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Cooling and freshening at 8.2 ka on the NW Iceland Shelf recorded in paired $\delta^{18}O$ and Mg/Ca measurements of the benthic foraminifer *Cibicides lobatulus*

Ursula Quillmann [a,*], Thomas M. Marchitto [a], Anne E. Jennings [a], John T. Andrews [a], Birgitte F. Friestad [b]

[a] Department of Geological Sciences and Institute of Arctic and Alpine Research, University of Colorado, Boulder Colorado, USA
[b] Bjerknes Centre for Climate Research, University of Bergen, Norway

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ABSTRACT

A shallow marine sediment core from NW Iceland provides evidence for a brief cooling and freshening at ~8200 cal yr BP, consistent with the hypothesis that the catastrophic outburst flood of the proglacial lakes Oijbway and Agassiz caused the 8.2 ka event. This is the first high-resolution record reconstructing near-surface temperatures and $\delta^{18}O_{sw}$ by paired measurements of Mg/Ca and $\delta^{18}O_{calcite}$ of a benthic foraminifer. We developed a new Mg/Ca temperature calibration for *Cibicides lobatulus*. Our down-core Mg/Ca derived temperature reconstruction dates the 8.2 ka cooling event between ~8300 cal yr BP and ~8100 cal yr BP, which is similar to the timing and 160-yr duration recorded in the Greenland ice cores. The near-surface temperature drop of ~3 to 5°C during the 8.2 ka event was accompanied by lighter $\delta^{18}O_{sw}$ values. Synchronously to the changes in the geochemical proxies, the percentages of two Arctic benthic foraminifers increased and the percent calcium carbonate decreased. Our record, combined with several others from the region, suggests that the freshwater outburst spread far from the source into the high-latitude North Atlantic. This freshwater input could have directly caused substantial high-latitude cooling, with reduced North Atlantic Deep Water formation amplifying the climatic impact.

Introduction

The high-latitude North Atlantic contains large, poorly understood feedbacks between ice, freshwater, deep-water formation, and temperature that are believed to have played large roles in past abrupt climate changes, and that may be relevant for future climate change. Of particular interest is the so-called 8.2 ka event, because this short-lived climate excursion holds important information about potential causes of abrupt climate changes during the otherwise climatically stable Holocene epoch.

The 8.2 ka event was first documented in Greenland ice core records as a sudden drop in surface air temperatures, with a magnitude ranging from 3 to 8°C, depending on various reconstructions (e.g., Alley et al., 1997; Kobashi et al., 2007). Based on analysis of these ice cores the cooling anomaly during the 8.2 ka event was brief, lasting only 160.5 ± 5.5 yr with a defined cooling peak lasting only 69 ± 2 yr (Thomas et al., 2007). The cooling has been attributed to the catastrophic drainage of the proglacial lakes Agassiz and Ojibway into the North Atlantic during the final deglaciation of the Laurentide Ice Sheet (Barber et al., 1999) (Fig. 1). The timing of the outburst flood has been constrained to the interval between 8040 and 8490 cal yr BP (Lewis et al., 2012), which is later but overlapping with the previously proposed timing between 8160 and 8740 cal yr BP (Barber et al., 1999). It has been hypothesized that the estimated 200,000 km^3 freshwater input from these proglacial lakes spread across the subpolar North Atlantic and suppressed the formation of North Atlantic Deep Water (NADW), thus reducing meridional heat transport and altering atmospheric circulation (e.g., Alley et al., 1997; Barber et al., 1999; Clark et al., 2001; Renssen et al., 2001; Kleiven et al., 2008; LeGrande and Schmidt, 2008; Praetorius et al., 2008; Wiersma et al., 2011). The 8.2 ka cooling could have taken place, however, without changing the rate of northward transport of Atlantic Water, as model results by Clarke et al. (2009) demonstrated. They showed that the freshwater could have promoted sea-ice growth and altered atmospheric circulation, causing substantial cooling over Greenland. In contrast, Condron and Winsor (2011) illustrated in a model study that the proposed outburst flood may have been carried southward in a narrow coastal current to the subtropics, without ever having affected the surface or deep-water regime in the subpolar North Atlantic. The oceanographic changes associated with the 8.2 ka event thus remain controversial.

Paleoceanographic data can help to constrain the potential cause and magnitude of the 8.2 ka event. The 8.2 ka event, in turn, provides a test bed for modeling changes in the density structure of the North Atlantic under near-interglacial boundary conditions. The freshwater input during the 8.2 event lends itself as an analog for present-day freshwater input into the North Atlantic, both from the melting of the Greenland Ice Sheet and the export of sea ice from the Arctic Ocean.

* Corresponding author. Fax: +1 303 492 6388.
E-mail address: Ursula.Quillmann@Colorado.EDU (U. Quillmann).

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Climate anomalies around 8200 yr ago have been tracked around the North Atlantic region (see review by Rohling and Palike, 2005). However, a number of these records show a multi-centennial-scale climate perturbation that started ~8600 cal yr BP and ended ~8000 cal yr BP, onto which the 8.2 ka event cooling may have been superimposed. Rohling and Palike (2005) suggest that the multi-centennial-scale perturbation may be part of a repeating pattern throughout the Holocene, and hence not strictly related to the briefer 8.2 ka event. Here we follow Thomas et al. (2007) in applying the term "8.2 ka event" only to the brief interval between 8247 and 8086 cal yr BP as recorded in the Greenland ice core records. There is limited knowledge of how this brief 8.2 ka event is expressed in the ocean, because open-ocean sedimentation rates are generally low and thus provide poor temporal resolution compared to the highly resolved Greenland ice-core records. Even though the cause of the 8.2 ka event is believed to be known, unequivocal proof is lacking that the sudden and catastrophic freshwater input from the proglacial lakes Agassiz and Ojibway caused the 8.2 ka cooling via an impact on the North Atlantic. In contrast, it has been suggested that the lake drainage caused cooling over Greenland because of an atmospheric response to the altered relative distributions of exposed land, water, and glacial ice in North America (Dean et al., 2002).

In this study, we focused our attention on a narrow time interval encompassing the 8.2 ka event as recorded in the Greenland ice-core record. We aimed to determine the evolution and magnitude of ocean temperature change in the North Atlantic, and whether the cooling was accompanied by a freshening. This was accomplished by investigating sediment core MD99-2266 from the mouth of Ìsafjarðardjúp, a large and shallow fjord in NW Iceland (Figs. 1 and 2). Previous results on the Holocene paleoceanographic evolution at this site, including a comparison to inner fjord conditions, were published in Quillmann et al. (2010). In this new study we performed a higher resolution analysis on the interval from 8400 to 7600 cal yr BP, encompassing the 8.2 ka event.

Regional setting and oceanography

Our study site, MD99-2266 (66°13.77′N, 23°15.93′W, 106 m water depth), is located at the mouth of Ìsafjarðardjúp (Figs. 1 and 2). Ìsafjarðardjúp continues beyond the fjord mouth onto the continental shelf as a shelf trough. The Ìsajarðardjúp trough is slightly shallower seaward of the fjord mouth, but the depth of this threshold is between 110 and 120 m in the center, as shown in the map generated by the Marine Research Institute of Iceland based on multibeam bathymetry data (Fig. 2). The bathymetric map therefore shows no topographic barrier that would restrict open ocean inflow to our 106 m study site at the present.

Three CTD (conductivity, temperature, depth) casts were taken in the Ìsafjarðardjúp fjord system in the summer of 1997 during a research cruise conducted by the Icelandic Marine Research Institute on board the *Bjarni Saemundsson* (Helgadottir, 1997). All three CTD casts display similar characteristics with a thin lid of runoff-influenced low-salinity water above 15 m (Fig. 2). The lack of a discernable intra-fjord spatial pattern in the subsurface salinities suggests that the terrestrial influence is mainly limited to the <15 m layer. All three CTD locations are landward of MD99-2266, and the modern benthic foraminiferal fauna at the three sites are dominated by *Cassidulina reniforme* and *Elphidium excavatum* f. *clavata* (Jennings et al., 2004) typical of Icelandic fjords,

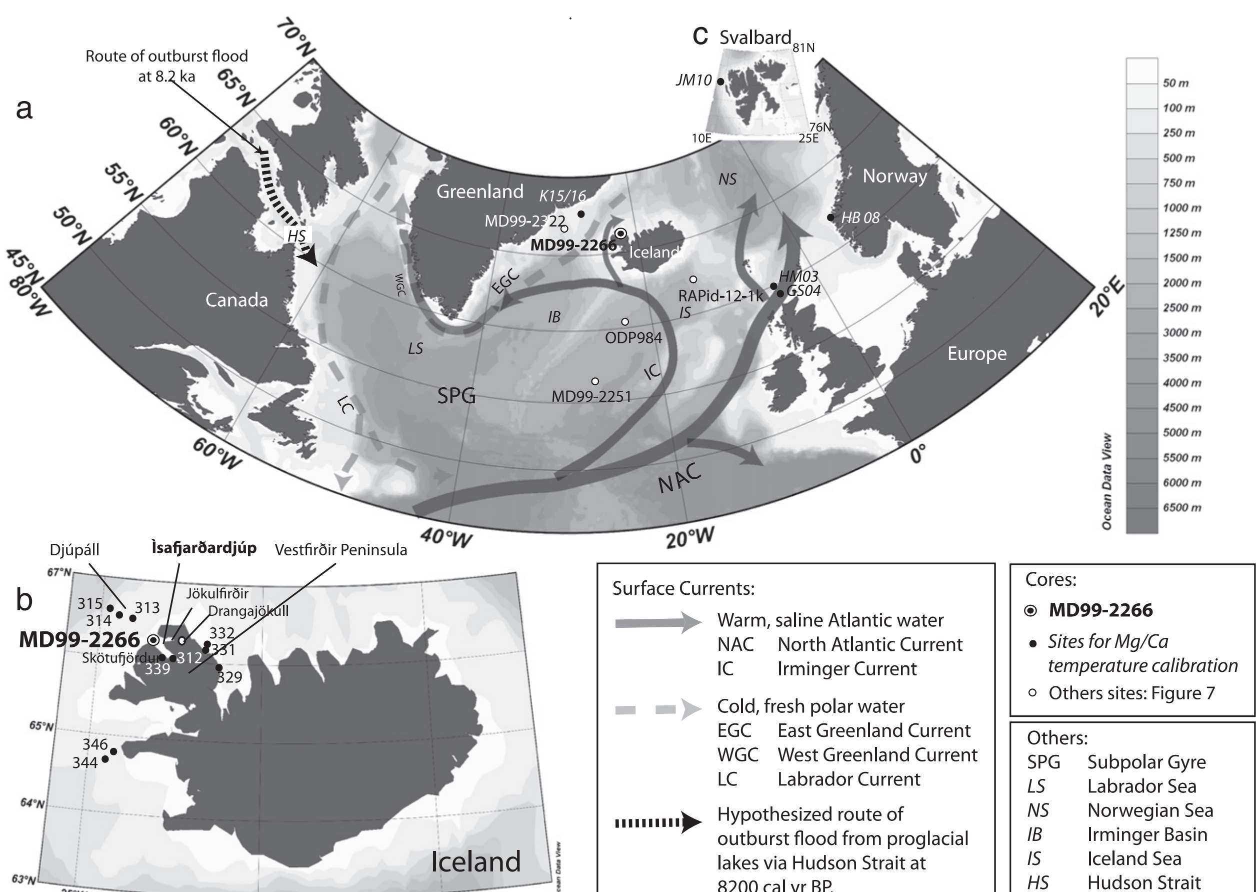

Figure 1. Location map. Panel a shows the high latitude North Atlantic with schematic surface current regime and the hypothesized route of the outburst flood of the proglacial lakes Agassiz and Ojibway at ~8200 cal yr BP. Also shown are Greenland, Faroe Island, and Norwegian sites for Mg/Ca temperature calibrations (black dots) and other regional temperature reconstructions shown in Fig. 7 (white dots). Panel b zooms in on Iceland and provides details about site MD99-2266 (this study). Also shown are the B997 sites for Mg/Ca temperature calibration (black dots). Inset c shows the Mg/Ca temperature calibration sites from Svalbard.

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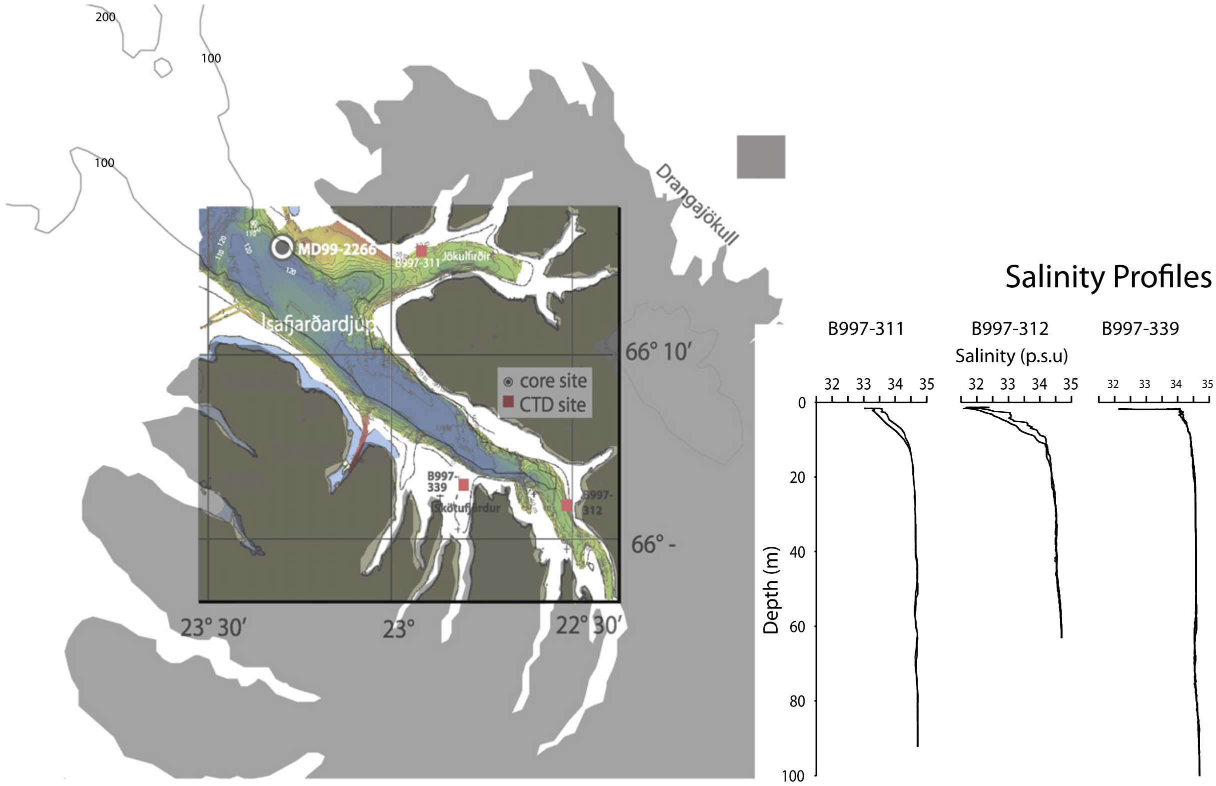

Figure 2. Bathymetry and CTD salinities. The multibeam bathymetry of the Ìsafjarðardjúp fjord system (courtesy of the Marine Research Institute of Iceland) is superimposed onto a map of the region. The depths (m) of various contours at the fjord entrance (pers. comm. Gudrun Helgadottir) are indicated in white font. The salinity profiles from three inner-fjord CTD casts taken during the summer of 1997 are shown, with location marked by square symbols (Helgadottir, 1997).

unlike the fauna at MD99-2266, which displays modern Icelandic shelf fauna. We therefore conclude that the seafloor at the MD99-2266 site is characteristic of open shelf conditions today.

The Vestfirðir Peninsula lies close to the marine polar front, where the northward-flowing, warm and saline Atlantic Water of the Irminger Current (IC) meets the southward-flowing, cold and fresh Arctic Water of the East Greenland Current (Fig. 1). More precisely it is the eastern branch of the IC that is present off western Iceland, where it flows clockwise around the island and is known as the North Iceland Irminger Current (NIIC). Today the Atlantic Water in the NIIC occupies the upper ~200 m of the water column, capped by a ~10- to 20-m-thick lid of fresher water from terrestrial runoff along the NW Icelandic shelf (www.hafro.is). The IC is a branch of the North Atlantic Current (NAC). The NAC is an extension of the Gulf Stream and flows northeastward across the North Atlantic along the southern and eastern boundary of the subpolar gyre, and constitutes the most northern upper limb of the ocean conveyor belt.

Material and methods

MD99-2266

MD99-2266 is a 10-cm-diameter Calypso piston core measuring 38.90 m in length, all of which is Holocene in age. It was raised during the 1999 IMAGES V cruise, Leg III, aboard the *R/V Marion Dusfresne* (Labeyrie et al., 2003).

Chronology

The chronology for MD99-2266 is based on 20 accelerator mass spectrometer (AMS) radiocarbon (^{14}C) dates and the depth of the 10,200 ± 120 cal yr BP Saksunarvatn tephra, as discussed in Quillmann et al. (2010) (Fig. 3). The dates were converted to calibrated years using the CALIB Radiocarbon Calibration online program, version 5.0.2. (Stuiver et al., 1998) with an ocean reservoir correction of 400 yr ($\Delta R = 0$) and an uncertainty of 50 yr in ΔR. Recalibration using the latest version of CALIB (6.1) would change only one of the dates, and by just 14 yr. Kristjansdottir et al. (2007) identified the 8019 ± 277 cal yr BP Suduroy tephra (Wastegard, 2002) on the North Iceland shelf and used it to conclude that the central estimate of the local reservoir age was 430 yr at that time. The interval investigated here, from 7600 to 8400 cal yr BP, is bracketed by dates at 1398–1399 cm (7314 ± 105 cal yr BP) and at 2237–2238 cm (8816 ± 170 cal yr BP) (Fig. 3). Four additional dates between 8101 and 8284 cal yr BP lie within this interval, but we omitted one date at 1784–1785 cm (8252 ± 118 cal yr BP) (Fig. 3B). Although this date falls well within the 2σ range of the underlying date, its inclusion would have introduced a slight age reversal. The combination of dense AMS dating and a reasonably well-constrained reservoir age lead us to estimate that our age control at ~8200 cal yr BP is accurate to about one century. According to Channell et al. (2012), MD99 cores are severely stretched by more than 50% and caution is in order when interpreting sedimentation rates. Regardless of any stretching, our sampling interval of ~10 cm affords an average temporal resolution of ~18 yr.

Geochemical proxies: $\delta^{18}O_{calcite}$ and Mg/Ca

Cibicides lobatulus was chosen because this species is cosmopolitan and epifaunal, and comprised at least 10% of the foraminiferal assemblage in every sample. We added 40 samples between 7600 and 8400 cal yr BP to the already existing $\delta^{18}O_{calcite}$ record (Quillmann et al., 2010). We submitted about ten > 150 µm *C. lobatulus* specimens per

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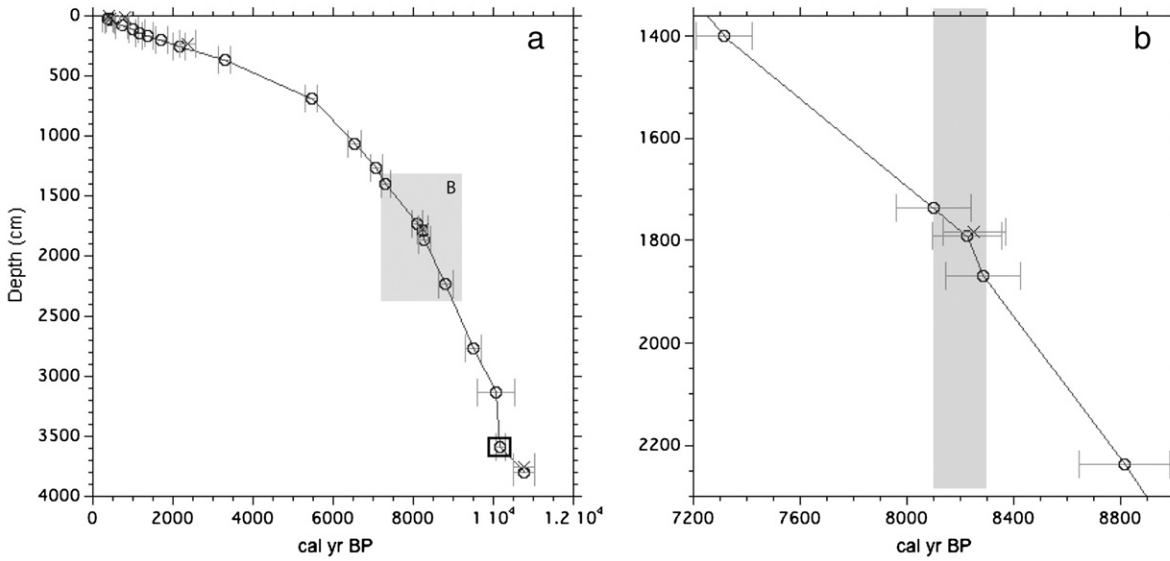

Figure 3. Chronology for MD99-2266 based on ^{14}C AMS dates and the depth of the Saksunarvatn tephra deposit. Dates with their $\pm 2\sigma$ bars are shown. Dates incorporated into the age model are indicated by open, dark gray circles and omitted dates by a light gray x. The date of the Saksunarvatn tephra is indicated by a boxed open circle. A 0%-weighed curve fit was applied to the age-depth data to obtain the age model. In Panel A the age model for the entire core is shown, as published in Quillmann et al. (2010). In Panel B the close up of the two dates bracketing study interval and the four dates falling within the study interval are shown. The vertical light gray band highlights the timing of the 200-yr climate perturbation found in this study.

sample for isotope analysis to the Curry lab at the Woods Hole Oceanographic Institution. Measurements were made on a Finnigan MAT 252 mass spectrometer with a Kiel Carbonate Device. The results are given in conventional δ-notation in ‰ relative to the VPDB standard, with precision better than 0.07‰.

For the Mg/Ca temperature calibration, about 25–35 >150 μm *C. lobatulus* specimens were picked from 31 surface sediment samples collected in the high-latitude North Atlantic. For the paleoceanographic reconstruction, about 40–50 >150 μm *C. lobatulus* specimens were picked from core samples in MD99-2266; abundances of *C. lobatulus* sufficient for Mg/Ca analysis were present in 21 MD99-2266 samples.

Mg/Ca ratios were measured on a Thermo-Finnigan Element2 magnetic-sector single-collector ICP-MS, following methods of Rosenthal et al. (1999) and Marchitto (2006). Crushed samples were cleaned reductively and oxidatively, in a Class-1000 clean lab, according to the cleaning protocol from Boyle and Keigwin (1985) as modified by Boyle and Rosenthal (1996). Long-term 1σ precision for Mg/Ca ratios is 0.5% (Marchitto, 2006).

It has been suggested that in addition to temperature, Mg/Ca ratios potentially are affected by partial dissolution and by the saturation state with respect to calcite (ΔCO_3^{2-}). Partial dissolution of foraminiferal calcite potentially could lower Mg/Ca ratios, thus mimicking a cooling in the reconstructed temperatures, as calcite formed in warmer waters (higher Mg/Ca ratios) is likely more susceptible to dissolution (Rosenthal et al., 2000; Dekens et al., 2002). However, benthic foraminifers spend their entire life cycle in relatively constant water temperatures and thus are expected to be resistant to the partial dissolution effect. Bottom-water ΔCO_3^{2-} seems to influence benthic Mg/Ca ratios during growth, but the effect is likely only significant in the deep sea where ΔCO_3^{2-} is low and varies much more than temperature does (Elderfield et al., 2006). In shallow sites as in this study, ΔCO_3^{2-} is highly saturated and likely is not a major influence on the Mg/Ca ratios. We also note that recent field studies (Ferguson et al., 2008; Mathien-Blard and Bassinot, 2009; Arbuszewski et al., 2010) suggest that the positive effect of salinity on planktonic foraminiferal Mg/Ca may be larger than that determined from live culturing (Lea et al., 1999; Kisakurek et al., 2008), particularly in very saline waters. In the absence of observations supporting a salinity effect on benthic foraminiferal Mg/Ca, we interpret our *C. lobatulus* record solely in terms of temperature.

Biological proxies

Jennings et al. (2004) and Rytter et al. (2002) have shown that changes in modern benthic foraminiferal assemblages on the Iceland shelf coincide with changes in oceanic currents and water masses. We present abundance results in the 106–1000 μm sand fraction for *C. reniforme* and *E. excavatum* f. *clavata*, both Arctic species, found today on the Icelandic shelf in Polar-dominated waters and in fjord environments with cold bottom-water temperatures (Jennings et al., 2004).

Sedimentological proxies

Total inorganic carbon (TIC) percentages were determined by coulometric titration of CO_2 following extraction from the sediment by acid volatilization (Engleman et al., 1985) at the USGS laboratories, Denver, Colorado. Weight percent TIC was converted to weight percent $CaCO_3$ by dividing by 0.12. The accuracy and precision for TIC, determined from hundreds of replicate standards, usually are better than 0.1 wt%.

Results

Mg/Ca temperature calibration for C. lobatulus

Because of its primarily epifaunal habitat, *C. lobatulus* should be a good recorder for bottom-water conditions, but it has not before been used for Mg/Ca-derived temperature reconstructions. We therefore developed a new Mg/Ca temperature calibration for *C. lobatulus* based on high-latitude North Atlantic surface sediment and core top samples. *C. lobatulus* can become detached after death and can be transported to deeper sites, so we limited our selection to samples taken from depths <700 m and chose only pristine, well-preserved individuals.

We measured 44 Mg/Ca ratios in surface sediment samples or core tops from 22 different locations in the high latitude North Atlantic: the west to north Icelandic shelf, the east Greenland shelf, the Norwegian shelf, Svalbard, and the Faroe Islands (Table 1; Fig. 1). The bottom-water temperature at each site was estimated from the CTD cast taken at the time of core collection, ranging from 0.01 to 9.50°C.

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Table 1
Site information for the *C. lobatulus* Mg/Ca temperature calibration. Four samples (indicated by ^{o}) were excluded from the final Mg/Ca temperature regressions because they were $>2\sigma$ outliers relative to initial regression equations. Seven samples (indicated by *) were analyzed at the University of Cambridge using a different cleaning protocol (see text for details) and were also excluded. The Iceland shelf cores were collected during the 1997 joint Icelandic and US cruise aboard the *R/V Bjarni Saemundsson* (B997) (Helgadottir, 1997). The East Greenland cores were collected on the inner shelf off Nansen Fjord during the 1991 joint Icelandic and US cruise aboard *R/V Bjarni Saemundsson* (B1191) (Andrews et al., 1991). Cores from Svalbard were collected during the 2010 cruise aboard *A/S Jan Mayen* (JM10) (Husum, 2010). The core from the west Norwegian margin (HB08-BC) was collected during a 2005 cruise to Byforden, Salhusfjorden and Herdlefjorden aboard the *B/V Hans Brattstrøm* (Hjelstuen et al., 2008). Cores from the Faroe Islands were collected during a cruise in 2003 aboard *R/V Håkon Mosby* (HM03-133) and in 2004 aboard *R/V G.O. Sars* (GS04-138).

Core name	Location name	Sample type	Water depth (m)	Latitude °N	Longitude °W	BWT (°C)	BWS (psu)	Mg/Ca (mmol mol^{-1})
B997-312 s	NW Iceland fjord	Surface	72	65.98	−22.49	5.22	34.69	1.780
B997-312 s	NW Iceland fjord	Surface	72	65.98	−22.49	5.22	34.69	1.988
B997-339 s	NW Iceland fjord	Surface	104	66.02	−22.80	5.30	34.70	1.948
B997-313 s	Djúpáll, NW Iceland shelf	Surface	313	66.61	−23.93	5.87	35.00	1.809
B997-313 s	Djúpáll, NW Iceland shelf	Surface	313	66.61	−23.93	5.87	35.00	1.822
B997-314 s	Djúpáll, NW Iceland shelf	Surface	233	66.70	−24.18	6.03	35.05	1.734
B997-314 s	Djúpáll, NW Iceland shelf	Surface	233	66.70	−24.18	6.03	35.05	1.780
B997-315 s	Djúpáll, NW Iceland shelf	Surface	217	66.73	−24.33	6.04	35.05	2.050
B997-315 s	Djúpáll, NW Iceland shelf	Surface	217	66.73	−24.33	6.04	35.05	1.963
^{o}B997-329 s	N Iceland shelf	Surface	110	65.96	−22.30	3.07	34.61	2.084
^{o}B997-329 s	N Iceland shelf	Surface	110	65.96	−22.30	3.07	34.61	1.964
B997-331 s	N Iceland shelf	Surface	44	66.07	−21.64	6.71	34.26	1.896
B997-331 s	N Iceland shelf	Surface	44	66.07	−21.64	6.71	34.26	1.958
B997-332 s	N Iceland shelf	Surface	104	66.14	−21.59	5.54	34.49	1.937
B997-332 s	N Iceland shelf	Surface	104	66.14	−21.59	5.54	34.49	1.857
B997-344 s	W Iceland shelf	Surface	282	64.84	−24.37	6.90	35.12	1.884
B997-346 s	W Iceland shelf	Surface	320	64.93	−24.26	7.12	35.11	2.202
B1191-K15	East Greenland shelf	Surface	455	68.10	−29.27	0.96	34.75	1.231
B1191-K16	East Greenland shelf	Surface	100	68.11	−29.27	0.96	n/a	1.189
^{o}JM10-181-MCG	Svalbard, Kongsfjorden	Core top, 1–2	290	78.59	11.38	0.46	34.67	1.713
JM10-182-MCG	Svalbard, Kongsfjorden	Core top, 0–0.5	287	78.98	11.51	0.35	34.71	1.265
JM10-182-MCC	Svalbard, Kongsfjorden	Core top, 0.5–1	287	78.98	11.51	0.35	34.71	1.076
JM10-182-MCC	Svalbard, Kongsfjorden	Core top, 1–2	287	78.98	11.51	0.35	34.71	1.241
JM10-183-MCD	Svalbard, Kongsfjorden	Core top, 0–0.5	271	79.02	11.73	0.01	34.67	1.357
JM10-183-MCD	Svalbard, Kongsfjorden	Core top, 1–2	271	79.02	11.73	0.01	34.67	1.317
^{o}JM10-195-MCG	Svalbard, Kongsfjorden	Core top, 0.5–1	360	79.18	11.70	1.14	34.67	1.030
JM10-195-MCG	Svalbard, Kongsfjorden	Core top, 1–2	360	79.18	11.77	1.14	34.70	1.360
HB08-BC Gr 1	W Norway, Salhusfjorden	Surface	344	79.00	5.16	8.20	34.70	1.982
HB08-BC Gr 2	W Norway, Byfjorden	Surface	314	60.48	5.25	8.20	34.70	2.151
*HM03-133-13	Faroe Island, upper slope	Surface	180	62.67	6.08	8.68	35.29	2.148
HM03-133-14	Faroe Island, shelf	Surface	115	62.50	6.08	9.50	35.28	2.349
HM03-133-14	Faroe Island, shelf	Surface	115	62.50	6.08	9.50	35.28	2.408
*HM03-133-14	Faroe Island, shelf	Surface	115	62.50	6.08	9.50	35.28	2.411
*GS04-138-16	Faroe Island, slope	Surface	416	62.80	6.06	3.39	34.96	1.718
*GS04-138-17	Faroe Island, slope	Surface	462	62.80	6.15	2.88	34.95	1.652
*GS04-138-18	Faroe Island, slope	Surface	498	62.80	6.15	1.00	34.90	1.509
*GS04-138-20	Faroe Island, slope	Surface	614	62.85	6.13	−0.14	34.89	1.558
*GS04-138-21	Faroe Island, slope	Surface	652	62.86	6.13	−0.33	34.89	1.614

Since most of the cruises were performed during summer, we regress Mg/Ca against summer temperature by default, which may make a difference at the shallowest sites.

Most of the samples (37) were analyzed at INSTAAR, following the procedure described earlier. Sample size was generally ~25–35 individuals per run, and no sample consisted of <10 individuals. The pooled standard deviation of the sample splits was 0.083 mmol mol^{-1} (6 degrees of freedom). We used Fe/Ca measurements to screen for detrital contamination (Barker et al., 2003), especially in the Icelandic shelf sites where contamination by high-Mg tephra potentially can be a problem. We rejected five samples with Fe/Ca >0.5 mmol mol^{-1}, and one sample that was an extreme Mg/Ca outlier. Four of the remaining 31 samples were excluded from the final Mg/Ca temperature calibrations because they were $>2\sigma$ outliers relative to initial regression equations (gray crosses in Fig. 4).

Seven samples were analyzed in the Cambridge University Laboratory (one previously published in Elderfield et al. (1996)) and did not include the reductive step that we use at the INSTAAR ICP-MS lab. It has been proposed that the analytical offset between the two different cleaning protocols can be approximately corrected by subtracting 0.2 mmol mol^{-1} (Elderfield et al., 2006) or 15% (Barker et al., 2003; Rosenthal et al., 2004). Because this correction introduces additional uncertainty, we did not include these samples in the calibration equations. After subtraction of 15% (gray C's in Fig. 4) they agree reasonably well with the INSTAAR data.

C. lobatulus Mg/Ca ratios correlate positively with bottom-water temperature (Fig. 4). Both a linear regression and an exponential regression fit the Mg/Ca measurements well.

Linear regression ($\pm 1\sigma$ errors): (1)

$$\mathrm{Mg/Ca} = 1.20 \pm 0.04 + 0.116 \pm 0.008\ T$$

Exponential regression ($\pm 1\sigma$ errors): (2)

$$\mathrm{Mg/Ca} = 1.24 \pm 0.04 e^{0.069 \pm 0.005\ T}$$

The linear equation has an r^2 value of 0.90 ($p<0.0001$) and the exponential equation 0.89 ($p<0.0001$). The standard error of estimate for both regression fits is 0.12 mmol mol^{-1}, equivalent to 1.0°C using the linear slope.

From a thermodynamic point of view, the exponential regression fit is expected to capture the response of the Mg/Ca ratios to temperature and has traditionally been used for calibrations of planktonic foraminifera (Rosenthal et al., 1997). It has been proposed, however, that a linear response to temperature is a

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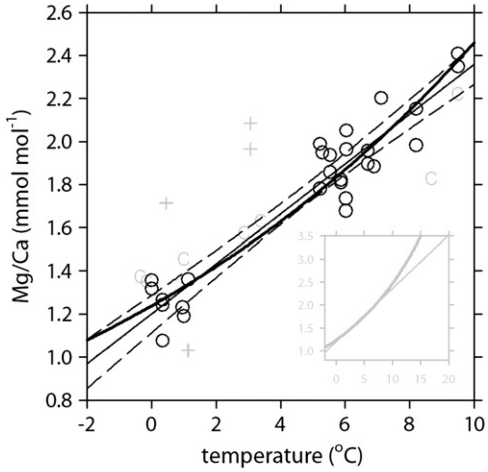

Figure 4. High-latitude North Atlantic surface sediment *C. lobatulus* Mg/Ca plotted versus CTD bottom-water temperature. Shown are the best fit linear regression (Eq. (1), fine solid line) with 95% confidence intervals (dashed lines) and best fit exponential regression (Eq. (2) thick solid line). The + symbols were rejected because they were >2σ outliers relative to initial regression equations. The C symbols indicate samples analyzed at the University of Cambridge that did not include the reductive step, and a −15% correction is applied here; since this correction introduces additional uncertainty, we did not include these samples in the calibration equations. The gray inset shows the linear and exponential extrapolations to the highest Mg/Ca values measured in MD99-2266.

better approximation in benthic foraminifera (Marchitto et al., 2007; Bryan and Marchitto, 2008). Physiological processes likely affect Mg incorporation into the shells and may mask the imprint of thermodynamics (Elderfield et al., 1996; Erez, 2003). We note that Eq. (1) is identical to the linear calibration for *Cibicidoides pachyderma* based on multicore tops collected in the Florida Straits over a temperature range of 5.8–18.6°C (Marchitto et al., 2007), and may support the assertion that neither the Florida data (Elderfield et al., 2006) nor the present data are significantly impacted by ΔCO_3^{2-}. Because this is an area of ongoing research, and because our linear and exponential fits are equally good, we apply both equations to our downcore reconstruction presented below.

Multi-proxy evidence for a climate excursion in MD99-2266 at 8200 cal yr BP

All of our proxies show a prominent excursion overlapping temporally with the 8.2 ka event recorded in the GISP2 (Grootes and Stuiver, 1997) and NGRIP (Rasmussen et al., 2006) $\delta^{18}O_{ice}$ records (Fig. 5). Although we cannot presently evaluate the uniqueness of this excursion in the context of the full Holocene record, we note that its timing and brief duration are fully consistent with the ice core record. The broadest excursion in MD99-2266, registered in the Mg/Ca record as a decrease of ~0.5 mmol mol^{-1}, lasted no more than ~200 yr. In the Greenland ice cores the 8.2 ka event extends from 8247 to 8086 cal yr BP (Thomas et al., 2007), having lasted 160.5±5.5 yr, with a peak cooling interval of 69±2 yr.

All other proxies registered a briefer excursion, lasting ~40 yr and falling within the Mg/Ca excursion interval.

A *C. lobatulus* $\delta^{18}O_{calcite}$ positive excursion started at ~8260 cal yr BP and ended at ~8220 cal yr BP, during which the $\delta^{18}O_{calcite}$ became enriched by ~0.4‰. The heaviest $\delta^{18}O_{calcite}$ value of the past 10,000 yr, and the largest positive $\delta^{18}O_{calcite}$ excursion of the entire record (albeit with data resolution varying through the core) coincides with the Mg/Ca low at 8250 cal yr BP. Enrichment of $\delta^{18}O_{calcite}$ can be caused by colder temperatures, heavier $\delta^{18}O_{sw}$ values, or a combination of both.

To test if the foraminiferal assemblages register a change coinciding with the excursion registered in the Mg/Ca and $\delta^{18}O_{calcite}$, we selected samples from before, during, and after the excursion. The percentages of *C. reniforme* spiked from a mean of 4% to a high of 11% between ~8240 and 8235 cal yr BP; the percentages of *E. excavatum* f. *clavata* spiked from a mean of 2% to a high of 15% at ~8240 cal yr BP. Both are Arctic species, found today in Polar waters on the Icelandic shelf and in fjord environments with cold bottom-water temperatures (Jennings et al., 2004). We therefore interpret the sudden increases of *C. reniforme* and *E. excavatum* f. *clavata* (albeit limited to one sample) as indicative of colder bottom waters.

An abrupt interval of low percent $CaCO_3$ started at ~8260 cal yr BP with a drop from ~18% to ~10% and ended at ~8220 cal yr BP, thus matching precisely the interval of the heavy $\delta^{18}O_{calcite}$ excursion. We interpret a drop in $CaCO_3$ as a decrease in bioproductivity. This may be consistent with a stronger halocline due to the outburst flood, since stratified waters become nutrient depleted quickly. Dilution by an abrupt increase of terrigenous (basaltic) material to the site is more difficult to rule out.

Temperature reconstruction

Converting the Mg/Ca ratios into temperature posed a challenge, because most of the measured Mg/Ca ratios from the early Holocene lie outside the modern calibration range. Our temperature calibration is limited to the modern distribution of *C. lobatulus* in the high latitude North Atlantic, where bottom temperatures do not exceed ~10°C. However, the early Holocene is generally regarded as a warm period in the high latitude North Atlantic (e.g., Eiriksson et al., 2000; Jennings et al., 2000; Justwan et al., 2008; Knudsen et al., 2008). *C. lobatulus* was apparently not displaced by this warmth, and indeed this species is known from as far south as the Georgia continental shelf today (Culver and Buzas, 1980). Our calibration data suggest that the relationship between *C. lobatulus* Mg/Ca and temperature can be described as linear or exponential; neither regression fit is significantly better or worse than the other, but the choice of equation becomes important when applied outside of the calibration range (Fig. 4 inset). Temperature reconstructions based on the linear regression fit not only yield higher overall temperatures but also give a greater temperature range than do the reconstructions based on the exponential regression fit (Fig. 6), and therefore show the greatest discrepancy compared to other North Atlantic records (see Discussion below).

Regardless of the temperature equation applied, during the 8.2 ka event we see a drop in Mg/Ca into the range of our calibration. Temperature dropped by ~5°C from ~13°C to 8°C (linear regression) or by ~3°C from ~11°C to 8°C (exponential regression). After recovery from the 8.2 ka event the temperature rose to an average of ~15°C (linear) or 12°C (exponential) over the subsequent 500 yr.

C. lobatulus has been suggested to calcify in the peak summer months and therefore may reflect peak summer temperatures (Scourse et al., 2004). Comparison between our temperature reconstruction and present-day summer temperatures at 50 m water depth, as measured in August 2008 and 2009 during routine hydrographic surveys (www.hafro.is), shows that the average temperature during the immediately post-8.2 ka interval in our record was warmer than today by ~7°C (linear equation) or ~4°C (exponential equation). The coldest reconstructed temperatures during the 8.2 ka event are the same as present-day summer water temperatures of ~8°C on the NW Iceland shelf.

Seawater $\delta^{18}O$ reconstruction

C. lobatulus $\delta^{18}O_{calcite}$ hypothetically could have been influenced by both cooling and freshening during the 8.2 ka event. The outburst flood likely would have cooled the Atlantic Water transported by the IC, resulting in $\delta^{18}O_{calcite}$ enrichment. However the isotopically light floodwaters would have pushed $\delta^{18}O_{calcite}$ in the opposite direction, potentially dampening the $\delta^{18}O_{calcite}$ cooling signal. To separate the relative contributions of temperature and $\delta^{18}O_{sw}$ to the $\delta^{18}O_{calcite}$ signal

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we used the best constrained benthic foraminiferal calibration available, which is based on the genera *Planulina* and *Cibicidoides* from the Bahamas (Lynch-Stieglitz et al., 1999):

$$\delta^{18}O_{calcite} - \delta^{18}O_{sw} + 0.27 = 3.38 - 0.21\ T, \tag{3}$$

where $\delta^{18}O_{calcite}$ is on the PDB scale and $\delta^{18}O_{sw}$ on the SMOW scale. The Mg/Ca-derived temperature was then used to extract the $\delta^{18}O_{sw}$ component captured in the $\delta^{18}O_{calcite}$ signal.

The lightest $\delta^{18}O_{sw}$ interval took place between ~8300 and 8100 cal yr BP, during which the $\delta^{18}O_{sw}$ values dropped by ~0.7 or 0.4‰, depending which temperature reconstruction is applied (linear or exponential, respectively) (Fig. 6). The interval of lighter $\delta^{18}O_{sw}$ values coincides with the interval of cooler temperature as would be expected if the 8.2 ka event was caused by the outburst flood of proglacial lakes Agassiz and Ojibway. In Figure 6 we show also the $\delta^{18}O_{sw}$ values corrected for global ice volume according to Fairbanks (1992).

In the modern surface ocean warmer than 5°C, a linear relationship exists between $\delta^{18}O_{sw}$ and salinity (Lynch-Stieglitz et al., 1999):

$$\delta^{18}O_{sw} = -14.34 + 0.419\ S. \tag{4}$$

Using this relationship the lighter $\delta^{18}O_{sw}$ values would suggest a drop in salinity of ~1.8 or 1.3 psu, depending on the temperature reconstruction used (linear or exponential, respectively). But likely the $\delta^{18}O_{sw}$-versus-salinity relationship near Iceland differed from today during the 8.2 ka event because of the input of isotopically light meltwater from Lakes Agassiz and Ojibway (LeGrande and Schmidt, 2008). This uncertainty precludes an accurate reconstruction of paleosalinity during the 8.2 ka event.

Discussion

Open shelf versus local Icelandic influences on MD99-2266

The high sedimentation rates at our site offer the advantage of high-resolution reconstruction of past marine environmental conditions. However, the close proximity to land potentially could overprint the regional marine signal with an influence from local fjord conditions, leading to the possibility that the cooling and freshening recorded in MD99-2266 is reflective of a local terrestrial signal and not a regional Atlantic signal.

As discussed earlier, the seafloor at site MD99-2266 today reflects open-shelf conditions with negligible impact from runoff. At the time of our study interval, the eustatic sea level was ~15–20 m lower than today (Lambeck and Chappel, 2001). Postglacial isostatic rebound of Iceland was rapid, and the extensive ~8600 cal yr BP Þjórsárhraun lava flow met the sea at ~15 m below modern sea level, indicating that uplift was completed by that date (Norddahl and Einarsson, 2001). Hence we infer that relative sea level was ~15–20 m lower than today during our study interval, though we note that this is largely offset by the sediment column being ~18 m thinner at that time. Furthermore, since the slight threshold beyond the fjord mouth is ~10 m deeper than our core site today, its sediment sequence would only need to have been ~8 m thinner in order for the core site to have remained completely unrestricted from the open ocean. In total, then, the depth of our core site was approximately the same as it is today, and bottom waters likely reflected open-shelf conditions with relatively little impact from land, assuming that runoff was not drastically higher than modern.

At present, Drangajökull, an icecap of ~145 km^2, lies within the Ìsafjarðardjúp drainage system (Fig. 2). There are no hydrographic data available from the rivers that drain Drangajökull (Andrews et al., 2008). But as indicated by the CTD data discussed earlier, modern runoff has no obvious impact on bottom waters near the mouth of Ìsafjarðardjúp. There is little known about the extent of the Drangajökull during the early Holocene, but the icecap was likely smaller than it is today if Drangajökull responded to the climatic optimum in a similar way as Langjökull in west central Iceland. Langjökull, the second largest ice cap on Iceland today, appears to have been much smaller during the early Holocene (Larsen et al., 2012). Larsen et al. (2012) showed a brief advance (~100 yr) of Langjökull coinciding with the 8.2 ka event, but argued that even with this advance it was still much smaller than during the late Holocene and presently. Their findings agree with model results by Flowers et al. (2008) that simulate ice-cap expansion of Langjökull during the 8.2 ka event but still culminate in a much smaller size than today.

A paleoreconstruction in the tributary fjord Jökulfirðir (Core MD99-2265, water depth 43 m) showed that glacial and glacial river input from Drangajökull had ended by ~9800 cal yr BP, as suggested by a decrease in sedimentation rate (Ólafsdóttir, 2010a). However the same study pointed out the onset of unstable conditions at ~8600 cal yr BP punctuated by a bottom-water cooling and organic carbon decline at ~8400 cal yr BP. This might represent an advance of Drangajökull, but we suggest that the 43 m Jökulfirðir site would have been much more sensitive to such an event than our 106 m location at the mouth of Ìsafjarðardjúp (Fig. 2).

In summary, while we cannot completely rule out a terrestrial influence on MD99-2266 during the early Holocene and the 8.2 ka event, we have no reason to believe that our core location was significantly less representative of open-shelf conditions than it is today.

Climate perturbation in Iceland around 8200 cal yr BP

Geirsdottir et al. (2009), in a review paper on Holocene and Pleistocene climate fluctuations in Iceland, showed that many records from on and around Iceland show a climate perturbation near 8200 cal yr BP. Multiple marine records from the west to north shelf show a temperature depression, documented in proxies including foraminiferal assemblages, oxygen isotopes, alkenones, and $CaCO_3$ percentages (e.g., Eiriksson et al., 2000; Andrews and Giraudeau, 2003; Castaneda et al., 2004). But, as Geirsdottir et al. (2009) point out, many of these climate and environmental perturbations do not exactly coincide with the timing of the 8.2 ka event in the Greenland ice core record. Rather they span a longer period, starting by at least 8500 and lasting until ~8000 cal yr BP. Several lacustrine records from Iceland (e.g., Caseldine et al., 2006) likewise show a broad cooling between ~8500 and 8000 cal yr BP.

The cooling of the Irminger Current around 8200 cal yr BP was noted in a comparison study of a west Iceland site (246 m water depth from the Jökuldjúp trough) and a northwestern Iceland site (235 m water depth in the Djúpáll trough), the latter being close to MD99-2266 (Ólafsdóttir et al., 2010b). At the more southern location, transfer functions of benthic foraminifera indicate that the cooling started around ~8740 cal yr BP and reached its lowest temperature around 8200 cal yr BP. The same broad cooling interval is less pronounced at the northwestern site, with the bottom temperatures reaching a low of

Figure 5. Summary plot for MD99-2266 results. The upper panel shows the Holocene sections of the NGRIP (black, 20-yr mean: Rasmussen et al., 2006; Vinther et al., 2006) and GISP2 (gray, 2-m mean: Grootes and Stuiver, 1997) $\delta^{18}O_{ice}$ records, and the MD99-2266 *C. lobatulus* $\delta^{18}O_{calcite}$ record. The ages of the GICC05 chronology used for NGRIP were modified to a reference point of AD 1950. The black line in the MD99-2266 $\delta^{18}O_{calcite}$ record shows the published record from Quillmann et al. (2010), with ~50-yr resolution between 300 and 2700 cal yr BP and ~100-yr resolution between 2700 and 10,500 cal yr BP. The gray data show the samples that were added between 7600 and 8400 cal yr BP to improve the resolution to ~18 yr (this study). The lower panels zoom in on the interval from 7600 to 8400 cal yr BP, showing (a) NGRIP (black) and GISP2 (gray) $\delta^{18}O_{ice}$, (b) *C. lobatulus* $\delta^{18}O_{calcite}$, (c) *C. lobatulus* Mg/Ca, (d) percentages of Arctic benthic foraminifers *C. reniforme* (dashed line) and *E. excavatum* f. *clavata* (solid line) (note the reversed axis), and (e) percent $CaCO_3$. The medium gray vertical band highlights the low Mg/Ca ratio interval; the dark gray vertical band highlights the briefer interval of excursions in the other proxies. For both panels: arrowheads indicate radiocarbon dates in MD99-2266. Estimated age uncertainty in the 7600–8400 cal yr BP interval is shown by the horizontal bar.

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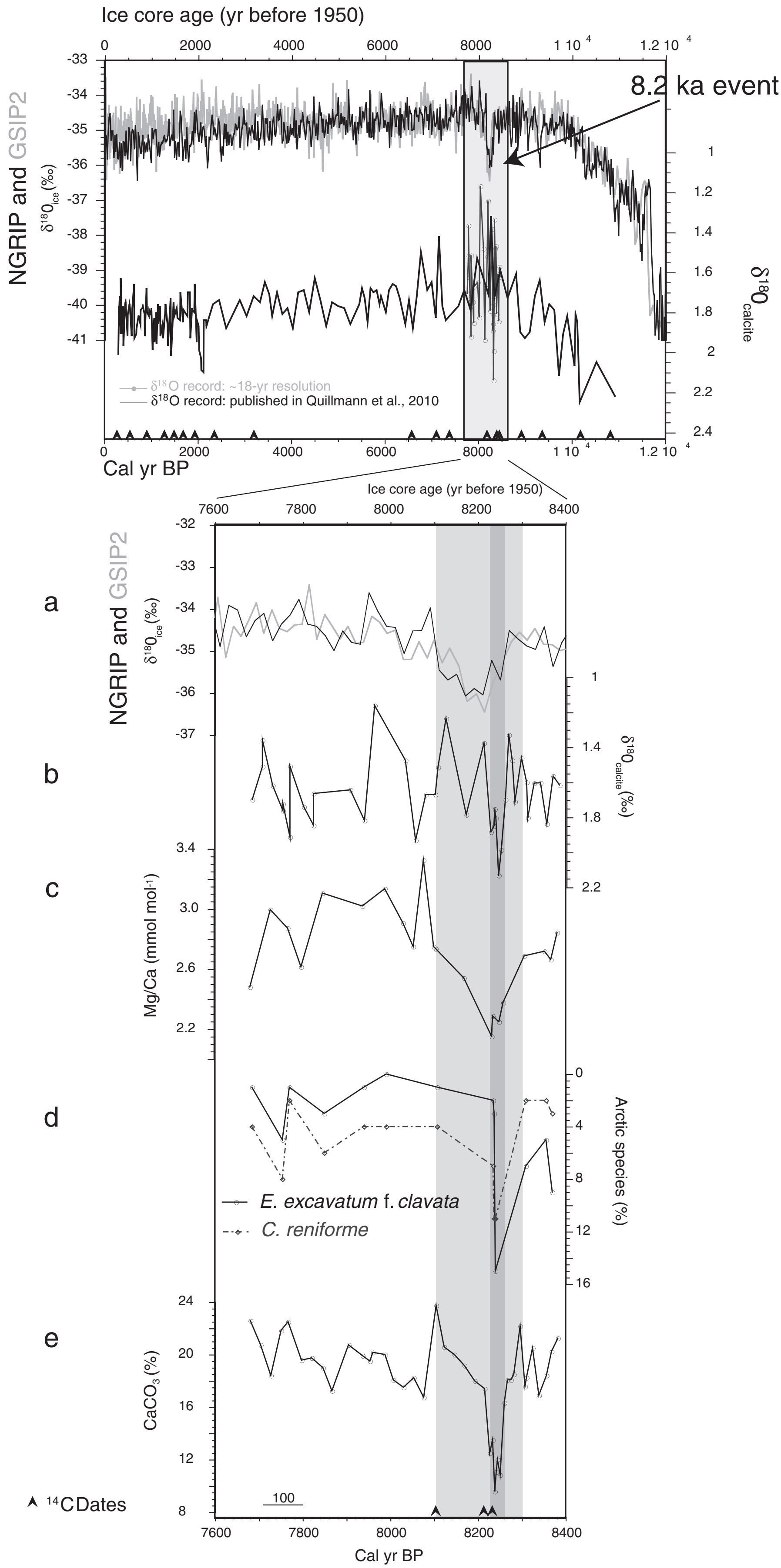
Ice core age (yr before 1950)
8.2 ka event
NGRIP and GSIP2
δ18O$_{ice}$ (‰)
δ18O$_{calcite}$
δ18O record: ~18-yr resolution
δ18O record: published in Quillmann et al., 2010
Cal yr BP
a
b
c
d
e
Mg/Ca (mmol mol-1)
Arctic species (%)
E. excavatum f. clavata
C. reniforme
CaCO3 (%)
14C Dates
100

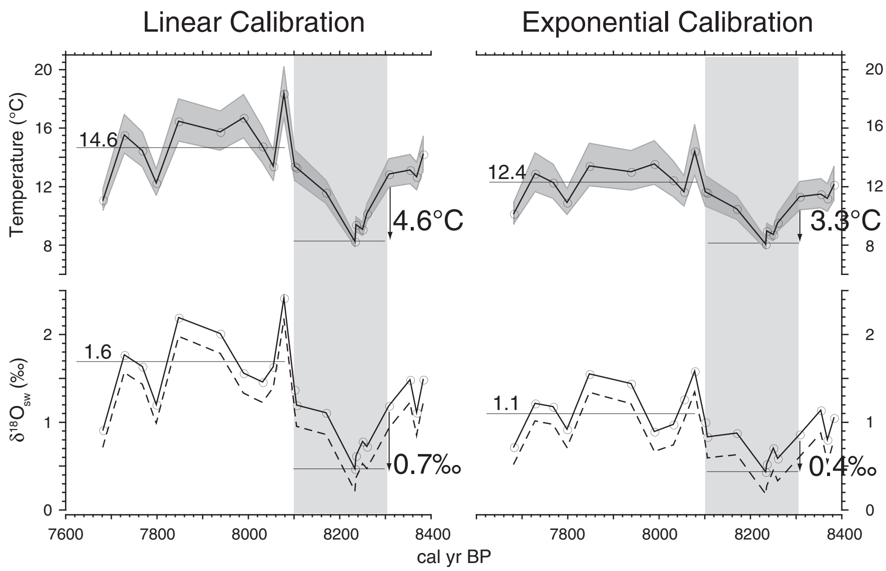

Figure 6. *C. lobatulus* temperature and $\delta^{18}O_{sw}$ reconstructions, based on the linear (left panels) and exponential (right panels) Mg/Ca temperature calibrations. The light gray vertical band highlights the 8.2 ka event interval, spanning from ~8300 to 8100 cal yr BP. The dark gray areas indicate the calculated $\pm 2\sigma$ temperature uncertainty based on the linear regression fit. The dashed lines in the lower two panels show the $\delta^{18}O_{sw}$ corrected for global ice volume (Fairbanks, 1992).

6.4°C at ~8200 cal yr BP. The temperature reconstruction by Ólafsdóttir et al. (2010b) is overall lower by several °C than our reconstruction, presumably because their site is deeper, and because they base their reconstructions on benthic foraminiferal transfer functions. As Ólafsdóttir et al. (2010b) pointed out, benthic assemblages reflect mean annual temperatures, whereas *C. lobatulus* has been suggested to calcify during peak summer months (Scourse et al., 2004) and we have calibrated Mg/Ca against summer temperatures.

In summary, multiple records from Iceland and the Iceland shelf show an excursion around 8200 cal yr BP, but it seems that the 8.2 ka event is embedded within a longer lasting climate perturbation. While there is a general consensus that the cooling at 8200 cal yr BP as recorded in Greenland ice cores was likely associated with a freshwater forcing, Geirsdottir et al. (2009) pointed out that causes for the longer period of climate perturbations in Iceland have not been resolved.

The 8.2 ka event as a regional signal dispersed in the North Atlantic

Although a climate anomaly has been tracked in a number of marine records from the North Atlantic region, the timing and duration of the observed anomalies are inconsistent (Rohling and Palike, 2005 and references therein). These inconsistencies have been attributed, in part, to imprecise chronologies and to low resolution in marine sediment cores relative to the well-dated, highly resolved Greenland ice core records (Rohling and Palike, 2005; Thomas et al., 2007).

Several regional near-surface records contain sufficient temporal resolution to potentially detect the 8.2 ka event. We can compare planktonic reconstructions with our benthic temperature reconstruction because our site is shallow (106 m today), so *C. lobatulus* records the same water conditions as many planktonic foraminiferal species that are commonly used in the subpolar North Atlantic to reconstruct surface or near-surface water temperatures. A temporal expression of the 8.2 ka event that is comparable to that in MD99-2266 can be seen in high-resolution core MD99-2322 on the east Greenland shelf, which is also affected by a subsurface flow of Atlantic Water from the IC beneath Polar surface waters (Jennings et al., 2011) (Fig. 7a). Jennings et al. (2011) attributed the anomaly to a cooling in the Atlantic Water rather than an increased influence of Polar Water to their site. The summer cooling at this site is estimated to be 1.8°C based on planktonic foraminiferal assemblages, compared to the ~3–5°C cooling recorded in MD99-2266 (Fig. 7b). Although regressed against summer sea-surface temperature, planktonic foraminiferal transfer functions integrate the signal of various growth seasons over the entire upper water column in which the various species calcified, from the mixed layer to below the thermocline (Andersson et al., 2010), potentially leading to a smaller 8.2 ka event excursion than the reconstruction based on *C. lobatulus* Mg/Ca at our site.

Using high-resolution core MD99-2251 from the Gardar Drift south of Iceland, Ellison et al. (2006) reconstructed Atlantic Water sea-surface temperatures derived from planktonic foraminiferal transfer functions, and combined them with $\delta^{18}O_{calcite}$ of the planktonic foraminifer *Globigerina bulloides* to infer $\delta^{18}O_{sw}$. They found a cooling event with a magnitude of ~1.2°C and a minor freshening centered around 8290 cal yr BP (Fig. 7c) (Ellison et al., 2006). The temperature depression is smaller than the cooling in our record, but the difference in the reconstruction methods might explain the discrepancy, as discussed above. However, Mg/Ca derived temperatures of *G. bulloides* from the same sample series (Farmer et al., 2008) did not record the same cooling interval as Ellison et al. (2006), but rather identified a series of small and rapid temperature fluctuations (Fig. 7d).

At Björn Drift (ODP 984) on the eastern flank of the Reykjanes Ridge south of Iceland, Mg/Ca of the planktonic foraminifer *Neogloboquadrina pachyderma* (dextral) (aka *Neogloboquadrina incompta* (Darling et al., 2006)) showed an ~2°C cooling (Fig. 7e) accompanied by a $\delta^{18}O_{sw}$ decrease of ~0.3‰ at ~8200 cal yr BP, but with lower temporal resolution than MD99-2266 (Came et al., 2007). A combined Mg/Ca-$\delta^{18}O$ record from the South Iceland Rise using the sub-thermocline planktonic foraminifer *Globorotalia inflata* showed a cooling of ~1°C and a $\delta^{18}O_{sw}$ drop of ~0.5‰ at ~8200 cal yr BP (Thornalley et al., 2009) (Fig. 7g). As was the case on Gardar Drift (Farmer et al., 2008), South Iceland Rise *G. bulloides* Mg/Ca did not record a cooling at this time (Thornalley et al., 2009).

Our record, combined with other high-resolution reconstructions discussed above, shows that the 8.2 ka event was associated with both a cooling and a freshening, and that the signal is seen far from the presumed source. The two most plausible scenarios are that the freshwater pulse from the proglacial lakes Agassiz and Ojibway entered the North Atlantic via the Hudson Strait and was entrained into the subpolar gyre where it cooled and freshened the Atlantic Water and the IC; and/or that the freshwater pulse dampened the formation of NADW and thus lessened the northward transport of warm, saline Atlantic Water to the high-latitude regions.

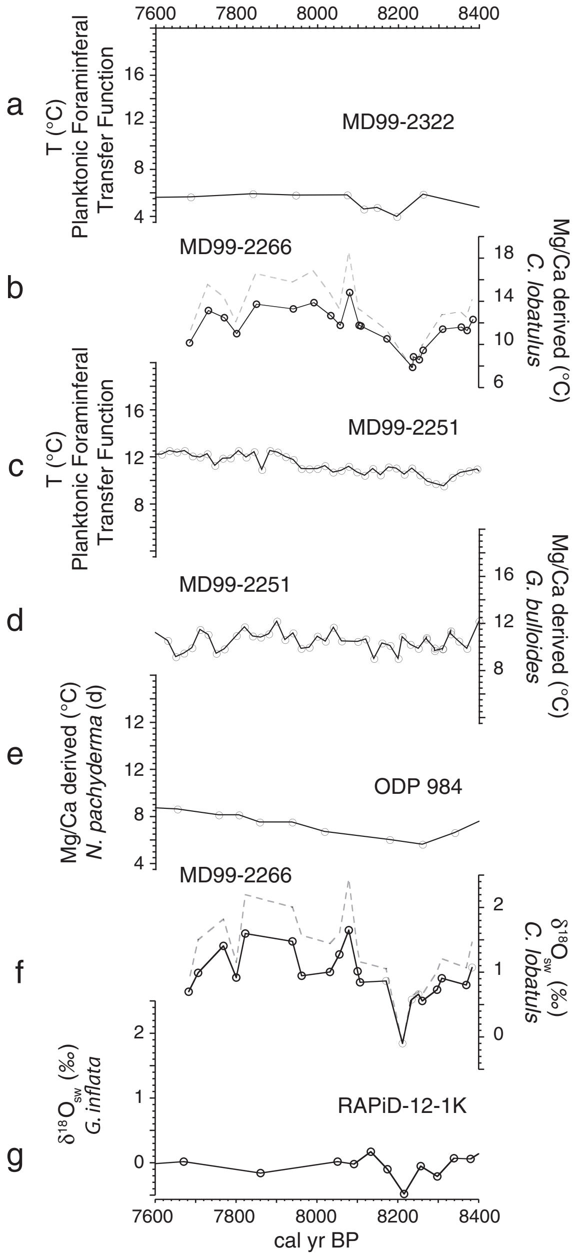

Figure 7. Comparison of the temperature and $\delta^{18}O_{sw}$ reconstructions in MD99-2266 to other regional reconstructions, plotted on the same scale. Temperature reconstructions are based on: (a) planktonic foraminiferal assemblages in MD99-2322 from the Denmark Strait (Jennings et al., 2011); (b) this study, using both the linear (dashed) and exponential (solid) Mg/Ca calibrations; (c) planktonic foraminiferal assemblages in MD99-2251 from Gardar Drift south of Iceland (Ellison et al., 2006); (d) planktonic foraminiferal (*Globigerina bulloides*) Mg/Ca, on the same core (Farmer et al., 2008) as in (c); and (e) planktonic foraminiferal *Neogloboquadrina pachyderma* (dextral) Mg/Ca in ODP984 on Björn Drift south of Iceland (Came et al., 2007). $\delta^{18}O_{sw}$ reconstructions are based on: (f) this study, using both the linear (dashed) and exponential (solid) Mg/Ca calibrations; and (g) *Globorotalia inflata* Mg/Ca and $\delta^{18}O$ in RAPiD-12-1K from the South Iceland Rise (Thornalley et al., 2009).

Clarke et al. (2009) modeled the lakes Agassiz and Ojibway outburst flood and concluded that it could have caused the cooling recorded in the Greenland ice cores by promoting sea ice growth and altering atmospheric circulation without having altered the Atlantic Meridional Overturning Circulation (AMOC) substantially. In this modeling study the sea-surface temperature near our study site dropped by ~4–6°C during the first several decades of the 8.2 ka event, compared to our observed cooling of ~3–5°C. Furthermore, Clarke et al. (2009) suggest an ~3.5–10 psu drop in salinity around Iceland, persisting for about 50 model years, which is considerably larger than our drop in $\delta^{18}O_{sw}$ suggests.

There is, however, convincing proxy evidence that changes in deep-water circulation did occur during the 8.2 ka event. Sortable silt grain size decreased around 8.2 ka both in core MD99-2251 (Ellison et al., 2006) and ODP 984 (Praetorius et al., 2008), suggesting reduced flow speed of NADW. Kleiven et al. (2008) also inferred reduced NADW flow speed at site MD03-2665 south of Greenland on the basis of magnetic sediment grains, accompanied by a dramatic decrease in benthic $\delta^{13}C$.

Our observations challenge the results from a recent modeling study by Condron and Winsor (2011). They showed in a high-resolution global, ocean-ice circulation model that the freshwater from the outburst flood was likely confined in a narrow, buoyant coastal current, flowing southward along the continental shelf to the subtropical North Atlantic. According to the results of their modeling study, the freshwater from the outburst flood did not spread to the subpolar North Atlantic and thus would not have reached the Iceland area nor the sites where deep water is formed. Our observations instead suggest that the freshwater was entrained into the subpolar gyre and spread toward Iceland, either via the Irminger Current or via an expanded gyre.

Conclusions

We aimed to determine if the brief 8.2 ka event as recorded in Greenland ice cores left a resolvable imprint in the marine realm, and if this imprint is consistent with the hypothesized catastrophic outburst flood of the proglacial lakes Oijbway and Agassiz. The high sedimentation rates at our site allowed for an ~18-yr sampling resolution, permitting us to assess the evolution of the 8.2 ka event at the mouth of Ìsafjarðardjúp in NW Iceland. We focused on the interval between 8400 and 7600 cal yr BP.

Our approach was novel as it is the first high-resolution record that reconstructs near-surface temperatures and $\delta^{18}O_{sw}$ by paired measurements of Mg/Ca and $\delta^{18}O_{calcite}$ of a benthic foraminifer. We developed a new Mg/Ca temperature calibration for *C. lobatulus* based on 27 surface sediment and core-top samples from the North Atlantic. We showed a cooling and freshening at 8.2 ka in accordance with the hypothesis that proglacial lakes drained into the North Atlantic via the Hudson Strait. We recorded the signal far from the source, implying a widespread impact on North Atlantic sea-surface conditions. Such a large-scale influence could hypothetically have cooled the region on its own, with a resulting reduction in the AMOC amplifying the effect.

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