\text{O}$ records (Fig. 2) indicate a rapid decrease of $\sim 1.5\text{‰}$ from LGM values of $\sim 4.5\text{‰}$ to 3.5–3‰ (Hagen, 1999a; Hagen and Hald, 2002). The interpolated ^{14}C ages of the $\delta^{18}\text{O}$ transition in these cores is ~ 13.9 and 15.6 ^{14}C ka BP. South of our study area the $\delta^{18}\text{O}$ transition in core DS97-7P (Kuijpers et al., 2003) (63.523°N, -38.658°W, 1843 m wd) is dated at 13.7 ^{14}C ka BP. These rapid changes in the $\delta^{18}\text{O}$ could be caused by either the input of isotopically light meltwater due to global ice sheet mass loss (Clark et al., 2004; Lambeck et al., 2014; Peltier et al., 2015; Vasskog et al., 2015; Bigg et al., 2011), or by the influx of warm Atlantic Water (Jennings et al., 2006; Knudsen et al., 2003).

The “unknown” element in establishing a chronology for events is the value(s) for the ocean reservoir correction, ΔR , although the stratigraphic sequence of events in our cores will remain unaltered. There is evidence for variations in ΔR during MIS2 in the northern North Atlantic of up to 1000 yr (Butzin et al., 2005; Sarnthein et al., 2015; Thornalley et al., 2015; Stern and Lisiercki, 2013); however, dates associated with the

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187 Vedde and Saksunarvatn tephras, show that no large-scale ΔR correction is required
188 across the Pleistocene/Holocene transition (Andrews et al., 2002a; Jennings et al., 2002,
189 2014). Published calibrations have used a ΔR of 0–400 yr (e.g. Jennings et al., 2006;
190 Knudsen et al., 2003; Patton et al., 2017) but we recognize that two alternative arguments
191 can be used to proceed from ^{14}C dates to calibrated dates, and thus we are forced to work
192 within the constraints of a multiple hypotheses framework (Chamberlin, 1890). Many
193 estimates of larger ΔR 's are based on correlations between the marine and Greenland Ice
194 Sheet isotopic records---this requires a close linkage between a climate record from an
195 elevation of ~3 km and a near-surface marine record. The alternative is that the near
196 surface marine $\delta\delta^{18}\text{O}$ record was responding to freshwater forcing linked to Heinrich
197 events (Bigg et al., 2011) that is dated ca 17.5 cal ka BP (Bigg, 2016; Bigg et al., 2011).
198 A similar degree of change in $\delta^{18}\text{O}$ was noted at the H-1 onset in the Labrador Sea
199 (Andrews et al., 1994) associated with a massive freshwater outburst from Hudson Strait
200 (Hesse and Khodabakhsh, 2016). Support for an earlier, hence small ΔR , also comes
201 from the detailed identification of tephras in core PS2644 (Voelker and Haflidason, 2015)
202 (not shown---just north of B997-326 (Fig. 1; 67.866°N, 21.765°W, 777 m wd); the
203 resulting depth/age model indicates that the abrupt $\delta^{18}\text{O}$ transition in *N. pachyderma* s
204 occurred ~18 cal ka BP.

205 On Figure 1C (and in Suppl. Table 1) we show the calibrated dates with a ΔR of 0
206 (we used Oxcal (Bronk Ramsey, 2008)). This results in deglaciation dates of between 15
207 and 17 cal ka BP and the $\delta^{18}\text{O}$ transition in JM96-1225 at ~16.2 cal ka BP indicating that
208 deglaciation was coeval with or immediately post-Heinrich event 1 *sensus stricto*
209 (Hemming, 2004; Hesse and Khodabakhsh, 2016). However, given the Greenland ice

210 cores $\delta^{18}\text{O}$ records (Vinther et al., 2008; Rasmussen et al., 2014) and regional marine
211 $\delta^{18}\text{O}$ records (Thiagarajan et al., 2014) the abrupt $\delta^{18}\text{O}$ decrease in JM96-1225 (Fig. 2)
212 may date from the Bolling/Allerod (B/A) transition at ~ 14.7 cal ka BP (Rasmussen et al.,
213 2014). If the assumption is made that that the marine $\delta^{18}\text{O}$ transition is coeval with is
214 event, then a larger ΔR is required. In order to derive approximate calibrated radiocarbon
215 dates for this second alternative hypothesis we use a ΔR of 1000 ± 200 for dates ~ 14 ^{14}C
216 ka BP and a ΔR of 0 ± 200 for dates ~ 10.7 ^{14}C ka BP. We linearly interpolated between
217 these two (ΔR decrease of 303 yr ky^{-1}) in order to obtain a calibrated date. This approach
218 has its limitations, but it serves to provide an internally consistent data set (Suppl. Table
219 1 Table 2). These calibrated basal dates (Fig. 1C,) indicate deglaciation between ~ 13 and
220 15 cal ka BP with median age estimates ~ 13.8 cal ka BP compared to ~ 15 cal ka BP with
221 a $\Delta R = 0$. For cores MD99-2264 and JM96-1213 we developed depth/age models
222 (Blaauw and Christen, 2011) for the two alternative ΔR s (Suppl. Material and Suppl. Fig.
223 1).

224 Results: Core sites and records

225 We first summarize the analysis of the regional grain-size data and then present these and
226 other proxy data on a core-by-core basis, although the available data are not identical
227 from core to core. Our reporting of the $\delta^{18}\text{O}$ on foraminifera is not corrected for vital
228 effects or relative sea level (e.g. Smith et al., 2005). For consistency with earlier papers,
229 we will initially refer to the calibrated dates with $\Delta R = 0$, but with an understanding that
230 these estimates are "less than or equal to." Supplemental. Table 1 contains all calibrated
231 estimates. Data embedded in Figures 3,4, 6 and 7 are archived (www.Pangeae.de).

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232 **Grain-size spectra:** Sediment grain-size distributions are important indicator of
233 the mode of sediment erosion, transportation, and deposition (Prins et al., 2002; Weltje
234 and Prins, 2007). Rapid retreat of ice sheets and tidewater ice streams is often marked
235 by massive ice-rafted debris (IRD) signal (Andrews, 2000; Bond et al., 1992), but it can
236 also be signaled by meltwater dominated sedimentation (Hesse and Khodabakhsh, 2016).
237 Grain-size modes (GSMs) (Andrews et al., 2016; Perner et al., 2016) were
238 obtained using k-mean fuzzy clustering (Minasny and McBratney, 2002) on the logratio
239 42 transformed fractions (Aitchison, 1986)---we did not exclude the coarse sand (IRD
240 component) (Hass, 2002). Fuzzy clustering allows samples to be mixtures of the main
241 GSMs. The degree of mixing, termed the “confusion index (CI)”, varies between 0 (a
242 single cluster membership—no confusion) to a maximum of 1. Gradistat data on the
243 samples indicated that many of the samples were polymodal, suggesting transport by
244 more than one process (Curry, 1960). The performance indicators (Fig. 3A) distinguished
245 3 discrete GSMs. The GSM designations are robust as the confusion index is generally
246 low with only 10% having significant contributions ($CI > 0.2$) from more than one GSM
247 (Fig. 3B). The average grain-sizes were computed for each of the 3 GSM’s (Fig. 3C) and
248 indicate that the majority of the samples (GSM #C, $n = 119$) are classified as poorly
249 sorted medium silts. The average GSM #A ($n = 21$, Fig. 3C) is a very poorly sorted, very
250 coarse silt, and the average GSM #B ($n = 22$) is a coarse silt. A bi-plot of mean grain-
251 size versus sorting supports the 3-fold discrimination of the grain-size spectra, although
252 indicating substantial variability in the range of sorting in GSM #C (Fig. 3D) as does a
253 plot of the percentages of medium sand versus medium silt (Fig. 3E). The samples
254 classified as GSM #C are very poorly sorted with coarser grain-size represent fine-

grained glacial marine diamictos. Ice-rafting of medium and coarser sand is evident in most of the GSM #C as shown by the positive skewness and 500 μm size mode (Figs. 3C and 3F). Additional commentary on the GSMs is reserved for the discussion of the individual core data (below Figs 4, 6 and 7). In these figures under "Lithology" we plot the lithofacies and the descriptive name for the sediment, such as "medium silt," for those sections of the cores where we have grain size data.

North Iceland, Reykjafjarðaáall Trough, core B997-326PC: The Húnaflói/Reykjafjarðaáall Trough (HT, Fig. 1A) provided a bathymetric outlet for a major ice stream that flowed northward to the Iceland Sea (Bourgeois, 2000; Hubbard, 2006; Patton et al., 2017, Principato et al., 2016). Core B997-326PC1 (Fig. 4) lies just south of the average AD 1870-1920 April limit of sea ice (Divine and Dick, 2006) (Fig. 1A). Several cores on the northern limit of the trough (B997 -323, -326) penetrated a matrix supported fossiliferous diamicton (Dmm), with dates between 25 and 44 ^{14}C ka BP (Andrews and Helgadottir, 2003). The contact between the basal diamicton and the overlying glacial marine sediments (Fig. 4) is dated 13.8 ± 0.2 ^{14}C ka BP (16.2 ± 0.3 cal ka BP (Suppl. Table 1). A Borrobol-type basaltic tephra was tentatively identified at 140 cm in B997-326PC2 and correlated with similar tephras just above the basal diamictos in B997-326, -332 and -323 (Andrews and Helgadottir, 2003; see www.noaa.gov/paleo). The Vedde tephra may be present at 125 cm (Andrews and Helgadottir, 2003). A date at 18 cm of 9.7 ± 0.3 cal ka BP indicates that the core either did not recover most of the Holocene or has a compressed Holocene record.

Counts of the $> 2\text{mm}$ (IRD) fraction indicate consistent ice-rafting over the interval from ~ 16 to 9.8 cal ka BP (Fig. 4). Nearly all the grain-size spectra between the

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278 base and 125 cm are classified as GSM #C (Fig. 3) with only two samples classified as
279 #A. The qrtz wt% ($\leq 1\%$) is much lower than the present-day input ($\sim 6\%$) (Andrews
280 and Eberl, 2007), probably representing the dominance of local quartz-poor sediments.
281 Calcite and total organic carbon (TOC) (Fig. 4) also remain low until the uppermost
282 centimeters, which date from the last 1 cal ka BP (Andrews and Helgadottir, 2003).
283 Although the sea ice biomarker IP₂₅ was often present, it was generally not above the
284 limit of quantification. However, the co-occurring sea ice biomarker C_{25:2}, is extremely
285 well correlated with IP₂₅ on the North Iceland shelf (NIS) (Massé et al., 2011) and could
286 be readily quantified, although its concentration remained low throughout the record.
287 Foraminifera assemblage data in B997-323, which lies seaward of B997-326 (Fig.
288 1A), were reported in Andrews and Helgadottir (2003). The core recovered a basal
289 diamicton that extended from the base of the core to 80 cm. The planktonic foraminifera
290 numbers reach values of 100/g during deglaciation and are associated with relatively
291 heavy $\delta^{18}\text{O}$ values (3.5–4.0 ‰) (Fig. 10 in Andrews and Helgadottir, 2003) suggesting
292 deglaciation occurred early in the glacial/deglacial transition (e.g. Fig. 2). Immediately
293 above the basal diamicton the benthic foraminifera show high percentages ($\sim 20\%$) of the
294 chilled Atlantic Water species *Cassidulina neoteretis* (Jennings and Helgadottir, 1994),
295 indicating that the deglaciation was associated with the incursion of Atlantic Water.
296 Sparse stable isotope data were obtained on $\delta^{18}\text{O}$ *N. pachyderma* that showed a sharp
297 increase in $\delta^{18}\text{O}$ to 3.8‰ at 60 cm and that Andrews and Helgadottir (2003) tentatively
298 attributed to the Younger Dryas (YD) cold event (12.8 to 11.5 cal ka BP).
299 **NW Iceland, Djúpáll Trough:** Djúpáll is a shallow trough (~ 230 m wd) that extends WNW
300 from Ísafjarðardjúp to Denmark Strait (Fig. 1A). An independent ice cap (Hoppe, 1982;

Norðdahl, 1990) may have existed over NW Iceland and been separated from the main Iceland Ice Sheet during the LGM. A large ridge, interpreted, as a moraine lies seaward of the cores we discuss; the trough contains a large drift deposit (Fig. 5A) (Geirsdóttir et al., 2002; Ólafsdóttir, 2004). Publications on cores from this drift have focused on the last 12 cal ka BP (Andresen et al., 2005; Andrews et al., 2002b, 2013; Ólafsdóttir et al., 2010; Quillmann, 2006; Quillmann et al., 2010). However, several cores in Djúpáll, JM96-1234GC, MD99-2264, and B997-336PC (Andrews et al., 2008; Geirsdóttir et al., 2002), contain ^{14}C dates older than the Vedde tephra (Suppl. Table 1).

Core B997-338PC. This core is situated on the flank of the Djúpáll sediment drift (Fig. 5A). Some of the radiocarbon dates in an initial paper on B997-338PC were incorrect (Andrews et al., 2002b, 2013), probably because of the reworking of foraminifera from the pre-LGM diamicton. The revised chronology is based on two radiocarbon dates on small, articulated shells associated with rhyolitic tephtras (Andrews et al., 2013) (Fig. 6A). The two rhyolitic tephtras, some 100 cm apart (Suppl. Table 2), might indicate Borrobol-type (BT) tephtras (Andrews et al., 2013; Davies et al., 2004; Lind et al., 2016) and thus not in keeping with $\Delta R=0$; however, two rhyolitic tephtras were noted in the NGRIP core and dated ~ 15.2 and 15.5 cal ka BP (Mortensen et al., 2005).

The upper 20 cm of core B997-338PC consists of reworked Vedde ash (Fig. 5A) thus consistent with the ^{14}C date of 11.56 ± 0.17 ka BP from 21 cm (Suppl. Table 1). The core (Fig. 6A) consists of a basal diamicton of compacted overridden glacial marine sediments (dry densities of $\sim 1.5 \text{ g/cm}^3$), overlain by poorly sorted, sandy mud or muddy sand ($\sim 1 \text{ g/cm}^3$) with some coarse IRD (Chesley, 2005); glacial marine deposition started

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324 ~15 cal ka BP. In detail, the GSMs are dominated by #C with an upper interval of #B
325 (Figs. 3 and 6A). The basal diamicton samples are classified as GSM #C, and they
326 comprise the very poorly sorted component (Fig. 3D). Sediment mineralogy is
327 dominated by local basalt sources and the qrtz wt% is very low (<1 %), hence there is
328 little evidence for a contribution by ice-rafting from Achaean/Paleoproterozoic Greenland
329 bedrock. Age estimates ($^{39}\text{Ar}/^{40}\text{Ar}$) on basalt IRD clasts do, however, indicate that some
330 clasts were derived from the E Greenland early Tertiary outcrop (Principato, 2003;
331 Principato et al., 2006). The sea ice biomarkers IP₂₅ and C_{25:2} were generally below the
332 limits of detection, although C_{25:2} could be quantified in a few cases indicating limited
333 occurrence of sea ice.

334 The modest levels of calcite % (Fig. 6A) indicate there was some marine
335 productivity. The foraminifera consist of sparse planktonic numbers and a fauna
336 dominated by species associated with glacial marine conditions, notably *Cassidulina*
337 *reniforme* and *Elphidium excavatum* f. *clavata*, which also are the main species in the
338 modern fauna (Jennings et al., 2004a). The chilled Atlantic Water species *C. neoteretis*
339 occurs in low percentages (Fig. 6A). Sparse stable isotope data were obtained on
340 *Cibicides lobatulus*, a common epifaunal species, that occurs in current dominated
341 environments (Jennings et al., 2004b) and which showed $\delta^{18}\text{O}$ ratios ~2‰ during
342 deglaciation.

343 **Core MD99-2264** was retrieved from the center of the Djúpáll drift (Geirsdóttir et
344 al., 2002; Olafsdottir, 2004) (Fig. 5A). Estimates of modern (last 2-3 decades) wt%
345 quartz, IP₂₅, and $\delta^{18}\text{O}$ *N. pachyderma* (s) were obtained on the accompanying box core
346 MD99-2263 of 2%, 0.004 µg/g, and 2.2‰ respectively (Andrews et al., 2009). The

347 Vedde tephra was identified at 11.8 m depth. Articulated shells in the basal diamicton
348 Unit A, Fig. 6B) have finite ^{14}C ages of ~ 53 ka BP, whereas dates above a rhythmite unit
349 (Unit C) range in age between 15.7 and ~ 12.5 cal ka BP (Figs. 6B). This core has a very
350 high rate of sediment accumulation allowing for unsurpassed detail of events between
351 deglaciation and deposition of the Vedde tephra (Suppl. Fig. 1). Rates of accumulation
352 averaged 200 yr/100 cm or 125 yr/100 cm depending on the chosen ΔR (Suppl. Table 1,
353 Fig. 6B). This, however, does mean that some elements of the record can be diluted
354 especially those we express in percents/g. This core was not included in Patton et al.
355 (2017) paper.

356 During the initial deglaciation coarse IRD deposition occurred (Fig. 6B, Unit B).
357 The sediment between 11 and 15 m (Unit D) varied between sand to clayey silt, with clay
358 %s between 4 and 23%. Calcite wt% was slightly higher in the diamicton than in the
359 overlying sediment but increases slowly and steadily above 20 m core depth. The same
360 pattern is seen in the wt% of quartz with values usually less than 2% below ~ 16 m. The
361 benthic foraminifera are dominated by the glacial marine Arctic species *C. reniforme* and
362 *E. excavate* f. *clavata* ---these two species are also present in large percentages in the
363 basal diamicton. There are, however, low percentages of the chilled Atlantic Water
364 species *C. neoteretis* as well as the boreal species *C. laevigata*, with a notable but short
365 lived peak in percentages ~ 1850 cm (~ 14.5 cal ka BP) suggesting relatively warm
366 conditions (Fig. 6B). Stable $\delta^{18}\text{O}$ ratios from the glacial marine sediments ≤ 26 m on
367 the near-surface planktonic foraminifera *N. pachyderma* (s), and on the benthic
368 foraminifera *C. reniforme* (Olafsdottir, 2004) (Fig. 6B) have some outliers. Median $\delta^{18}\text{O}$
369 ratios of 3.0 ‰ and 3.94‰ respectively are intermediate in their values relative to ratios

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370 before and after the B/A transition in Denmark Strait (Hagen, 1999a; Ólafsdóttir, 2004).

371 The large difference between the benthic and planktonic $\delta^{18}\text{O}$, especially between 17 and

372 18 m indicates the presence of a substantial meltwater cap. No IP_{25} or $\text{C}_{25:2}$ were detected

373 in samples from 1215 and 1260 cm depth (~ 12.8 to 13.1 cal ka BP) and coarse IRD is

374 rare above 20 m (~ 14.3 cal ka BP).

375 Thus, the 4 basal dates from this area of Iceland range in age between $15.3 - 16.2$

376 cal ka BP ($\Delta R = 0$) versus $13.9 - 14.7$ cal ka BP ($\Delta R = \sim 1000$), and present two

377 potentially quite different scenarios in terms of their glaciological interpretation(s).

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379 **East Greenland:** Kangerlussuaq Trough is a deep cross-shelf trough that trends nearly

380 N-S from Kangerlussuaq Fjord to the shelf break south of Denmark Strait (Dowdeswell

381 et al., 2010; Mienert et al., 1992; Stein, 1996). Stein (1996) mapped a series of

382 recessional moraines along the Trough (Fig. 1A).

383 **Core PO175/1-5** is a 3.2 m long core collected from 501 m water (Mienert,

384 1990), thus from a depth generally below the limit for iceberg scouring (Syvitski et al.,

385 2001). A series of short gravity cores were also collected nearby in ~ 300 m of water and

386 record conditions over the last ca. 150 yr, including both sediment mineralogy and IP_{25}

387 abundance (Alonso-Garcia et al., 2013). The seismic record across PO175/1-5 (Stein,

388 1996) sampled Unit L of Stein (Fig. 5B). The radiocarbon dates indicate that the bulk of

389 the sediment was deposited between ~ 17 and 15^{14}C ka BP (Fig. 7A), although the dates

390 show some reversals, probably from reworking of earlier glacial marine sediments

391 (Suppl. Fig. 1). No grain-size data are available.

392 The quartz and calcite wt% are relatively constant throughout (Fig. 7A) but the
393 quartz is less than 50% of the wt% recorded for recent decades in PO176/GKC 9
394 (Alonso-Garcia et al., 2013), whereas the wt% of calcite has a median value of 0.5%
395 versus 3.3 % in PO176/GKC 9 (i.e. modern). The relatively low quartz wt% indicates
396 that the sediment was predominantly derived from erosion of the E. Greenland early
397 Tertiary basalt outcrop (Andrews et al., 2015). The foraminifera data show fluctuating
398 numbers/g for both the benthic and planktonic species. What is especially noticeable,
399 given the link with cooled Atlantic Water (Jennings and Helgadottir, 1994), is the
400 presence of *C. neoteretis*, which peaks between 150 and 60 cm, whereas the glacial
401 marine species of *C. reniforme* and *E. excavatum* occur in relatively constant percentages
402 although higher in the early part of the record. The $\delta^{18}\text{O}$ data on the near-surface
403 planktonic species *N. pachyderma* (s) showed relatively little variation about a median
404 value of 2.8‰.

405 Both sea ice biomarkers (IP₂₅ and C_{25:2}) show similar trends (Fig. 7A), with low
406 concentration ~300–140 cm depth, after which, values showed an increasing, but
407 variable trend between ~140–60 cm. At 40 cm, a reduction in IP₂₅ and C_{25:2}
408 concentrations coincided with a decrease in the quartz data, although there is a generally
409 poor relationship between the biomarkers and quartz wt%. Highest IP₂₅ and C_{25:2}
410 concentrations were observed during the YD (~25 cm), before returning to lower
411 concentrations towards the top of the core.

412 **JM96-1213GC** is a 540 cm long core that was retrieved in 557 m of water in the
413 inner/mid- section of the deep Kangerlussuaq Trough (Hald, 1996)(Fig. 1A and 5C).
414 The lithofacies consist of two intervals of diamicton with high IRD, counts and two units

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415 of fine-grained “wispy” laminated muds (Fig. 7B); the basal diamicton (Dmm) is
416 interpreted as a till (Jennings et al., 2002). Four radiocarbon dates were obtained
417 between the top of the diamicton and the Vedde tephra at ~100 cm (Jennings et al., 2002)
418 (Fig.7B; Suppl. Table 1). The two basal deglaciation dates from cores in the
419 Kangerlussuaq Trough range in age between ~17.6 and 16.2 cal ka BP ($\Delta R = 0$) as
420 opposed to 14.7 to 15.8 cal ka BP with the variable ΔR (Suppl. Table 1 and Suppl. Fig. 1:
421 Fig. 7B).

422 The GSM modes show cohesive sequences that partly mimic the lithofacies. The
423 basal diamicton and another at ~350 cm are classified as GSM #B and (Fig. 3 and 7B),
424 although this mode is not restricted to the Dmm facies, but frequently it coincides with
425 increased IRD input. The YD interval is marked by a marked light $\delta^{18}\text{O}$ peak in the polar
426 species *N. pachyderma* (s) (Jennings et al., 2006) and a switch to GSMs #C and A (Figs.
427 3 & 7B). The qXRD analysis indicated that calcite was close to the detection limit
428 throughout ($< 1\%$) whereas quartz had limited variability around a median wt% of 4.7
429 with a maximum of 7%. The mineralogy is dominated by pyroxene and plagioclase
430 indicating glacial erosion and transport from the nearby early Tertiary basalt outcrop
431 (Andrews et al., 2015).

432 The biomarker data exhibit a generally different trend to that of quartz wt%,
433 although both are lowest below ~420 cm depth (Fig. 7B). Thus, an alternating profile in
434 IP_{25} and $\text{C}_{25:2}$, with lowest values during the YD and a peak ~80 cm (Fig. 7B), are not
435 observed in the quartz profile, which exhibits a general increase above ~420 cm.

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437 **Discussion**

438 We know little about the stability of the regional ice streams prior to the LGM, although
439 inferences have been made based on nearby deep-sea marine records (Andrews et al.,
440 1998; Hagen and Hald, 2002; van Kreveld et al., 2000). The basal deglaciation dates on
441 cores across Denmark Strait suggest a near synchronous retreat of the Iceland and the
442 East Greenland ice sheets across their respective shelves, although the precise timing is
443 ambiguous because of the uncertainty in ΔR . Estimates of the rate of retreat from the E
444 Greenland shelf break to Kangerlussuaq Fjord depend on which ocean reservoir
445 correction is used, but in both N Iceland and E. Greenland the ice sheets were back to the
446 present coastline by ~ 12.2 cal ka BP (e.g. Patton et al., 2017).

447 The geochemistry of the rhyolitic tephra in B997-338 (Andrews et al., 2013)
448 closely matches those of the Borrobol tephra (Suppl. Table 2). On the N Iceland shelf,
449 Gudmundsson et al (2012) identified a "Borrobol-like" tephra in MD99-2272 and
450 attached an age to it of 14.5 cal ka BP; this identification was used in the Xiao et al
451 (2017) paper. The similarity coefficient (Sarna-Wojcicki et al., 1984) between the
452 B997-338 tephra and the geochemistry of the type Borrobol (Lind et al., 2016) suggests
453 the closest match with the geochemistry at 90 cm. If correct, this supports the larger ΔR
454 age estimates. Borrobol-like tephra have not yet been noted from the Kangerlussuaq
455 Trough and in our N Iceland cores the basaltic tephra (Suppl. Table 2) are more difficult
456 to assign to a specific source (Gudmundsson et al., 2012). If the larger ΔR correction is
457 correct, and the evidence points toward that conclusion, then the rate of retreat along the
458 Kangerlussuaq Trough (Fig. 1A) approached ~ 200 m/yr and are comparable with recent
459 changes in Greenland ice streams (Khan et al., 2014).

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460 There is strong evidence that a major change in deep-water circulation at ~15 cal
461 ka BP resulted in abrupt warming at the onset of the Bølling/Allerød (B/A) interstadial
462 (Thiagarajan et al., 2014; Thornalley et al., 2011). This transition is recognized as a
463 “tipping point” (Turney et al., 2016) in the Northern Hemisphere’s climate system
464 (Praetorius and Mix, 2014, 2016) and it is also marked by an abrupt decrease in the
465 $\delta^{18}\text{O}$ of precipitation falling on both the NGRIP site and the nearby Renland Ice Cap
466 (Vinther et al., 2008)(Fig 1A). However, we cannot be absolutely certain that this
467 atmospheric-derived event was coeval with the abrupt decrease in the $\delta^{18}\text{O}$ of *N.*
468 *pachyderma* (s) in Denmark Strait (Fig. 2), which might rather reflect surface ocean
469 response to freshwater inputs (Bigg et al., 2011),
470 or forcing by atmospheric CO_2 variations (Marcott et al., 2014). Glaciological
471 modeling of the retreat histories of the Greenland and Iceland ice sheets have to consider
472 appropriate calibrated age data and for both ice sheets the fall-back position is to use ΔR
473 = 0 (e.g. Tarasov and Peltier, 2002; Patton et al., 2017).
474 Time dependent processes that influence ice sheet stability include changes in
475 relative sea level, insolation, and ocean warming. On Figure 8A we plot the median
476 calibrated ages and their 95% age range for the two alternative ΔR ’s for the deglaciation,
477 and the median dates for the two alternative ΔR corrections (Fig. 8B). Note that the
478 Patton et al. (2017) paper used a ΔR of 24 ± 23 yr, so their modeled retreats match our
479 $\Delta R = 0$ chronology, and the Sinclair et al (2016) paper also used a $\Delta R = 0$. The
480 difference of ~1.5 cal ky between the two possible chronologies is most marked at a
481 critical time that would place initial deglaciation close to 16 versus ~14.5 cal ka BP
482 (Figs. 6B and 7B). Between 20 and 12 cal ka BP global sea level rose by nearly 60 m

(Lambeck et al., 2014) (Fig. 8C) but rates of change varied dramatically (Fig. 8D). Summer insolation at 65°N rose steadily over this period (Fig. 8E). Syvitski et al (1999) and Norðdahl and Ingólfsson (2015) questioned whether the retreat of the Iceland Ice Sheet could have been driven by the increase in sea level associated with deglaciation (Clark et al., 2002)--- rapid increases in sea level occurred at ~17 and ~15 cal ka BP (Fig. 8D). The argument that rapid deglaciation is associated with relative sea level changes, driven by glacial isostasy and eustasy, is not new (e.g. Hughes, 2011; Tanner, 1965) and has been invoked as an Heinrich event mechanism (Andrews, 1998). However, the local relative sea level (RSL) requires knowledge of the rate of glacial isostatic recovery, which away from the coast (i.e. observations) must be modeled (Peltier et al., 2015; Roy and Peltier, 2015). Models of RSL for our area (Fig. 8C) indicate that glacial isostatic recovery would have resulted in a modest decrease of water depth at ice fronts.

Sinclair et al. (2016) suggested that some of the discrepancy between modeling of the post-LGM history of the GIS and field observations, including E Greenland (Lecavalier et al., 2014), might be due to the lack of ocean forcing in the ice sheet models. On Figure 9 we compare our proxies from the deglaciation to ~12 cal ka BP, and note, where available, the present-day values from our surface/core top samples, and on Figure 10 we illustrate our conclusions in the form of a simplified 3-part scenario.

Specific conditions during deglaciation (Figs. 4, 6 and 7) included: 1) significant amounts of coarse > 2 mm IRD during deglaciation indicating a flux of debris-laden icebergs, 2) variable sea ice conditions at the East Greenland sites compared to relatively infrequent drift ice occurrence on the N Iceland shelf (as indicated by low sea ice biomarker and qrtz wt%)---both drift ice proxies are lower than late Holocene values

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506 (Cabedo-Sanz et al., 2016) (Fig. 9C & F), 3) the presence of *C. neoteretis* (Fig. 9E)
507 indicates that ice retreat was associated with the incursion of Atlantic Water, 4) $\delta^{18}\text{O}$
508 values of *N. pachyderma* (s), significantly lighter than during the LGM and a noticeable
509 E to W trend in the *N. pachyderma* (s) $\delta^{18}\text{O}$ signal (Fig. 9D), and 5) significantly higher
510 percentages of calcite off N Iceland.

511 In our cores the evidence for a distinct YD event is not dramatic, although there is
512 an increase in $\delta^{18}\text{O}$ in B997-323 and -322 (Fig. 4) whereas on the E Greenland shelf the
513 YD is marked by a distinct light $\delta^{18}\text{O}$ signal (Jennings et al., 2006). The YD might also
514 marked by erosional intervals in Djúpáll (Fig. 4, 5, 6 and 7) and be indicated by the
515 increase in quartz and k-feldspar and by the large difference in benthic versus planktonic
516 $\delta^{18}\text{O}$ of 2.93‰ in MD99-2264 (Ólafsdóttir 2004), suggestive of a meltwater cap. In this
517 core the grain-size data show that silt and clay % increased between ~12.7–12.2 cal ka
518 BP whereas the sand fraction increased to 70-90% from ~12.2–11.8 cal ka BP. In JM96-
519 1213GC (Fig. 7B) both sea ice biomarkers decrease to non-detectable values during the
520 YD, which suggests either a lack of sea ice or extensive (perennial) landfast sea ice (Figs.
521 9C and 10). In a number of other Arctic/sub-arctic records, the YD is often characterized
522 by significantly elevated IP_{25} abundances (Cabedo-Sanz et al., 2013; Müller et al., 2014;
523 Belt et al., 2015; Méheust et al., 2016); however, ice conditions are somewhat different
524 for the East Greenland sites, insofar as drift ice delivery can be impeded by the presence
525 of near-permanent thick landfast sea ice, such that reduced sea ice biomarker content may
526 be associated with the latter conditions (e.g., Alonso-Garcia et al., 2013). In contrast,
527 higher IP_{25} (and $\text{C}_{25:2}$) abundances during the BA and Preboreal periods (e.g., Fig 7B)
528 potentially reflect reduced landfast ice cover with increased drift ice reaching the cores

529 sites (Alonso-Garcia et al., 2013). Such an explanation cannot be used to account for the
530 low/absent IP₂₅ content for the N Iceland sites, however, which presumably reflects
531 limited sea ice conditions overall (Fig. 10). In any case, despite the strength of the $\delta^{18}\text{O}$
532 YD signal on the Greenland Ice Sheet (Alley, 2000) the YD cold period (GS 1) is not a
533 striking event in our marine cores.

534 Our study and conclusions are based on a broad geographic coverage of sites (Fig.
535 1 and 10) and multi-proxies, but is focused on the area around Denmark Strait. However,
536 a recent biomarker-based study of sea ice conditions, also on the North Icelandic Shelf
537 but further north and east of our core sites, provides further context to our findings.
538 Thus, Xiao et al. (2017) analyzed IP₂₅ and phytoplankton biomarkers in a core (MD99-
539 2272) located ca. 67°N; 18°W, concluding that sea ice conditions were extensive during
540 the Bolling-Allerod, more seasonal during the YD and absent during the early Holocene.
541 Unfortunately, deglacial conditions were not reported from this core. However,
542 differences with our own findings, especially during the Bolling-Allerod and YD, likely
543 reflect the contrasting nature of the sea ice conditions apparent in each study.
544 Specifically, our proxy data almost certainly reflect drift ice conditions (by comparison
545 with modern analogue signatures of biomarkers and qtz content), while the data of Xiao
546 et al. (2017) are interpreted in terms of local biomarker production and thus indicative of
547 a retreating seasonal sea ice margin as observed in many previous IP₂₅-based sea ice
548 reconstructions (e.g., Belt et al., 2015; Berben et al., 2017; Hörner et al., 2016; Müller et
549 al., 2009, 2012; Müller and Stein, 2014; Stein et al., 2017). Common to both studies,
550 however, are generally reduced sea ice conditions during the YD compared to the
551 Bolling-Allerod. Interestingly, the data of Xiao et al. (2017) show a return in IP₂₅ content

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552 during the mid-Holocene – starting ca. 6.5 cal ka BP – a feature that has been reported
553 previously from an additional core from the same study site (JR51-GC35) and another
554 core located slightly further west (MD99-2269; ca. 66.38°N; 20.51°W) (Cabedo-Sanz et
555 al., 2016), both of which were interpreted as reflecting changes to drift ice conditions,
556 especially as IP₂₅ content had previously been shown to be well correlated with
557 observational sea ice records from the region (Massé et al., 2008). Indeed, the presence of
558 IP₂₅ in MD99-2272 during the mid-Holocene (Xiao et al., 2017) is accompanied by
559 increased phytoplankton biomarker content, which contrasts the Bolling-Allerod and YD
560 profiles where high IP₂₅ is accompanied by relatively low phytoplankton biomarker
561 concentrations, a combination consistent with extensive spring sea ice cover (Belt and
562 Müller, 2013; Belt et al., 2015; Berben et al., 2017; Hörner et al., 2016; Müller et al.,
563 2009, 2012; Müller and Stein, 2014; Stein et al., 2017. As such, the IP₂₅ data in MD99-
564 2272 likely reflect a shift from local seasonal sea ice occurrence during the Bolling-
565 Allerod and YD to drift ice in the Holocene. Although confirmation of the former is
566 challenging due to absence of any modern analogue for the study region, the various
567 studies highlight significant localized sensitivity with respect to sea ice conditions post-
568 deglaciation.

569 **Conclusions**

570 Figure 10 presents a visual capsule of ice sheet extent and ocean surface conditions for
571 the LGM, initial deglaciation, and the YD. Specific points are:

- 572 ❖ Uncertainty in the ΔR correction represents a major limitation in establishing
573 detailed regional marine-based correlations.

- 574 ❖ Detailed search for, and geochemical fingerprinting of tephras > 11 and < 15 ^{14}C
575 ka BP would potentially resolve the ΔR uncertainty, especially the identification ,
576 if present, of the Borrobol tephra and ^{14}C dating of associated sediments. Tephra
577 data from B997-338 points toward the use of a large ΔR in calibrating
578 deglaciation dates.
- 579 ❖ The use of the Greenland Ice Sheet's isotopic records as a 1: 1 template for coeval
580 changes in glacier and ocean response potentially ignores the different response
581 time-scales between the atmosphere, oceans, and cryosphere.
- 582 ❖ Deglaciation of the shelves on either side of Denmark Strait was rapid, probably
583 starting ~ 15 (or ~ 17) cal ka BP with ice margins on the outer coasts by Younger
584 Dryas time.
- 585 ❖ Deglaciation was initiated by the incursion of Atlantic Water and aided by a
586 combination of bathymetry and changes in relative sea level.
- 587 ❖ Biomarker, stable isotope, and mineral proxies indicate that differing conditions
588 prevailed on the N Iceland versus E Greenland shelves during deglaciation and
589 the YD with more drift ice on the East Greenland shelf.
- 590 ❖ Further changes in sea ice conditions are evident further east with a general
591 temporal shift from seasonal local sea ice to absent or drift ice conditions during
592 the Holocene.
- 593 ❖ Evidence for a return to glacial conditions during the YD cold event is sparse and
594 complex in our marine archives.

Acknowledgements

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605

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607 this manuscript.

608

609 Tables

610 Table 1: Core locations---latitude, longitude, and water depth. Available proxies for the
611 six cores are also listed. Note that because of space on Fig. 1C not all sites could be
612 plotted.

613

614 Suppl. Table 1: Radiocarbon dates, errors and calibrated ages

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616 Suppl. Table 2: Geochemical data from cores B997-338, --326, and -323 and comparison
617 with the Lipari obsidian standard (Hunt and Hill, 1996)

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619 **Figure captions**

620 Figure 1A: Map of the Denmark Strait area showing the primary cores (solid squares),
621 cores noted in the text and providing information (filled stars; Suppl. Table 1). Dotted
622 blue line = average position of the sea ice edge (15% coverage) April AD 1870-1920
623 (Divine and Dick, 2006). KT = Kangerlussuaq Trough, H = Húnaflói, the Djúpáll Trough
624 is located at core MD99-2264. Surface Atlantic Currents: Irminger Current = IC, North
625 Iceland Irminger Current (NIIC); cold surface Arctic/Polar Currents: East Greenland
626 Current = EC, East Iceland Current = EIC. B) Map of basal uncorrected ¹⁴C dates and
627 whether the core penetrated a basal diamicton (Dmm). Black squares show the two cores
628 with stable isotope records across the glacial/deglacial transition. C) The calibrated ¹⁴C
629 dates with values based on a ΔR of 0 versus the sliding scale ΔR estimates (Table 2).
630 The number in brackets along Kangerlussuaq Fjord (11.8 cal ka BP) is the average
631 cosmogenic date (Dyke et al., 2014).

632 Figure 2: *Neogloboquadrina pachyderma* s δ¹⁸O records from north and south of
633 Denmark Strait (Hagen, 1999) illustrating the transition from glacial to interstadial
634 values, and the differences in the radiocarbon dates that we assume is associated with
635 regional variations in the ocean reservoir correction (see text).

636 Figure 3: Analysis of the grain-size spectra. A) Plot of the 3 performance indicators in
637 FuzMe (Minansy, 2010; Minasny and McBratney, 2002) (MPE= modified partition
638 entropy, FPI = fuzziness performance indicator, S = Compactness and separation validity
639 function) indicating the optimal solution of 4 Grain-size mode (GSM) clusters; B)
640 Probability plot of the Confusion Index (CI) (0 = certainty of cluster membership, 1 =
641 equal mixture of the 3 grain-size modes C) Plot of the average grain-size spectra for the 3

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3 642 GSMS; D) Plot of the mean grain-size (μm) and sorting; E) Plot of the percentages of
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5 643 medium sand versus medium silt; and F) Plot of aspects of the shape of the grain-size
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8 644 spectra, Skewness versus Kurtosis (Folk and Ward, 1957; Blott and Pye, 2001).
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10 645 Figure 4: Downcore plots of data from B997-326 and -323 (Fig. 1) on the N Iceland
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12 646 shelf (see Tables 1 and 2) for available data and radiocarbon dates. Data are on
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15 647 percentage values unless otherwise noted. Stable isotope data ($\delta^{18}\text{O}$ NPs) for *N.*
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17 648 *pachyderma* (s). Red dashed lines show the possible location of basaltic tephras and
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19 649 solid red lines (also Figures 6 and 7) show the position of rhyolitic tephras (T). The
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21 650 calibrated dates from B997-323 are listed as (cal ka 323).
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24 651 Figure 5: A) Seismic (chirper) plot across the Djúpáll Drift (Geirsdottir et al., 2002;
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26 652 Olafsdottir, 2004). The insert shows the bathymetry and location of the cross-section. B)
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28 653 Seismic section along the outer Kangerlussuaq Trough showing the relative location of
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30 654 the sites for BS1191-K18B/JM96-1216 and 175PO/1-5 in unit L (from Stein, 1996). C)
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32 655 Seismic stratigraphy at the site of JM96-1213GC.
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35 656 Figure 6A & B: Downcore plots of data from the Djupal Trough, NW Iceland (Fig. 1) A)
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37 657 B997-338 showing the location of the rhyolitic tephras (red dashed lines) and calibrated
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39 658 dates $\Delta R = 0$. B) MD99-2264. The depth/age plots show the models for a $\Delta R = 0$ in blue
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42 659 and the variable ΔR in red—the Vedde tephra is indicated by the red dashed line. The
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44 660 vertical dashed line shows depth of the ~14.5 cal ka BP for either model. The $\delta^{18}\text{O}$ data
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46 661 are for *C. reniforme* (benthic) (Cr) and *N. pachyderma* (s) (planktonic)(Nps). The plot
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49 662 (B- P ‰) shows the $\delta^{18}\text{O}$ difference between the benthic and planktonic paired
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52 663 measurements.
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664 Figure 7A and B: Downcore plots of cores from the Kangerlussuaq Trough (Fig. 1A,
665 Table 1) A) PO175/1-5, and B) JM96-1213. The depth/age plots show the models for a
666 $\Delta R = 0$ in blue and the variable ΔR in red—the Vedde tephra is indicated by the red
667 dashed line. The vertical dashed line shows depth of the ~14.5 cal ka BP for either
668 model. The $\delta^{18}\text{O}$ data are on the near-surface planktonic foraminifera *N. pachyderma* (s).
669 Figure 8: Global and regional parameters operating in the period between 12 and 17 cal
670 ka BP. A) Plot of median calibrated ages and the 95% confidence interval for sites
671 discussed in this paper (Fig. 1, see Table 2) using $\Delta R = 0$ (red) and $\Delta R \leq 1000$ yr (blue).
672 The vertical solid blue and red lines represent the median ages for the basal dates for the
673 two ΔR options; C) Eustatic sea level (Lambeck et al., 2014) and estimates of relative sea
674 level change (RSL) at specific latitudes and longitudes (Peltier, person. commun. 2016).
675 D) Rate of change of global sea level (C); and E) July Insolation at 65°N (Berger and
676 Loutre, 1991);
677 Figure 9: Box plots of surface & near-surface environment proxies across Denmark Strait
678 (east to west) between 12 to ~15 cal ka BP--- —numbers give the wt% of modern samples
679 showing: A) Weight % quartz; B) Weight % calcite; and C) IP25; D) $\delta^{18}\text{O}$ for the near-
680 surface polar planktonic foraminifera *Neogloboquadrina pachyderma*; E) Map of the %
681 of *C. neoteretis* for the deglacial interval > 12 cal ka BP; F) Diene II.
682 Figure 10: Simplified scenarios of conditions in Denmark Strait at the end of MIS2 and
683 during the initial period of deglaciation. Red triangles represent icebergs and the white
684 patches represent sea ice. Black dots refer to the core locations as listed in Table 1. Red
685 arrows represent warm surface (Atlantic Water) currents and the dark blue represents the

686 export of cold Polar Water. A) Ice Sheet extent during the LGM; B) conditions during
687 deglaciation, and C) Conditions at the YD interval.
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Figure 1A: Map of the Denmark Strait area showing the primary cores (solid squares), cores noted in the text and providing information (filled stars; Suppl. Table 1). Dotted blue line = average position of the sea ice edge (15% coverage) April AD 1870–1920 (Divine and Dick, 2006). KT = Kangerlussuaq Trough, H = Húnaflói, the Djúpáll Trough is located at core MD99-2264. Surface Atlantic Currents: Irminger Current = IC, North Iceland Irminger Current (NIIC); cold surface Arctic/Polar Currents: East Greenland Current = EC, East Iceland Current = EIC. B) Map of basal uncorrected 14C dates and whether the core penetrated a basal diamicton (Dmm). Black squares show the two cores with stable isotope records across the glacial/deglacial transition. C) The calibrated 14C dates with values based on a ΔR of 0 versus the sliding scale ΔR estimates (Table 2). The number in brackets along Kangerlussuaq Fjord (11.8 cal ka BP) is the average cosmogenic date (Dyke et al., 2014).

279x361mm (300 x 300 DPI)

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Figure 2: *Neogloboquadrina pachyderma s* $\delta^{18}\text{O}$ records from north and south of Denmark Strait (Hagen, 1999) illustrating the transition from glacial to interstadial values, and the differences in the radiocarbon dates that we assume is associated with regional variations in the ocean reservoir correction (see text).

258x329mm (300 x 300 DPI)

Figure 3: Analysis of the grain-size spectra (n = 191). A) Plot of the 3 performance indicators in FuzMe (Minansy, 2010; Minansy and McBratney, 2002) (MPE= modified partition entropy, FPI = fuzziness performance indicator, S = Compactness and separation validity function) indicating the optimal solution of 4 Grain-size mode (GSM) clusters; B) Probability plot of the Confusion Index (CI) (0 = certainty of cluster membership, 1 = equal mixture of the 3 grain-size modes C) Plot of the average grain-size spectra for the 3 GSMs; D) Plot of the mean grain-size (μm) and sorting; E) Plot of the percentages of medium sand versus medium silt; and F) Plot of aspects of the shape of the grain-size spectra, Skewness versus Kurtosis (Folk and Ward, 1957; Blott and Pye, 2001).

226x214mm (300 x 300 DPI)

Fig. 4

269x352mm (300 x 300 DPI)

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A

B

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Fig. 5

279x361mm (300 x 300 DPI)

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Fig. 6

279x361mm (300 x 300 DPI)

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Fig. 7

279x361mm (300 x 300 DPI)

Figure 8: Global and regional parameters operating in the period between 12 and 17 cal ka BP. A) Plot of median calibrated ages and the 95% confidence interval for sites discussed in this paper (Fig. 1, see Table 2) using $\Delta R = 0$ (red) and $\Delta R \leq 1000$ yr (blue). The vertical solid blue and red lines represent the median ages for the basal dates for the two ΔR options; C) Eustatic sea level (Lambeck et al., 2014) and estimates of relative sea level change (RSL) at specific latitudes and longitudes (Peltier, person. commun. 2016). D) Rate of change of global sea level (C); and E) July Insolation at 65°N (Berger and Loutre, 1991);

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Figure 9: Box plots of surface & near-surface environment proxies across Denmark Strait (east to west) between 12 to ~15 cal ka BP--- -numbers give the wt% of modern samples showing: A) Weight % quartz; B) Weight % calcite; and C) IP25; D) $\delta^{18}O$ for the near-surface polar planktonic foraminifera *Neogloboquadrina pachyderma*; E) Map of the % of *C. neoteretis* for the deglacial interval > 12 cal ka BP; F) Diene II.

178x207mm (300 x 300 DPI)

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Figure 10: Simplified scenarios of conditions in Denmark Strait at the end of MIS2 and during the initial period of deglaciation. Red triangles represent icebergs and the white patches represent sea ice. Black dots refer to the core locations as listed in Table 1. Red arrows represent warm surface (Atlantic Water) currents and the dark blue represents the export of cold Polar Water. A) Ice Sheet extent during the LGM; B) conditions during deglaciation, and C) Conditions at the YD interval.

254x190mm (96 x 96 DPI)

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Supplemental Material

Methods

Coarse ice-rafted sediment >2 mm: The number of clasts > 2 mm were counted in 2-cm vertical increments on X-radiographs of the split cores (Andrews et al., 1997; Geirsdottir et al., 2002). These counts are used as the most definitive evidence of ice-rafting debris (IRD) associated with deposition from icebergs (Andrews, 2000)(Andrews, 2000).

Grain-size: Lithofacies were based on visual inspection and examination of X-radiographs on split cores. Grain-size data were obtained on 162 samples from cores B997-322, -323, -326, -328 and JM96-1213 with boundaries between 2000 and 2 µm. The samples were processed on the Malvern long-bed laser system after treatment with hydrogen peroxide to remove any organics. Samples were not decalcified as detrital carbonate can be an important signal of the export of sediment from or from the extensive limestone outcrops in N Greenland and the Canadian Arctic Archipelago. Furthermore, decalcification of samples from MD99-2269 (Andrews et al, 2003) resulted in little change in grain-size properties. Grain-size statistics were obtained using Gradistat (Blott and Pye, 2001)(Blott and Pye, 2001). Rather than just report variations in parameters, such as mean grain-size, we sought to more rigorously characterize the grain-size distributions (Weltje and Prins, 2003; Weltje and Prins, 2007).(Weltje and Prins, 2003; Weltje and Prins, 2007)

X-ray diffraction of < 2 mm sediment fraction: Quantitative X-ray diffraction (qXRD) analysis of the < 2 mm sediment fraction was undertaken using the method outlined by Eberl (Eberl, 2003)(Eberl, 2003) in which 10% by weight of ZnO is added to the sample, milled for 5 minutes in a McCrone mill, dried, and then loaded into a Siemens D5000 X-ray diffractometer. Each sample runs for 100 minutes with scanning taking place in 0.02 2-theta steps between 5 and 65° 2-theta, and measured intensities recorded every 2 secs. The data are then downloaded and weight% of minerals estimated using the Excel macro program Rockjock v6 (Eberl, 2003). qXRD data from the area have been reported previously (Andrews, 2008; Andrews and Eberl, 2007). The QXRD data have been used in a sediment unmixing model (Andrews and Eberl, 2012) to better understand the varying contributions from possible bedrock source areas (Andrews et al., 2015).

Light Stable Isotopes: $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data have been obtained on both benthic and planktonic foraminifera for our sites, but at varying levels of resolution. Details on the processing of the samples are given in Olafsdottir (Olafsdottir, 2004). Data were obtained on the dominant polar planktonic foraminifera *Neogloboquadrina pachyderma* s and on several species of benthic foraminifera. We compare our data with $\delta^{18}\text{O}$ values of *Neogloboquadrina pachyderma* s from cores in the area of Denmark Strait, namely JM96-1225, -1228 (Hagen, 1999b; Hagen and Hald, 2002) and MD99-2323 (Dunhill, 2005) (Fig. 1). There is also a considerable body of information on both modern (surface) and Holocene $\delta^{18}\text{O}$ ratios on both planktonic and benthic foraminifera from both shelves (Castaneda et al., 2004; Jennings et al., 2011; Knudsen et al., 2004; Kristjansdottir et al., 2016 ; Smith et al., 2005).

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Foraminifera assemblages: Details of the sample processing for foraminifera are given in Olafsdottir (Olafsdottir, 2004) and Jennings et al (Jennings et al., 2004). Interpretations are based on the analysis of the occurrence of species in seafloor sediments collected in the 1990s (Jennings and Helgadottir, 1994; Jennings et al., 2004; Rytter et al., 2002).

Biomarkers: Biomarker analyses (IP₂₅ and C_{25:2}) were performed on 107 samples from cores B997-326 (n = 21), -338 (n = 8), JM96-1213 (n = 43) and 175PO/1-5 (n = 35), using methods described previously (Belt et al., 2012; Belt et al., 2013). Briefly, two internal standards, 7-hexylnonadecane and 9-octylheptadec-8-ene (each 0.10 µg) were added to freeze-dried sediment samples (ca. 3 g) to permit quantification of IP₂₅ and C_{25:2}. Samples were then extracted using dichloromethane/methanol (3 x 3 mL; 2:1 v/v; ultrasonication; 15 min). Dried total organic extracts were then partially purified using column chromatography (silica; hexane; 6 mL) and further fractionated into saturated and unsaturated components using silver ion solid-phase extraction (SPE) material (100 mg; Supelco discovery® Ag-Ion). Saturated and unsaturated (including IP₂₅ and C_{25:2}) hydrocarbons were with hexane (1 mL) and acetone (2 mL), respectively. Analysis of purified fractions was carried out using gas chromatography–mass spectrometry (GC–MS) as described previously (Belt et al., 2012). Mass spectrometric analysis was mainly carried out in single-ion monitoring (SIM) mode. IP₂₅ and C_{25:2} were identified on the basis of their characteristic GC retention indices and mass spectra obtained from standards, while quantification was achieved by comparison of mass spectral responses of selected ions (IP₂₅: *m/z* 350; C_{25:2}: *m/z* 348) with that of the internal standard (*m/z* 350) and normalized according to relative response factors and sediment masses (Belt et al.,

2012). Analytical reproducibility was monitored using standard sediment with a known concentration of IP₂₅ (8 %, n = 8).

Depth/age radiocarbon models:

In sites MD99-2264 and JM96-1213 there are sufficient dates over a reasonable depth span to construct depth/age curves using the two ΔR options (Suppl. Table 1). We used the flexible Bayesian program (Bacon) (Blaauw and Christen, 2011) to reconstruct depth/age plots (Suppl. Fig. 1)--- in Figures 6B and 7B these are simplified and only the median ages are plotted. Accumulation rates were high at both sites and both models tended to overestimate the age of sediment at the depth of the Vedde tephra (Suppl. Fig. 1) indicating either a change in the prior or a need to change ΔR . In both cores the rates of accumulation are essentially constant. The variations of age versus depth in PO175/1-5 (Suppl. Table 1) were such that we have not used an age model in Fig. 7A but the results of the two alternative ΔR s are plotted on Suppl. Fig. 1.

Caption Suppl. Fig. 1: Outputs from the Bayesian calibration program "Bacon" (Blaauw and Christen, 2011, 2016) for MD99-2264, JM96-1213, and PO175/105 showing the results with the two different ocean reservoir corrections (see text). The single "best" model is the faint red dashed line. Grey stippled areas are the 95% confidence limits

Suppl. Table 1:

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92 This table lists all the radiocarbon dates used in this paper and shows the number of
93 the date listed on Fig. 8 (col A), material dated (B), original data and error (E & F).
94 Cols G to K give the calibrated 95% age range, median and sigma error with a $\Delta R =$
95 0, and cols L to R give the estimated ΔR , error, 95% age range, median date and
96 error for calibrations with the variable ΔR (see text).

97
98 **Suppl. Table 2:**
99

100 Geochemical analyses for tephras noted in this paper. Data from the University of
101 Colorado's microprobe analysis of the Lipari obsidian standard (Hunt and Hill,
102 1996)is listed first followed by the tephras in B997-338 and -323 and -326
103 (Andrews and Helgadottir, 2003; Andrews et al., 2013).

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Suppl. Table 1

Core	Number Fig. 8	Depth	P or B	14C date	error
ICELAND					
HM107-05	1	381	P	14100	140
	2	394	B	13690	100
	3	394	SHELL	13980	90
B997-326PC		18	B	9040	110
		111	B	9580	100
		125	Vedde/Saks		
	4	182	P&B	13835	215
		198	B	23570	340
B997-323		48	Vedde?		
	5	59	B	13100	130
	6	71		13400	190
		90	B	>25,00	
B997-338		21	B	11560	170
			tephra		
	7	98.5		13020	220
	8	193	shell	13507	78
		198	tephra		
	9	210	shell	13235	62
		319.5	B	31900	1700
		412	B	34600	640
MD99-2264			Vedde		
	10	1169	shell	11170	90
	11	1339	shell	12080	800
	12	1705	shell	12714	86
	13	2183	shell	13130	91
	14	2749	shell	13469	89
		37010	shell	53436	800
JM96-1234		349	B	14030	70
B997-336		501	shell	13680	70
HU90030-006LCF	15	1175	B	12690	195
	16	1235	B	12810	205
	17	CC	B	13105	85
MD99-2256	18	1184	shell	11840	90
MD99-2259	19	2048	B	12790	120
E. GREENLAND					
HU93030-007PC	20	2261	shell	13790	80
	21	7.5	P--Nps	15270	120

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4	MD99-2260	22	11 P--Nps	16230	150
5	JM96-1216/2		55 BF	8180	80
6			95 BF	9580	100
7					
8		23	210 shell	10625	90
9		24	246 shell	12040	80
10		25	BF	14450	150
11	BS1191-K18B		4.5	1680	50
12			37	5215	75
13			62	9292	80
14			77	9240	90
15					
16					
17		26	97 B	12325	80
18		27 >150	N p s	12470	205
19		28	N p s	12085	115
20		29	B	12865	305
21					
22					
23					
24	PO175/1-5		29.5 P & B	13100	110
25		30	59.5 P & B	11995	145
26			95 B	13300	145
27					
28		31	125 shell	13024	120
29		32	194 B	14465	200
30		33	311 B	14845	190
31	JM96-1213		100		
32				Vedde	
33			199 Triboculina	14680	170
34		34	320.5 Bryozoa	12630	80
35		35	389 B	13690	230
36		36	439 N labradorica	13356	108
37		37	473 Ex. Clavata	13830	270
38					
39					
40					
41	MD99-2317	38	1851 Ex. Clavata	11567	88
42		39	1982 bryozoa	11950	110
43					
44	MD99-2322		2436 shell	10442	82
45	JM96-1214	40	562 B	11380	80
46		41	469 Bryozoa	12700	110
47	JM96-1215	42	495 P & B	12260	100
48		43 >587	B	12900	50
49	JM96-1225		47.5 P	12455	65
50			52.5 18O transition		
51			57 P	15380	130
52					
53	JM96-1228		213 P	15140	185
54			220 18O transition		
55			231 P	16270	160
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B = benthic forams

P = planktonic forams

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5	deltaR = 0 fro to	range	Median	error	linear est deltaR	error	
6							
7	17000	16145	855	16544	222	1000	200
8	16265	15684	581	15974	150	906	200
9	16680	16070	610	16361	151	994	200
10	10362	9236	1126	9786	280		
11	11067	9878	1189	10458	290		
12			0				
13	16891	15557	1334	16176	324		
14	27954	26660	1294	27420	307	3900	
15			0				
16	15581	14438	1143	15072	271		
17	16154	15014	1140	15553	288	818	
18			0				
19	13349	12698	651	13032	170	261	100
20			0				
21	15636	14108	1528	14877	403	703	200
22	15993	15384	609	15715	151	851	200
23			0				
24	15568	15114	454	15296	117	768	200
25	40848	32665	8183	36092	2073		
26	40226	36889	3337	38698	825		
27			0				
28	12854	12538	316	12673	78	142	200
29	16346	11801	4545	13860	1122	418	200
30	14778	13997	781	14317	209	610	200
31	15479	14750	729	15152	172	736	200
32	15952	15316	636	15650	163	839	200
33							
34	16704	16182	522	16423	131	1000	200
35	16192	15736	456	15962	116	900	200
36							
37	15091	13815	1276	14379	345	603	
38	15231	13918	1313	14558	354	639	
39	15388	14724	664	15114	165	730	200
40	13481	13131	350	13309	88	346	
41	15037	14068	969	14494	261	633	
42							
43	16340	15834	506	16105	126	936	
44	18394	17789	605	18087	151	1384	200
45							
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3	19496	18791	705	19106	183	1675	200
4	8951	8481	470	8689	122		
5	10683	10215	468	10447	125		
6	12382	11464	918	11944	212	-23	
7	13697	13316	381	13488	97	406	200
8			0			1136	200
9	1337	1128	209	1238	50		100
10	5755	5392	363	5576	86		100
11	10318	9850	468	10109	114		100
12	10239	9732	507	100031	131		100
13	14010	13568	442	13795	108	492	200
14	14792	13464	1328	14003	335	536	200
15	13806	13311	495	13545	127	420	200
16	15163	14094	1069	14623	296	656	200
17			0				
18	15537	14563	974	15089	226	727	200
19	13785	13185	600	13460	152	392	
20	15906	14975	931	15420	231	788	200
21	15709	14826	883	15267	215	704	200
22	17633	16458	1175	17078	299	1141	200
23	18043	17074	969	17589	247	1256	200
24			0				
25			0			1206	200
26	14554	13885	669	14139	163	585	200
27	16642	15268	1374	15965	344	906	200
28	15829	15163	666	15489	173	805	200
29	16985	15355	1630	16175	406	948	200
30							
31	13250	12815	435	13042	109	263	200
32	13667	13196	471	13407	115	379	200
33	11915	11250	665	11571	179	-78	200
34	13055	12671	384	12845	98	206	200
35	14862	13938	924	14320	242	606	200
36	13961	13470	491	13718	125	473	200
37	15052	14313	739	14720	196	667	200
38	14107	13770	337	13937	86	532	200
39			0	16180			
40	18529	17900	629	18205	160	1420	200
41	18387	17494	893	17936	227	871	200
42			0				
43	19547	18810	737	19153	194	1690	200
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from	to	Range	Median	error	Iceland
15741	14165	1576	15021	417	
15230	13844	1386	14523	373	
15569	14068	1501	14813	393	
10362	9236	1126	9786	280	
11067	9878	1189	10458	290	
		0		100	
161230	13561	147669	14797	667	
		0			
		0	12100		
14756	13344	1412	13918	354	
		0			
		0			
13313	12611	702	12949	186	
		0			
14875	13195	1680	13850	420	
15097	13727	1370	14327	365	
		0			14
14751	13437	1314	13982	328	average
		0			median
		0			
		0	12100	100	
13026	11835	1191	12502	298	
15815	11150	4665	13314	1172	
14034	13134	900	13568	228	
14655	13365	1290	13891	317	
15091	13695	1396	14298	370	
15610	14119	1491	14888	393	
15196	13865	1331	14513	359	
14210	12915	1295	13575	320	E Greenland
14511	12996	1515	13667	354	
14605	13348	1257	13870	306	
13375	12596	779	12977	206	
14143	13146	997	13633	257	
15295	13924	1371	14615	367	
16984	15564	1420	16246	350	

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4	17862	16460	1402	17190	354	
5			0			
6			0			
7			0			
8						
9	13485	12675	810	13098	209	
10	16156	14405	1751	15405	428	
11	1471	981	490	1229	119	
12	5859	5305	554	5576	143	
13	10434	9687	747	10087	184	
14	10366	9608	758	10013	188	
15	13749	12830	919	13297	226	
16	14012	12794	1218	13413	309	
17	13559	12670	889	13129	226	
18	14956	12946	2010	13795	495	
19			0			
20	14637	13343	1294	13871	317	average
21			0			
22	14991	13466	1525	14095	397	median
23	14647	13269	1378	13818	336	
24	16221	14303	1918	15410	478	
25	16685	14971	1714	15810	421	
26			0	12100	100	
27	16447	14742	1705	15642	411	
28	13986	13090	896	13519	224	
29	15531	13675	1856	14550	474	
30	14993	13535	1458	14144	377	
31	15755	13736	2019	14688	523	
32						
33	13164	12094	1070	12701	263	
34	13453	12629	824	13043	216	
35			0			
36	13130	12060	1070	12667	267	
37	14065	13112	953	13570	241	
38	13708	12772	936	13250	231	
39	14150	13271	879	13701	231	
40	13836	12936	900	13391	221	
41			0			
42	17080	15670	1410	16356	355	
43	17556	16040	1516	16794	390	
44			0			
45	17897	16478	1419	17223	360	
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from	to	median cal	
14605	13348	13870	16544
15741	14165	15021	15974
15230	13844	14523	16361
16230	13561	14797	15072
14756	13344	13918	15553
13313	12611	12949	14877
14875	13195	13850	15715
15097	13727	14327	15296
13026	11835	12502	14317
15815	11150	13314	15152
14034	13134	13568	15650
14655	13365	13891	14379
14210	12915	13575	15114
14511	12996	13667	13309
13375	12596	12977	14494
14143	13146	13633	16105
15295	13924	14615	
14644	13094	13820	
14706	13171	13759	15224

from	to	median cal	
13485	12675	13098	
16156	14405	15405	
13749	12830	13297	
14012	12794	13413	14623
14637	13343	13871	15420

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14991	13466	14095	15267
14647	13269	13818	17078
16221	14303	15410	17589
16685	14971	15810	14139
13986	13090	13519	15965
15531	13675	14550	15489
14993	13535	14144	16175
15755	13736	14688	14320
13164	12094	12701	13718
13453	12629	13043	14720
13130	12060	12667	13937
14065	13112	13570	
13708	12772	13250	15264.6154
14150	13271	13701	13.9-14.2
14554	13265	13897	
14150	13269	13701	

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Suppl. Table 2

		SiO2	TiO2	2O3	FeO*	MnO
Lipari obsidian standard	Mean	74.12	0.07	12.98	1.67	0.07
CU Lab n = 10	sd	0.86	0.03	0.41	0.21	0.02
B99-338						
90 cm	Mean	78.8	0.2	13.1	3.3	0.1
n = 3	SD	0.5	0.0	0.3	0.1	0.0
99-101 cm	Mean	74.0	0.3	14.0	3.8	0.1
n = 9	SD	4.8	0.1	1.9	0.7	0.0
190 cm	Mean	74.0	0.3	14.0	3.8	0.1
n = 9	SD	4.8	0.1	1.9	0.7	0.0
192-194 cm	Mean	77.1	0.3	11.7	5.7	0.1
n = 9	SD	1.5	0.2	1.4	2.4	0.1
Borrobol type						
n = 10	Lind et al 2016--Table 2 Mean	76.67	0.13	12.71	1.56	0.04
	standard deviation	0.35	0.01	0.25	0.17	0.01
	Similaritycoefficient					
	Borrobol v 90 cm	0.997				
	Borrobol v 100 cm	0.742				
	Borrobol v 90 cm	0.742				
	Borrobol v 193 cm	0.830				
Basalt tephras						
B997-326	156 cm	49.18	0.53	14.18	8.42	0.17
B997-323	69 cm mean	50.04	2.20	13.16	12.96	0.23
	sd	0.64	0.47	0.56	0.89	0.02
	94 cm mean	50.68	3.35	13.28	13.69	0.24
	sd	1.56	1.01	1.26	1.33	0.03

MgO	CaO	Na2O	k2O	P2O3	Total
0.04	0.77	3.74	5.03	0.01	98.48
0.01	0.11	0.16	0.25	0.02	

0.0	0.5	1.4	3.0	0.0	100.2
0.0	0.0	0.5	0.3	0.0	
0.1	1.1	2.4	3.4	0.0	99.2
0.2	0.9	1.4	0.8	0.0	
0.1	1.1	2.4	3.4	0.0	99.2
0.2	0.9	1.4	0.8	0.0	
0.0	0.4	1.4	3.1	0.0	99.8
0.0	0.0	0.5	0.6	0.0	
0.08	0.75	4.03	4.01	0.01	
0.02	0.06	0.47	0.12	0.01	
10.28	15.76	1.39	0.04	0.05	100.00
6.95	10.96	2.47	0.21	0.28	99.47
0.60	0.44	0.21	0.06	0.29	0.83
5.38	9.40	2.86	0.57	0.38	99.83
1.54	1.54	0.56	0.41	0.16	1.25

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279x361mm (300 x 300 DPI)

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330x408mm (300 x 300 DPI)