Asynchronous instability of the North American-Arctic and Greenland ice sheets during the last deglaciation

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Abstract

The chronology of deglacial meltwater pulses from the Laurentide Ice Sheet is well documented. However, the deglacial history of the North American-Arctic (north-eastern Laurentide and Innuitian) and western Greenland ice sheets draining into the Labrador Sea via Baffin Bay is less well constrained. Here we present new high-resolution, radiocarbon-dated records from the central Baffin Bay spanning ~17 to 10 kyr BP and documenting the full deglacial history of Baffin Bay. Sedimentological and geochemical data confirm the presence of two periods of enhanced detrital carbonate delivery, termed Baffin Bay Detrital Carbonate Events (BBDCs). These events are dated to ~14.2–13.7 kyr BP and ~12.7–11 kyr BP. They are synchronous across Baffin Bay and their mineralogical signature indicates a common source of detrital carbonate from northern Baffin Bay. The first event, BBDC 1, postdates Heinrich Event 1 and the second event BBDC 0 predates the recently revised timing of Heinrich Event 0. The onset of the BBDC events appears not to be systematically linked to Greenland temperature change as they occur during both interstadial and stadial periods. This indicates that deglaciation of North American-Arctic and western Greenland ice sheets with the associated iceberg and meltwater discharge were decoupled from the dominant North Atlantic climate mode, where iceberg discharge events from the Laurentide Ice Sheet occurred during stadial periods.

1. Introduction

The transition from the Last Glacial Maximum to the Holocene was characterised by rapid changes in temperature, ice sheet volume and ocean circulation (Carlson and Clark, 2012; Clark et al., 2012; Shakun and Carlson, 2010). In the Arctic and sub-Arctic regions, this transition is associated with two distinct events of ice sheet instability, known as Heinrich Events 1 and 0 (see a review in Clark et al., 2012). Heinrich Events, massive discharges of icebergs mostly from surges of the Laurentide Ice Sheet (Heinrich, 1988), are well documented in Labrador Sea (Andrews et al., 2012; Pearce et al., 2015; Stoner et al., 1996) and North Atlantic (Bond et al., 1997; Channell et al., 2012; Hemming, 2004) sedimentary records and their timing is well constrained. Heinrich Events are associated with cold stadial conditions in the North Atlantic, but their causal relationship to stadial coolings is unclear (Bond et al., 1993; Broecker, 1994; Marcott et al., 2011). Notwithstanding their exact relationship with

the regional climate evolution, these events were associated with large meltwater injections into the North Atlantic and as such likely to have had an impact on deep-water formation, especially in the Labrador Sea, thus modulating the global climate evolution of the last deglaciation (McManus et al., 2004; Thornalley et al., 2011). Most climate hypotheses and paleoclimate models assume that deglacial meltwater discharge into the North Atlantic was associated with Heinrich Events sourced from the Hudson Bay. However, there exists a second potential meltwater route, from Baffin Bay (e.g. Simon et al., 2014). Compared to the well-known chronology of late glacial meltwater discharges from the Labrador Dome of the Laurentide Ice Sheet since the Last Glacial Maximum (LGM), meltwater discharges from ice sheets that drain into Baffin Bay are less well documented and still loosely dated: a gap we intend to fill here.

1.1 Regional setting and previous work

During the LGM, Baffin Bay was flanked by three major ice sheets that flowed into it: the Greenland Ice Sheet (GIS), north-eastern sector of the Laurentide Ice Sheet (LIS) and the south-eastern sector of the Innuitian Ice Sheet (IIS) (Dyke et al., 2002; England et al., 2006). We will term these as the North American-Arctic and western Greenland ice sheets. These ice sheets expanded into the fjord systems surrounding Baffin Bay and filled marine channels such as the Nares Strait/Smith Sound, and Lancaster and Jones Sound (Figure 1) (Dyke et al., 2002; Li et al., 2011) as well as resulting in the spreading of an ice shelf across the northern Baffin Bay (Marcott et al., 2011). The GIS advanced through Disko Bugt and the Uummannaq Trough as far as the shelf edge on the western Greenland margin (Ó Cofaigh et al., 2013; Ó Cofaigh et al., 2013a).

Baffin Bay has a counter-clockwise oceanic circulation (Tang et al., 2004). The colder East Greenland Current and warmer Irminger Current join to form the West Greenland Current (WGC) that enters Baffin Bay via the Davis Strait and travels north along the west coast of Greenland. Paleoceanographic records from the Uummannaq region of the west Greenland shelf suggest the establishment of 'warm' WGC prior to 14 kyr BP (Sheldon et al., 2016). At the head of Baffin Bay the WGC recirculates south, incorporating Arctic waters and forming the Baffin Island Current (BIC). The BIC travels south along Baffin Island, with some recirculation east towards Greenland, while the rest flows into the Labrador Sea via the Davis Strait, forming part of the Labrador Current (Figure 1).

As well as the direct sources of sediment from the adjacent landmasses and shelves, this ocean current configuration provided a southerly path for iceberg-rafting and meltwater discharge from the North American-Arctic ice sheets (England et al., 2006; Simon et al., 2014), as well as a northerly path from the western GIS and its previously glaciated shelf (Jennings et al., 2014; Ó Cofaigh et al., 2013). In the northern region of Baffin Bay, many of the major paleo-ice streams draining the IIS and north-eastern LIS, such as the Nares Strait/Smith Sound and Lancaster Sound (Figure 1) are flanked by large areas of Palaeozoic limestones and dolostones (Hiscott et al., 1989). This is in contrast to the older Archean Nagssugtoqidian Mobile Belt (NMB) of the central west Greenland coast. Discrete discharge events originating from northern Baffin Bay are hence recorded in marine sedimentary records as layers rich in detrital carbonate, generally ice-rafted southward along the path of the BIC (Andrews et al., 2011). In contrast with the calcite-rich mineralogical signature of detrital carbonate layers originating from the Hudson Strait, ice-rafted detrital carbonate layers found in Baffin Bay are characterised by a calcite/dolomite ratio $\sim 0.25:1$ (Andrews et al., 2014, 1998, 1995; Simon et al., 2014).

Two BBDC events have been previously identified during the last deglaciation (Aksu and Piper, 1987; Andrews et al., 1998; Hiscott et al., 1989; Simon et al., 2014, 2012). Magnetostratigraphic dating of these records has been recently used to suggest that these events were not in phase with the North Atlantic Heinrich Events (Simon et al., 2014, 2012) but much uncertainty exists in such a paleomagnetic-based chronology, demonstrated by its recent revision (Simon et al., 2016). This uncertainty derives from the ambiguity in tuning noisy curves to a target curve that, in itself, includes dating uncertainty. If periods of iceberg and meltwater discharge from the ice sheets draining into Baffin Bay (i.e. BBDCs) were asynchronous Heinrich Events, this could have implications for the history of meltwater delivery into the Labrador Sea and North Atlantic. Obviously, a robust chronology would allow here for a better evaluation of the relationship between iceberg discharge and deglacial climate change (Bond et al., 1993; Broecker, 1994).

Deep sea sediments in Baffin Bay are affected by poor preservation of biogenic carbonate, particularly during the Holocene (Aksu, 1983), which has so far hindered the construction of a radiocarbon chronology for the deglaciation and thus a robust assessment of the melting history of the

surrounding ice sheets. Here we present two new deglacial records from Baffin Bay (Figure 1), from sites where a better preservation of benthic and planktonic foraminifera is observed. Better preservation is likely due here to the cored sediment depths near the carbonate compensation depth (CCD), ranging from 600–900 m in Baffin Bay (Aksu, 1983), as compared to deeper cores (> 2000 m water depth, e.g. Simon et al., 2012). Combined with the use of a new preparation technique for radiocarbon dating of ultra-small samples (Bard et al., 2015; Wacker et al., 2013), we are now in a situation to (i) precisely determine the ages and durations of BBDC events and (ii) set a reliable deglacial chronology of meltwater discharge events from the North American-Arctic and western Greenland ice sheets.

Figure 1. Location of Baffin Bay, surrounding land masses (BI = Baffin Island), ‘modern’ oceanography and core location used in this study (SL 170 and SL 174). The location of core PC 16 (Simon et al., 2014, 2012) is also shown. Warm currents originating in the North Atlantic are shown by solid red lines (WGC = West Greenland

Current). Cooler currents of Arctic origin are shown by blue dashed line (BIC = Baffin Island Current; LC = Labrador Current). Areas of meltwater/water influx into the Baffin Bay are also shown (DS = Davis Strait; DB = Disko Bugt; UT = Uummannaq Trough; NS = Nares Strait; JS = Jones Sound; LS = Lancaster Sound; HB = Hudson Bay; HS = Hudson Strait).

2. Methods and Material

2.1 Core locations

Gravity cores GeoTü SL 174 (68° 31.88' N 63° 19.82' W, MSM09/02-0467/3) and GeoTü SL 170 (68° 58.15' N 59° 23.58' W, MSM09/02-0455/13) were retrieved aboard the *RV Maria S. Merian* research cruise MSM09/02 (Kucera et al., 2014) in 2008. Cores are herein referred to by number only. SL 174 (777.5 cm in length) was retrieved from 1559 m water depth off the coast of Baffin Island. SL 170 (683 cm in length) was retrieved from 1078 m water depth in the eastern region of Baffin Bay, to the west of Disko Bugt (Figure 1). PARASOUND™ surveys at the core sites and magnetic susceptibility profiles (Figure S1) carried out at 1 cm intervals on board, show no evidence of turbidites or hiatuses in sediment deposition, thus indicating promising records for a cross-basinal study of deglacial history of the ice sheets surrounding Baffin Bay. Sediment cores were stored at <4°C and split lengthways into archive and working halves.

2.2 Chronology

Age models were constructed for both cores using Accelerator Mass Spectrometry (AMS) ¹⁴C dating of biogenic carbonate. To this end, foraminifera were picked from the >150 µm fraction of processed sediment samples in both cores. Either planktonic (*Neogloboquadrina pachyderma* sinistral) or mixed benthic foraminifera samples were dated. The selection depended on availability of sufficient material at the intervals of interest and where possible, replicate dates based on planktonic and benthic foraminifera within the same sample were obtained. Also included is one date based on mollusc fragments from core SL 170. Two methods were used for the radiocarbon dating; traditional ¹⁴C measurements with AMS on samples converted to graphite and a new method, where CO₂ from ultra-small amounts (~0.5 mg) of carbonate (Wacker et al., 2013) were directly analysed with a compact AMS facility equipped with a gas ion source at the Laboratory for Ion Beam Physics, ETH Zurich.

Some of these samples were also leached before measurement (Bard et al., 2015). There is good agreement between the methods, suggesting the dating procedure is robust.

The chronology for both cores was constructed using BACON (Blaauw and Christen, 2011), a Bayesian accumulation model code using an autoregressive gamma process. The code is open-source and run in the program 'R'. This method of constructing an age-depth model is more flexible than classical age-depth modelling, such as CLAM (Blaauw, 2010), as it allows for variable sedimentation rates between radiocarbon dates. Initially, all AMS ^{14}C radiocarbon dates were considered in the development of the age model. The AMS ^{14}C ages and their associated errors were calibrated within the age-depth modelling process, using the Marine13 radiocarbon age calibration curve (Reimer, 2013). In addition, we applied a local reservoir correction (ΔR) of 140 ± 35 years as estimated from historical but pre-bomb samples (Lloyd et al., 2011), currently used for radiocarbon date calibrations of Disko Bugt sedimentary sequences (Jennings et al., 2014; Lloyd et al., 2011; Ouellet-Bernier et al., 2014; Perner et al., 2013, 2011). Since the local reservoir correction in high latitude regions remains poorly constrained over the deglaciation, we carried out a detailed sensitivity assessment of the chronology of the two cores against the choice of local reservoir correction. These results are thoroughly addressed and the effects of changing local reservoir correction are evaluated in the discussion section 4.2.

Lab code	Depth (cm)	Material	Weight (mg)	^{14}C age (yrs)	^{14}C (+/- yrs)	^{14}C age benthic - planktonic	Min age (yrs BP)	Max age (yrs BP)	Mean age (yrs BP)
ETH-55690	97-100	planktonic (<i>N.pachyderma</i>) and mixed benthic	0.32	9793	120		10262	10961	10592
Beta-344508	122-125	planktonic (<i>N.pachyderma</i>) and mixed benthic	4.16	10390	40		11089	11585	11262
ETH-55691	142-145	mixed benthic	0.56	10997	110		11512	12346	12018
ETH-58356	169-170	mixed benthic	0.91	11010	85		12088	12586	12370
Beta - 344505	196-199	planktonic (<i>N. pachyderma</i>)	8.54	11410	50		12547	12787	12672
Beta - 344506	196-199	mixed benthic	12.72	11150	50	-260			
KIA 40767	215-219	planktonic (<i>N. pachyderma</i>)	1.1	12000	80		13035	13514	13288

Beta - 344507	233- 236	planktonic (<i>N. pachyderma</i>)	3.42	12580	60	13569	14134	13.879
ETH- 58357	278- 279	planktonic (<i>N. pachyderma</i>)	0.91	14510	120	15852	17401	16807
ETH- 58358	294- 295	planktonic (<i>N. pachyderma</i>)	0.98	15060	110	17122	17984	17587
*ETH- 58360.1	384- 385	planktonic (<i>N. pachyderma</i>)	0.57	22380	289	25831	26701	26235
*ETH- 58360.3	384- 385	planktonic (<i>N. pachyderma</i>) replicate	0.73	22640	190			

Table 1. Radiocarbon dates and modelled ages for core SL 174. All ^{14}C radiocarbon dates with the ETH lab code were measured as gas samples for this study at the Laboratory for Ion Beam Physics, ETHZ; other codes are for ^{14}C dates measured using the traditional AMS ^{14}C method. All ^{14}C ages were calibrated using the Marine13 dataset (Reimer, 2013) and a ΔR of 140 ± 35 years was applied in the age modelling software BACON. Maximum and minimum ages represent the 95% (2σ uncertainty) in the age model. No dates were considered outliers by the BACON age model. *The ages obtained from sample interval 384-385 cm were not included in the deglacial age model (see text for details).

Lab code	Depth (cm)	Material	Weight (mg)	^{14}C age (yrs)	^{14}C (+/- yrs)	^{14}C age benthic - plankton	Min age (yrs BP)	Max age (yrs BP)	Mean age (yrs BP)
ETH-55678	24-27	mixed benthic	0.47	9668	112		9930	10432	10226
ETH-55679	35-37	mixed benthic	0.81	9460	80		10095	10500	10310
ETH-55680	55-57	mixed benthic	1.03	9833	83		10376	10747	10572
ETH-55681	74-76	planktonic (<i>N. pachyderma</i>)	0.79	9901	82				
ETH-55682.1	74-76	mixed benthic	0.87	10028	87	127	10652	10966	10797
ETH-55682.2	74-76	mixed benthic (replicate)	0.86	10090	97	189			
ETH-55683.1	98-100	mixed benthic	0.81	10243	80		10877	11200	11059
ETH-55683.2	98-100	mixed benthic (replicate)	0.90	10232	137				
ETH-55684	116-118	planktonic (<i>N. pachyderma</i>)	0.44	11042	107				
ETH-55685	116-118	mixed benthic	0.93	10274	86		11012	11471	11213

Beta - 344504	136-139	mollusc fragments	12.22	10080	50			
ETH-55686	136-139	mixed benthic	0.71	12990	117			
ETH-58351	159-160	mollusc fragments	0.85	10755	85	11380	11974	11694
ETH-58352	159-160	mixed benthic	0.96	10905	85			
ETH-58353	180-181	mixed benthic	1.08	10671	85	11516	12105	11845
ETH-55687	266-269	mixed benthic	0.92	11267	100	12251	12656	12498
ETH-58354	288-290	mixed benthic	0.97	11150	75	12414	12739	12601
ETH-55688	399-402	planktonic (<i>N. pachyderma</i>)	0.69	11597	104	13040	13369	13213
ETH-55689	399-402	mixed benthic	0.87	11944	92	347		
KIA 40766	484-488	planktonic (<i>N. pachyderma</i>)	1.20	12730	60	13879	14573	14138
ETH-58355	636-637	planktonic (<i>N. pachyderma</i>)	0.85	14640	130	16435	17462	16992

Table 2. Radiocarbon dates and modelled ages for core SL 170. All ^{14}C radiocarbon dates with the ETH lab code were measured for this study as gas samples at the Laboratory for Ion Beam Physics, ETHZ; other codes are for ^{14}C dates measured using the traditional AMS ^{14}C method. All ^{14}C were calibrated using the Marine13 dataset (Reimer, 2013) and a ΔR of 140 ± 35 years was applied in the age modelling software BACON. Maximum and minimum ages represent the 95% (2σ uncertainty) in the age model. ^{14}C dates shown in grey were considered outliers and thus removed from the final age model presented here (see supplementary information for details).

2.3 Physical sediment properties

Computerised tomography (CT) and high-resolution colour linescan (LS) imaging were carried out on the archive halves of both cores at the University of Bremen, Germany. Lithofacies were determined from these along with visual analysis, with particular interest in areas of coarser, potentially ice-rafted sediment (Figures 2 and 3). L^* , a measure of lightness and component of the CIE/ $L^*a^*b^*$ colour space, was measured during linescan image acquisition. L^* is unit-less value, but represents lightness (in this case of sediment) on a scale from darkest black ($L^* = 0$) to brightest white ($L^* = 100$). L^* values are often used as a proxy for relative carbonate content in marine sediment cores; high L^* values indicating higher carbonate content (e.g. Balsam et al., 1999). To determine grain size distribution

throughout the core and also to obtain samples for radiocarbon dating, the working half of both cores was continuously sampled in slices of 1 cm thick slices. A portion of the slice corresponding approximately to 10 g dry weight was freeze dried, weighed, washed over 0.063 mm screen until all fine fraction was removed, the sieve residue was collected, dried and weighed again. The proportion of sediment in this > 63µm fraction was calculated as the ratio between the dry weight of the residue and dry weight of the bulk sample.

2.4 X-ray fluorescence (XRF) and quantitative X-ray diffraction (qXRD) analysis

XRF scanning was used to determine elemental properties in the cores. Data were collected every 1cm along the archive half surface of both cores with the XRF Core Scanner II™ (AVAATECH™ serial No. 2) at MARUM, University of Bremen. Sediment intervals containing large amounts of calcium (Ca) were noted, as in previous studies, as an indication of detrital input from source areas surrounding northern Baffin Bay (Aksu and Piper, 1987; Andrews et al., 2014, 1985; Simon et al., 2014). These potential detrital carbonate layers were sub-sampled for qXRD analysis to establish their mineralogy. qXRD was measured on a Bruker D8 Advance™ diffractometer at the University of Bremen. Of particular interest was the proportion of dolomite and calcite within detrital carbonate rich layers. For full details of XRF and qXRD methods see supplementary information.

3. Results

3.1 Deglacial chronology

The chronology of the two sediment cores is based on a total of 28 radiocarbon dates (Table 1 and 2). The distribution of the calibrated ages reveals that the uppermost portions of both cores, which contained fine-grained sediment and were barren of biogenic carbonate, corresponds to the later part of the Holocene with oldest ages being > 10 kyr BP (Table 1 and 2). The age models of both cores reveal continuous deposition during the last deglacial and early Holocene period (~17-10 kyr BP), albeit at different resolutions. The interval of highest sedimentation rates and thus of clastic delivery appears synchronous across the two cores (~12.6 kyr BP).

In core SL 174, the barren Holocene section corresponds to the top 98 cm. Further downcore, the progression of ages indicates undisturbed sedimentation (Figure 2). As a result, the Bayesian age model could be constructed using all ages, including a replicate age measurement on both planktonic

and benthic foraminifera in one sample interval. A large amount of planktonic foraminifera in the sample interval 384–385 cm allowed for a technical replicate; the extracted foraminifera were split and resulting aliquots were dated separately. The replicate planktonic-benthic age differed by about the same amount as the technical replicate in the oldest dated sample at 384–385 cm (260 ^{14}C years) (Table 1). A (calibrated) date of ~ 17.6 kyr BP at 294–295 cm is the first date bracketing the deglacial period. An older (calibrated) ^{14}C age of ~ 26.2 kyr BP obtained from 384–385 cm indicates that the lower part of the sediment sequence in the core is older than the LGM. The overlying 10 radiocarbon dates included in the age model indicate that the corresponding 197 cm-long interval covers the period ~ 17.6 – 10.6 kyr BP (Figure 2). The initial part of the deglacial record of SL 174 is characterised by low sedimentation rates of 14 – 23 cm kyr $^{-1}$ up to 234 cm downcore (~ 13.9 kyr BP), above it increasing about twofold to 28 – 32 cm kyr $^{-1}$. Highest sedimentation rates of 85 – 95 cm kyr $^{-1}$ are observed between 197 and 143 cm (~ 12.7 – 12.0 kyr BP).

A total of 21 ^{14}C ages, including 3 planktonic-benthic replicate ages, 2 benthic (technical) replicates and 2 benthic-mollusc fragment replicate ages were obtained for core SL 170 (Table 2). There were insufficient foraminifera in the top 24 cm and bottom 46 cm of the core for radiocarbon dating. Three radiocarbon ages fell outside the 95% uncertainty range of the age-depth model (see Figure S2 and supplementary information for details). These ages were removed from the final age model run, including one replicate planktonic age (116–118 cm) and two replicate benthic-mollusc fragment ages (136–139 cm). The remaining replicates showed excellent reproducibility, with plankton-benthos age differences of the same size as differences between benthic replicates. Based on the retained 18 ages, the corresponding 613 cm sedimentary sequence could be dated to the period ~ 17.0 – 10.2 kyr BP (Figure 3). The calculated sedimentation rate was low (~ 56 cm kyr $^{-1}$) during the deposition of the 636–486 cm interval (~ 17.0 – 14.1 kyr BP). A twofold increase in sedimentation rate (up to ~ 96 cm kyr $^{-1}$) occurred at ~ 14.1 kyr BP, and from then on the sedimentation rate remained high, varying between 87 and 257 cm kyr $^{-1}$ (Figure 3).

3.2 Physical sediment properties

In core SL 174 the interval 295–240 cm (~ 17.6 – 14.2 kyr BP) is characterised by laminated muds (Figure 2). Within this interval, the > 63 μm fraction remains below 20% and L^* values fluctuate

219 around 45. The following section is coarsening upward towards a thick layer of massive pebbly mud
220 between 240–195 cm (~14.2–12.6 kyr BP). Gravel-sized clasts embedded in the finer matrix, likely
221 representing dropstones, are abundant within this interval. Between 195 and 163 cm (~12.6– 2.2 kyr
222 BP), a layer of darker sediment (L^* value ~40) composed of a finer laminated mud, with lower > 63
223 μm fraction content ($\leq 10\%$). A layer of massive pebbly mud, with dropstones, 30% > 63 μm fraction
224 and L^* values of 50–60, are found between 163–111 cm (~12.2–10.9 kyr BP). The rest of the core is
225 composed of laminated mud and a much reduced content of the > 63 μm fraction. Occasional
226 dropstones are found until 75 cm. A brown mud layer of less than 5 cm caps the top of the core.

227 The base of core SL 170 to 510 cm (~17.0–14.6 kyr BP) is characterised by fine stratified mud (Figure
228 3) with a very low content of the > 63 μm fraction. Above 510 cm, a thin layer of pebbly mud with
229 dropstones is followed by a sandy mud layer with a > 63 μm fraction content of up to 50% (505–470
230 cm; ~14.5–13.9 kyr BP). Up to 393 cm (~13.2 kyr BP), the core then consists of stratified fine muds.
231 Laminated sandy and pebbly muds characterise the sediment from this point on, with abundant
232 dropstones and recognisable bioturbation between 360 and 320 cm (~12.9 to 2.7 kyr BP). Another
233 pebbly layer corresponds to the period ~11.5 to 11.1 kyr BP (145–110 cm). The top 60 cm of the core
234 is characterised by a sandy mud layer (> 63 μm 10–25%) with occasional dropstones. While L^* values
235 fluctuate here more frequently than in SL 174, coarser layers between 320 and 70 cm are composed
236 of consistently lighter sediment (L^* ~47–49). The top of the core is capped with a ~ 4 cm thick brown
237 mud layer.

Figure 2. Chronology and lithostratigraphy for core SL 174. The age-depth relationship is based on 10 ¹⁴C dates (filled circles). Error bars indicate minimum and maximum ages (95% uncertainty, 2σ) for dated intervals from the age model. Sedimentation rate (shown on log scale) was calculated between each dated interval, the black line indicating the mean sedimentation rate and grey area indicating the error (95% uncertainty). Computerised tomography (CT) and colour linescan images (LS) are shown as well as sediment lightness (L*) measurements. The percentage of sediment > 63 µm is shown in green. Ca/Sr ratio downcore is shown in purple. The lithofacies are defined from CT, LS images and visual inspection of the core.

Figure 3. Chronology and lithostratigraphy for core SL 170. The age-depth relationship is based on 18 ^{14}C dates (filled circles). Error bars indicate minimum and maximum ages (95% uncertainty, 2σ) for dated intervals from the age model. Sedimentation rate (shown on log scale) was calculated between each dated interval, the black line indicating the mean sedimentation rate and grey area indicating the error (95% uncertainty). Computerised tomography (CT) and colour linescan images (LS) are shown as well as sediment lightness (L^*) measurements. The percentage of sediment $> 63 \mu\text{m}$ is shown in green. Ca/Sr ratio downcore is shown in orange. The lithofacies are defined from CT, LS images and visual inspection of the core.

3.3 Detrital carbonate layers: XRF and XRD analysis

X-ray fluorescence (XRF) scans were used to characterise elemental properties of the sediment and thus identify detrital carbonate layers that were found in this time interval in Baffin Bay sediments during previous studies. We use the ratio of calcium (Ca) to strontium (Sr) counts to define these layers, as used in studies of North Atlantic and Labrador Sea sedimentary sequences, to identify detrital carbonate layers (Channell et al., 2012; Hodell et al., 2008; Pearce et al., 2013; Winsor et al., 2012). In core SL 174, the first increase in Ca/Sr (> 30) occurs between 240 and 217 cm. A subsequent peak is seen between 200 and 185 cm, and the thickest one between 165 and 115 cm (Figure 2). Outside these intervals, Ca/Sr values remain between 5 and 30. In core SL 170, three intervals with high Ca/Sr ratios (> 15) are observed at depths of 510–460, 285–190 and 145–110 cm (Figure 3). Outside these intervals Ca/Sr ratios remain between 5 and 13.

qXRD measurements were carried out on a subset of samples where high Ca/Sr values were observed. Of particular interest was the proportion of dolomite and calcite within layers with high Ca/Sr values, which were expected to contain detrital carbonate. The results of qXRD analyses indicate that in these carbonate-rich layers from SL 174, total carbonate percentages are between 23 and 55% with a high proportion of dolomite (17–40%) relative to calcite (7–17%). The proportion of total carbonate in detrital carbonate layers is lower in SL 170 (4–16%). However as found in SL 174, there is a consistently higher proportion of dolomite (4–12%) as compared to calcite (1–4%) in all samples analysed. All data presented here will be made available online at the PANGAEA portal.

4. Discussion

4.1 Identification and timing of Baffin Bay detrital carbonate (BBDC) events

In both cores, the base of the oldest detrital carbonate layer is associated with an almost twofold increase in sedimentation rates. In core SL 174, the detrital carbonate delivery event is characterised by a very pebbly mud layer with abundant dropstones, possibly indicating iceberg rafting (Figure 2). In SL 170 this layer is associated with the deposition of a less coarse, sandy layer, with occasional dropstones (Figure 3). Within the age model uncertainty, this layer (and thus the event that led to its deposition) appears synchronous on both sides of Baffin Bay, beginning ~14.2 kyr BP and ending ~13.7 kyr BP (Figure 4). This event corresponds to the first observed detrital carbonate delivery in

Baffin Bay during the last deglaciation and thus appears to correspond to BBDC 1 as described in previous Baffin Bay records (Andrews et al., 2014, 1998, Simon et al., 2014, 2012).

A subsequent increase in detrital carbonate input is inferred to start at ~12.7 kyr BP in SL 174 and ~12.6 kyr BP in SL 170, i.e. at about the same time within the age model uncertainty. This second detrital carbonate layer is associated with an almost four-fold increase in sedimentation rates in both cores, peaking at ~12.6 kyr BP (Figures 2 and 3) and by lithology and physical properties similar to those of the older BBDC 1 in each core. However, peak accumulation rates at ~12.6 kyr BP are associated with different geochemical signatures on both sides of Baffin Bay. On the Greenland side, the Ca/Sr values are at their maximum, whereas on the Baffin Island side, there is a distinct minimum in Ca/Sr (Figure 4). The manifestation of this event in SL 174 is almost identical to that observed in the deeper and more northern core PC 16 (Figure 1; Simon et al., 2014, 2012). In both western Baffin Bay cores, this event is characterised by a decrease in the proportion of > 63 μm fraction (Figure 2) and an increase in Ti/Sr ratios (Figure S3), a proxy for increased terrigenous (vs. biogenic) input. The common internal structure of this event in both PC 16 and SL 174 could therefore reflect a regional change in the source and pattern of sediment delivery to the basin. Indeed, this finer, Ti-rich sediment is thought to indicate a lateral mode of sediment delivery from ice stream advance/retreat in Baffin Bay (Aksu and Piper, 1987; Hiscott et al., 1989). More recent provenance studies identify a predominantly western Greenland (Uummannaq) source of these fine grained layers (Simon et al., 2014). The end of this event in both cores studied here is characterised by a secondary peak in Ca/Sr at ~11.4 kyr BP followed by a gradual decline in Ca/Sr until ~11.0 kyr BP. The timing of this event and its structure on the Baffin Island side suggests it is the previously identified BBDC 0 event (Andrews et al., 1995; Jennings et al., 2014; Simon et al., 2014, 2012).

Our identification of these detrital carbonate layers as BBDC layers is further supported by qXRD results (Figure 5). The bulk sediment composition in both BBDC events shows calcite/dolomite ratios consistently lower than ~0.25:1 in cores SL174 and SL 170. When compared with core PC 16, the total carbonate and dolomite therein decreases with increasing distance from the northern Baffin Bay source area, to which core PC 16 is most proximal and core SL 170 most distal along the path of the counter-clockwise circulation (Figure 1). This suggests that ice-rafting of dolomite rich carbonates

308 from the major northern ice streams, such as Nares Strait/Smith Sound and Lancaster Sound,
309 followed the southward path of the present-day Baffin Island current and re-circulated east toward the
310 centre of Baffin Bay north off Davis Strait (Figure 1), as also suggested by Andrews et al. (2011). The
311 composition of the BBDC 1 layer is distinctly different from Heinrich Event 1 layers from the Hudson
312 Strait (Andrews et al., 2012, 1995) and from this we can assume that icebergs discharged from the
313 Hudson Strait did not recirculate north into Baffin Bay (Figure 5) BBDC 0 has a higher proportion of
314 carbonate (and dolomite) perhaps indicating larger amounts of detrital carbonate among the ice rafted
315 debris than during the deposition of BBDC 1. In fact, composition of the presumed H0 event in the
316 Labrador Sea has a distinct signature more closely following that of BBDC events in the Baffin Bay
317 (Figure 5).

318 The synchronous nature of the BBDC 1 and 0 events that span Baffin Bay provides evidence for two
319 periods of major North American-Arctic ice-sheet instability events Baffin Bay during the last
320 deglaciation. This model of sediment delivery (Andrews et al., 1998; Hiscott et al., 1989) explains the
321 sedimentation pattern along the Baffin Island margin (PC 16 and SL 174), but the higher
322 sedimentation rate in SL 170 during both BBDC events requires an additional, local source of
323 sediment during these events. The proximity of SL 170 to the paleo-ice streams of Disko Bugt
324 suggests material delivered by the dynamic Jakobshavn Isbrae as the most likely local source of
325 clastic sediment. Indeed, these ice streams retreated via iceberg calving during the interval of BBDC 0
326 (Jennings et al., 2014) and sediment originating from nearby west Greenland could have potentially
327 caused dilution of the carbonate-rich signal in SL 170.

Figure 4. Timing of BBDC events in central Baffin Bay during the last deglaciation. Elevated Ca/Sr ratios for SL 170 (b) and SL 174 (c) depict the two major BBDC events during the deglaciation, highlighted by grey shading. Radiocarbon dates for each record are shown by triangles of corresponding colour. Average sedimentation rates are also shown for both cores in the corresponding colour (d). BBDC 1 post-dates the onset of Bølling warming (B) and continues into the Allerød (A) indicated by new analysis of Greenland ice core records (a) (Buizert et al., 2014) and the Heinrich 1 Event (H1) in the Atlantic (Carlson and Clark, 2012) (d). BBDC 0 occurs at the onset of the Younger Dryas (YD) stadial and ends apparently slightly after the YD-end, possibly overlapping the newest timing of North Atlantic Heinrich Event 0 in the Labrador Sea (d) (Jennings et al., 2015; Pearce et al., 2015).

Figure 5. Detrital carbonate composition during BBDC events in central Baffin Bay. Core PC 16 (Simon et al., 2014) has the highest concentration of carbonate and dolomite whilst the most distal core (SL170, this study) has the lowest. Composition of detrital carbonate layers found south of the Davis Strait (HU97048-007PC and HU87033-009) during Heinrich Event 1 are shown for comparison. Layers depositing during the presumed H0/BBDC 0 event have a higher dolomite component within detrital carbonate layers (Andrews et al., 2014, 2012). Allocation of these sediments to Heinrich layer timing is based on the original age models for these cores.

4.2 Age uncertainty due to variable local ^{14}C reservoir effect in Baffin Bay

The construction of a radiocarbon chronology and dating of the observed abrupt ice-sheet instabilities are associated with uncertainties, which deserve consideration prior to full discussion of their relation to regional as well as global climatic changes. Of particular importance is the effect of a local ^{14}C reservoir age (ΔR) in Baffin Bay; the magnitude and variability of which is not well constrained (e.g. Pearce et al., 2015). Previous radiocarbon chronologies in Baffin Bay and surrounding regions have employed ΔR values ranging from 0 years (e.g. Levac et al., 2001; Knudsen et al., 2008; Gibb et al., 2015) to 400 years in the northern Baffin Bay (e.g. Ledu et al., 2010). Recent studies in the Disko Bugt area use a ΔR of 140 ± 35 yrs (Jennings et al., 2014; Lloyd et al., 2011; Ouellet-Bernier et al., 2014; Perner et al., 2013, 2011), estimated from historical but pre-bomb samples. Using $\Delta R = 140 \pm$

35 yrs at the site of SL 174 is also in line with studies of regional reservoir correction in Arctic Canada, that indicate local reservoir correction of $\Delta R = 150 \pm 60$ yrs in the south east Baffin Island region (Coulthard et al., 2010). No studies to-date address radiocarbon dating and the effect of ΔR in the region of the deeper cores used here. Furthermore, we lack dateable material independent of any reservoir effect (e.g., pollen or tephra), in order to test for reservoir effects.

Instead, to assess potential variability in the timing of the observed BBDC events, we generated multiple age models with BACON under same settings but varying the ΔR and its associated uncertainty ($\pm$). Due to the higher resolution and higher number of available ^{14}C dates we focused on the core SL 170 and tested the impact of changing ΔR on the onset and termination of BBDC events (at cm depth). The results can be seen in Figure 6. While increasing ΔR from the published standard of $\Delta R = 140 \pm 35$ yrs (total correction = $\sim 540 \pm 35$ yrs) to $\Delta R = 400 \pm 100$ yrs (total correction = $\sim 800 \pm 100$ yrs) has the expected effect of making these events younger, the onset of the events even after application of these variable corrections remains within the 95% confidence intervals of the original age model. Importantly, none of the corrections align the timing of these events with Greenland temperature changes or Hudson Strait Heinrich Events (Figure 6).

Three sample intervals at depths 74–76 cm (SL 170), 399–402 cm (SL 170) and 196–199 cm (SL 174) yielded enough material for planktonic and benthic radiocarbon dating, providing potentially another means of assessing ΔR (Tables 1 and 2). The (uncalibrated) difference in ^{14}C ages between planktonic and benthic ranges from 127 to 347 yrs (mean = 221, standard deviation = 113) in core SL 170, whereas the planktonic age SL 174 is 260 years older than the benthic age. This is a feature occasionally observed in deep sea sediments and assigned either to the export of fresh young carbon from the photic zone to the sea floor or to distinct ventilation ages of sub-surface versus bottom waters (Bauer et al., 1995; Lund et al., 2011; Wu and Hillaire-Marcel, 1994). Whilst this does not permit a quantitative assessment of the offset between planktonic and benthic radiocarbon dates, it gives an indication that the difference does not vary by more than a few centuries. This small difference also allows us to accept the two-topmost dates (Holocene) for SL 174, which, due to low abundance of planktonic and benthic foraminifera, required a mixed sample for dating (Table 1).

A $\Delta R = 140 \pm 35$ yrs (total correction = 540 ± 35 yrs) is more than likely sufficient for correcting ages based on planktonic foraminifera. While Baffin Bay is and was subject to extensive and persistent sea-ice cover, planktonic foraminifera would inhabit the top ~150 m of the water column, with well ventilated water (Figure 1). Although our planktonic and benthic dates agree with each other (Tables 1 and 2), we nonetheless tested the difference between applying $\Delta R = 140 \pm 35$ yrs to all radiocarbon dates (age model presented here) and keeping the ΔR the same for radiocarbon ages based on planktonic foraminifera, while increasing it to $\Delta R = 400 \pm 50$ yrs for radiocarbon ages based on benthic foraminifera (Figure 6; variable ΔR). This also has the effect of making the onset and termination of the BBDC 0 event younger, but the new dates remain within the uncertainty (95% confidence interval) of the original age model. Differences are minimal in the deeper sections of both cores (BBDC 1), where most pre-Younger Dryas ^{14}C dates are based on planktonic samples. Small differences between models for core SL 174 (Figure 6), where most dates are based on planktonic material, also supports this finding. The disintegration of the Greenland and Innuitian (Ellesmere) ice sheets and subsequent opening of the Nares Strait was not achieved before ~10 kyr BP (Jennings et al., 2011; Zreda et al., 1999), so the effect on the age model of poorly-ventilated (older) Arctic water entering Baffin Bay can be largely disregarded.

The variability of ΔR through time is unknown in Baffin Bay. A larger ΔR correction of 700–800 yrs was found to be appropriate for the North Atlantic during the Younger Dryas stadial, owing to enhanced sea ice cover and reduced advection of surface waters into the area (Bard et al., 1994). The potential that ΔR could have been larger during the Younger Dryas and the early part of the last deglacial period is something that we cannot quantify in this study. However previous studies in Baffin Bay indicate that extensive sea ice conditions persisted throughout the deglacial until the early Holocene ~7.4–7.3 kyr BP (Gibb et al., 2015; Ouellet-Bernier et al., 2014). From this we could make the assumption that if there is an additional sea-ice correction needed, it would remain relatively constant. Nonetheless, the age model, run with various reservoir ages (Figure 6), gives us an indication as to the effect of increasing reservoir ages on the timing of the recorded Baffin Bay Detrital Carbonate events. Unless using a very high negative reservoir correction, it is not possible to align BBDC 1 with Heinrich Event 1 as recorded in the North Atlantic (Figure 4). The possibility that BBDC 1 was synchronous with the North Atlantic Heinrich Event 0 can also be rejected. Although Bond et

al., (1997) suggested H0 onset at ~12.8 kyr BP, more recent work (Hillaire-Marcel and de Vernal, 2008; Jennings et al., 2015; Pearce et al., 2015), points to a much younger age ~11.5 kyr BP, at the very end of the Younger Dryas (Figures 4 and 7). The onset of BBDC 1 inferred in this study (~14.1 kyr BP) is ~2400 years younger than the revised onset of H0. This would require applying an extremely high $\Delta R > 2000$ years to planktonic radiocarbon dates, which we consider unrealistic. Applying such high reservoir correction would imply that BBDC 0 post-dated the main discharge events from the Disko Bay (Jennings et al., 2014).

In summary, the sensitivity study indicates that the age model using the published local reservoir ($\Delta R = 140 \pm 35$ yrs) on all radiocarbon dates is unlikely to be affected by variable reservoir age effects to such degree that it would allow alternative phase relationship with Greenland temperatures and Labrador Sea Heinrich Events (Figure 4).

Besides local reservoir correction, the possibility of detrital carbonate contamination of the foraminifera tests used for radiocarbon dating, particularly within BBDC layers, also requires discussion. We noted previously that detrital carbonate delivered to the study core sites is likely of Paleozoic age; thus even a minute amount of contamination would yield either substantially older or 'blank' ^{14}C ages, i.e. beyond the range of radiocarbon dating. Radiocarbon dates within a detrital carbonate layer representing the 8.2 kyr BP event in Laurentian Fan sedimentary sequences, for example, required additional reservoir correction; Lake Agassiz waters were depleted in ^{14}C due to underlying Palaeozoic carbonates (Hoffman et al., 2012). Only three radiocarbon ages (including one replicate) had to be removed from the final age model of SL 170 (Table 2) as they caused small reversals (Figures S2). No age reversals occurred in SL 174 (Table 1). Comparisons of dates within BBDC layers with those outside them are consistently well within the error of the age-depth relationship and we can conclude from this that there was little contamination effect. Although the potential impact of a contamination by Palaeozoic carbonate would be significantly more important on ^{14}C -dating than on stable isotope compositions of foraminifera, it is noteworthy that stable isotope measurements in the northwest Labrador Sea of 'dirty' vs. 'clean' foraminifera tests, did not yield significant differences; the 'dirty' foraminifera tests, extracted from detrital carbonate-rich layers,

contained mostly clay-sized grains and minerals and there was little offset between these and the clean foraminifera (Gibb, 2014).

Figure 6. Testing the effect of local reservoir age on the timing of detrital carbonate events in the studied central Baffin Bay cores. Modelled age (kyr BP) for the onset and end of BBDC events 1 and 0 in both cores, using differing local reservoir age (ΔR) and associated uncertainty ($\Delta R \pm$) are shown. All other model-run parameters were kept the same. *variable ΔR run applied a $\Delta R = 140 \pm 35$ yrs to planktonics and $\Delta R = 400 \pm 50$ yrs to benthics ^{14}C dates. Shaded intervals refer to climatic events during the Holocene; YD = Younger Dryas, B/A = Bølling/Allerød, H1 = Heinrich Event 1

4.3 Asynchronous ice sheet instability during the last deglaciation

From the Last Glacial Maximum until BBDC 1, there is no evidence for enhanced detrital carbonate delivery to the Baffin Bay (see also Simon et al., 2014), suggesting only minimal discharge from any of the surrounding ice sheets until ~ 14.2 kyr BP. Our results confirm earlier inferences about the unlikelihood of a Heinrich Event 1 equivalent in Baffin Bay (Andrews et al., 2014, 1998, Simon et al., 2014, 2012) (Figures 4 and 7). Outside of BBDC layers, we observe no sedimentary evidence for any major ice sheet instability. Although precise sediment provenance data are not available, it appears that there is an out-of-phase relationship between the North American Arctic and western Greenland

ice sheets discharging into Baffin Bay and the Laurentide Ice Sheet calving via the Hudson Strait. Our age model test runs (see section 4.2) indicate that this out of phase behaviour is unlikely to be caused by variable local reservoir ages. The conclusion is further supported by the ^{14}C date in SL 174 at 384–385 cm, with an (uncalibrated) age of ~22.3–22.6 kyr BP. Although not included in the age model, it indicates the continuation of low sedimentation rates during the glacial period and thus no enhanced detrital input during Heinrich stadial 1. Furthermore, no Heinrich Event 1 equivalent event was noted in PC 16, north of SL 174 (Simon et al., 2014, 2012).

Contrary to earlier inferences (Simon et al., 2014, 2012), our high-resolution age model indicates that the onset of BBDC 1 postdates the onset of the Bølling interstadial as expressed in northern Greenland atmospheric temperatures (Buizert et al., 2014) by ~500 years (Figure 4). This difference is outside the 95% uncertainty in SL 170 and marginally within uncertainty in SL 174, indicating that the observed phase relationship may reflect a real temporal succession. In this scenario, the onset of the BBDC 1 event could reflect a lagged-nonlinear response (Carlson and Winsor, 2012) of the marine-terminating portions of the North American Arctic and western Greenland ice sheets to the Bølling warming. On the one hand, the onset of BBDC 1 occurred at the very end of meltwater pulse 1A (e.g. Deschamps et al., 2012), at the end of a period of rapid postglacial sea level rise. This would potentially deepen sills such as Davis Strait (e.g. Tarasov et al., 2012) (Figure 1) which in turn could have flooded ice margins. However the role of this mechanism would depend on the source of the meltwater pulse (North American and/or Antarctic ice sheets), which is still under debate (e.g. Liu et al., 2015). Thus we cannot directly implicate this as a mechanism for the onset of BBDC 1 in Baffin Bay. On the other hand, the timing of the BBDC 1 onset follows the spreading of warm water in the deep North Atlantic (Thiagarajan et al., 2014) and invigorated circulation (Liu et al., 2009) at the transition into the Bølling interstadial. This could have resulted in a greater inflow of the warm Irminger component of the West Greenland Current into Baffin Bay (Figure 1). The incursion of warmer sub-surface water was observed off St. Lawrence (Obbink et al., 2010) and the incursion of warmer Irminger Water was observed off southern Greenland at ~14 kyr BP (Jennings et al., 2006; Knutz et al., 2011; Winsor et al., 2012), and there is potential evidence for this incursion prior to 14 kyr BP in the Uummannaq Trough region (Sheldon et al., 2016). This incursion of warm sub-surface water has been implicated as a mechanism for enhancing basal melting of the Uummannaq Trough ice stream

(Sheldon et al., 2016) situated to the north of SL 170. There is little evidence of extensive ice-rafted debris in the BBDC 1 layer in SL 170, but the increased sedimentation rate at ~14.1 kyr BP could reflect increased delivery of local sediment driven by enhanced basal melting along the West Greenland margin (Figure 3), but this requires further investigation.

In western Baffin Bay, BBDC 1 sediments contain significant amounts of detrital carbonates (and abundant dolomite, Figure 5). The coarser sediment, containing abundant dropstones, is more characteristic of ice-rafted deposition. This layer is similar in timing and properties to a DC layer observed in cores from the Uummannaq Trough region (Sheldon et al., 2016) and in PC 16, north of SL 174 (Simon et al., 2014, 2012). Evidence of detrital carbonate layers are also found in the sediments from the Lancaster Sound Trough mouth (Li et al., 2011), although these layers remain undated, the authors assigning these events to the Heinrich Event timescale (Li et al., 2011). Similarly, records of detrital deposition in the Fram Strait, originating from the Arctic-Innuitian and Laurentide ice sheets, are roughly in line with the timing of the BBDC 1 (Darby and Zimmerman, 2008) (Figure 7). However, Darby and Zimmerman (2008) also suggest that Fram Strait ice-rafted debris (IRD) events could be (with age model uncertainties) a manifestation of Heinrich Events, a suggestion weakened by the chronology proposed here.

Although we cannot identify specific source areas in this study, it appears that the BBDC 1 event follows the model of sediment dispersal first proposed by Aksu and Piper (1987) and Hiscott et al., (1989). In this scenario, massive calving from destabilised ice sheets and glaciers surrounding northern Baffin Bay (North American-Arctic and north-western Greenland) resulted in the deposition of detrital carbonate layers along the path of the WGC/BIC (Figure 7). Whether the destabilisation of the North American-Arctic and western Greenland ice sheets and associated discharge were driven by the incursion of a warmer and stronger WGC (Figure 1) at the study sites requires further investigation. Alternatively, BBDC 1 could be a product of some internal ice-sheet instability as has been proposed for Heinrich events. However, we cannot assess these mechanisms (e.g. binge/purge oscillations (MacAyeal, 1993), the subglacial outburst-flow model or catastrophic ice shelf breakup model, summarised by Hesse and Khodabakhsh, 2016; Marcott et al., 2011) with the data presented here.

In contrast, the onset of BBDC 0 (~12.7 kyr BP) is coincident, within uncertainty, with the onset of the Younger Dryas stadial. Unlike the BBDC 1 event that, even with age model uncertainties, clearly postdates Heinrich Event 1, the difference between the timing of BBDC 0 and Heinrich Event 0 is not so clear. This rests on the uncertainties regarding the representation of Heinrich Event 0 in previously studied marine records. In earlier Labrador Sea core studies, the onset of Heinrich Event 0 was seen as coincident with the Younger Dryas (~12.8 kyr BP) (e.g. Andrews et al., 2012, 1995; Hillaire-Marcel et al., 1994; Stoner et al., 1996), but only a small number of these cores in shallower regions show clear evidence of a discrete Heinrich Event 0. Many cores from the deeper north-western Labrador Sea lack a detrital carbonate peak in this late deglacial interval (Andrews et al., 1995; Gibb et al., 2014a; Rashid et al., 2011). Of the few cores with a distinct Heinrich-type layer and adequate radiocarbon dating to constrain the upper and lower limit of the event, the duration of Heinrich 0 is ~12.8 to between 12.0–8.7 kyr BP (as summarised by Rashid et al., 2011). Within these poorly constrained age brackets, BBDC 0 from our records would then be coeval with the start of Heinrich Event 0. However all higher resolution cores present a revised timing of the Heinrich Event 0 onset following the Younger Dryas interval at ~11.5 kyr (Hillaire-Marcel and de Vernal, 2008; Jennings et al., 2015; Pearce et al., 2015) (Figures 4 and 7). This suggests that the onset of ice-sheet instabilities of the northern Baffin Bay predates the H0 discharge from Hudson Strait recorded mostly in the western Labrador Sea.

Evidence from geochemical data and sedimentation rates points to two modes of detrital deposition during the BBDC 0 event in western Baffin Bay (SL 174). Here, the initial carbonate rich and ice-proximal layer is followed by layer of Ti-enriched finer material (Figure S3), followed again by a layer more enriched in carbonate that persisted until the end of the event (Figure 4). While the coarser and more carbonate rich layers were likely transported by iceberg rafting as during BBDC 1 (Figure 2), the fine-grained layer in between requires a different transport mechanism.

Figure 7. Compilation of the timing of ice-rafted and detrital carbonate events in and around Baffin Bay during the late deglacial period. Map (left) shows the paths of detrital carbonate delivery in the Baffin Bay corridor region. Pie charts indicate the composition of BBDC layers in SL 174 and SL 170 (red, this study) and PC 16 (Simon et al., 2014); pie size is relative to the sedimentation rate during these periods. Letters indicate positions of other cores with detrital layers at various locations. The timing of these events according to their respective age models is shown right. Records used: Fram Strait (a) (Darby and Zimmerman, 2008); NW Baffin Bay (b) (Simon et al., 2014, 2012); SL 174 (c, this study); Disko Bugt Shelf (d) (Jennings et al., 2014); SL 170 (e, this study); Uummannaq Trough (f) (Ò Cofaigh et al., 2013; Sheldon et al., 2016); Labrador Sea (g) (Andrews et al., 2014) and North Atlantic Heinrich Events 1 (Carlson and Clark, 2012) and 0 (Pearce et al., 2015). The comparison of records suggests stochastic behaviour (discharge and iceberg rafting) of the ice sheets surrounding the Baffin Bay. IIS = Innuitian Ice Sheet, LIS = Laurentide Ice Sheet and GIS = Greenland Ice Sheet.

Sediments with such characteristics have been previously attributed to lateral mass flows resulting from rapid basal melting of ocean-terminating glaciers (Aksu and Piper, 1987; Simon et al., 2012). This characteristic composition and sedimentation sequence of BBDC 0 in west Baffin Bay is also seen in the more central PC 16, ruling out this layer as a turbidite. This rapidly deposited sub-layer of BBDC 0 could be seen as a dilution event of the more or less continuous deposition of the detrital carbonate-rich ice-rafted debris from northern Baffin Bay sources. Based on a SedUnMix sediment

549 analysis (c.f Andrews and Eberl, 2012; Simon et al., 2014) concluded that at least during the earlier
550 part of the deglaciation, these Ti-rich, finer layers may have been laterally transported from the
551 Uummannaq Trough area (Simon et al., 2014). Although paleoceanographic reconstructions in north-
552 western Baffin Bay are sparse and do not quite cover the onset of BBDC 0, the dominance of the
553 Atlantic Water species *Cassidulina neoteretis* (Perner et al., 2013; Seidenkrantz, 1995) at ~12.5 kyr
554 BP in a sediment core located south of the Nares Strait, suggests some influence of the WGC this far
555 north by this time (Knudsen et al., 2008).

556 Eastward, re-advance of the GIS through Disko Bugt to the shelf edge during the early Younger Dryas
557 and its subsequent retreat around 12.2 kyr BP would be a large source of ice rafted debris. Evidence
558 from sediment cores suggest initial retreat was via calving (Hogan et al., 2016; Jennings et al., 2014;
559 Ó Cofaigh et al., 2013; Rinterknecht et al., 2014), accounting for the elevated sedimentation rates on
560 the Greenland side of Baffin Bay recorded in SL 170 (Figures 3 and 7). This local source would have
561 amplified the effect of enhanced deposition of northern Baffin Bay-sourced detrital carbonate as found
562 in SL 170 and other cores in the Disko Bugt area (Jennings et al., 2014). Whatever the mechanism,
563 the effects of the discharge event from the North American-Arctic ice sheets surrounding the northern
564 portion of Baffin Bay was far-reaching; evidence of a dolomite-rich detrital carbonate component are
565 seen on the Greenland side of Baffin Bay in SL 170 (Figure 7e) and the nearby Disko Bugt shelf
566 (Jennings et al., 2014) (Figure 7d). Furthermore, evidence of a dolomite-rich detrital carbonate layer in
567 the Labrador Sea indicates that these events can travel overspill the Davis Strait (Andrews et al.,
568 2014, 2012, 1995) (Figures 5 and 7g).

569 The end of BBDC 0 at ~11 kyr BP appears concurrent with ice retreat from the carbonate terranes of
570 northern Baffin Bay (Zreda et al., 1999) and coincides with exposure dates of ~11.1 – 10.8 kyr BP
571 from the Lancaster Sound area (Ledu et al., 2010). By this time, transport of detrital carbonate from
572 the northern source of BBDC events was severely diminished. On the west Greenland margin,
573 exposure dates indicate that ice retreated across the Assiaat Sill (south of Disko Bugt) ~11 kyr BP
574 and reached the mainland ~10 kyr BP (Kelley et al., 2013; Young et al., 2013). In northern Disko Bugt,
575 older exposure ages ~11 kyr BP were found at the head of Vagait Strait, indicating that retreat was
576 somewhat faster in this area (Kelley et al., 2015). Similarly Rinterknecht et al. (2014) dated the

collapse of the Jakobshavn Isbrae ice stream at 12.2 ± 0.6 kyr BP. However, correcting for the improper uplift adjustment used, by accounting for the gravitational attraction of water towards ice (Cuzzone et al., 2016; Ullman et al., 2016), would place this collapse at the end of the Younger Dryas. Landward retreat of these ice streams would trap detrital material in fjord areas, exhausting the source of clastic material along this portion of the West Greenland coast to the seafloor at site SL 170.

5. Conclusions

Our new radiocarbon-based chronology indicates centennial synchronicity in the deposition of two distinct detrital carbonate layers across Baffin Bay during the last deglaciation and early Holocene (~17 – 10 kyr BP). We have determined that these layers represent the previously identified Baffin Bay Detrital Carbonate Events (BBDCs). Using the new chronology, we redefine the timing of these events as occurring between ~14.2 – 13.7 kyr BP (BBDC 1) and ~12.7 – 11 kyr BP (BBDC 0). The onset of BBDC 1 (~14.2 kyr BP) appears to have lagged Greenland warming and the event occurred completely out of phase with Heinrich Event 1. Iceberg and meltwater discharge from the North American-Arctic ice sheets during the latter part of the Bølling and into the Allerød interstadial implies the presence of significant amounts of icebergs and meltwater, which could have been channelled into the Labrador Sea through the Davis Strait. This finding questions the common assumption that iceberg discharge events caused reductions in Atlantic meridional overturning circulation and cold stadial conditions (Broecker, 1994).

The onset of BBDC 0 (~12.7 kyr BP) in the Younger Dryas stadial and its continuation into the Holocene (~11 kyr BP) suggest that instability of the North American-Arctic ice sheets could have been stochastic and independent of temperature change. This event was far-reaching and evidence of BBDC 0 is found in other cores on the western Greenland margin (Jennings et al., 2014) and the northwest Labrador Sea (Andrews et al., 2012, 1995). New records from the Labrador Sea suggest that BBDC 0 may have pre-dated the major discharge event from the Laurentide Ice Sheet via the Hudson Strait, Heinrich Event 0 (Pearce et al., 2015). The different timing of iceberg discharge events between Baffin Bay and the Labrador Sea therefore adds to the complexity of our understanding of abrupt climate change. Our results also highlight the importance of considering regional

responses/linkages of ice sheets to/with abrupt climatic changes not only during the last deglacial period, but likely during earlier ice retreat intervals.

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