Clim. Past, 10, 325–343, 2014
www.clim-past.net/10/325/2014/
doi:10.5194/cp-10-325-2014

Climate of the Past
Open Access

Multidecadal to millennial marine climate oscillations across the Denmark Strait (∼ 66° N) over the last 2000 cal yr BP

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Received: 13 June 2013 – Published in Clim. Past Discuss.: 11 July 2013
Revised: 15 December 2013 – Accepted: 17 December 2013 – Published: 14 February 2014

Abstract. In the area of Denmark Strait (∼ 66° N), the two modes of the North Atlantic Oscillation (NAO) and Arctic Oscillation (AO) are expressed in changes of the northward flux of Atlantic water and the southward advection of polar water in the East Iceland current. Proxies from marine cores along an environmental gradient from extensive to little or no drift ice, capture low frequency variations over the last 2000 cal yr BP. Key proxies are the weight% of calcite, a measure of surface water stratification and nutrient supply, the weight% of quartz, a measure of drift ice transport, and grain size. Records from Nansen and Kangerlussuaq fjords show variable ice-rafted debris (IRD) records but have distinct mineralogy associated with differences in the fjord catchment bedrock. A comparison between cores on either side of the Denmark Strait (MD99-2322 and MD99-2269) show a remarkable millennial-scale similarity in the trends of the weight% of calcite with a trough reached during the Little Ice Age. However, the quartz records from these two sites are quite different. The calcite records from the Denmark Strait parallel the 2000 yr Arctic summer-temperature reconstructions; analysis of the detrended calcite and quartz data reveal significant multi-decadal–century periodicities superimposed on a major environmental shift occurring ca. 1450 AD.

1 Introduction

The region of Denmark Strait (Figs. 1 and 2) is the northern end member for calculation of the North Atlantic Oscillation (NAO) index (Dawson et al., 2002; Hurrell et al., 2003; Kwok and Rothrock, 1999; Zhang et al., 2004) and is effected by variations in sea ice and freshwater, exported through Fram Strait which is associated with the Arctic Oscillation (AO) (Thompson and Wallace, 1998; Wang and Ikeda, 2000; Darby et al., 2012). Our records in this paper cannot detect multi-year oscillations, but they can resolve multidecadal to -millennial periodicities. Denmark Strait (Fig. 1) is one of the main gateways linking the polar and temperate realms. Strong thermal and salinity gradients exist along a series of marine fronts (Belkin et al., 2009) as northward advected Atlantic water comes in contact with southward-flowing polar and Arctic waters. Drift ice (Koch, 1945), in the form of sea ice and icebergs, is a pervasive feature of the East Greenland margin with landfast sea-ice retaining icebergs within the fjords for months or years (Dwyer, 1995). In contrast, the northern shelf of Iceland varies greatly in ice coverage from little or no drift ice to extensive coverage, and which has serious effects on farming and the fishery (Divine and Dick, 2006; Ogilvie et al., 2000; Ogilvie, 1997). In the 1870s for example the whaler, Captain David Gray, showed (1881) Iceland nearly encircled by drift ice. Blindheim et al. (2001) provided an important framework for paleoceanographic studies in the area of Denmark Strait (Fig. 2) by sketching out the major changes in surface oceanography that were associated with NAO positive/negative atmospheric modes (Hurrell et al., 2003; van Loon and Rogers, 1978) (Fig. 2c). This has also been extended to include the impact on intermediate and deep water circulation in the area of the subpolar gyre, with its center south of Denmark Strait (Sarafanov, 2009), and to the extent of sea ice in the region (Parkinson, 2000; Zhang et al., 2004). Yearly indexes (1899–2000 AD) for the NAO and AO are positively correlated but the explained variance is only modest ($r^2 = 0.3$) (Fig. 2d) (CRU, 2004; JISAO, 2004). However, the plot indicates that 70 % of the time the sign of the indexes match indicating that although the strength of the departures may vary the two

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oscillations are largely in-phase. However, the NAO and AO are a multi-year oscillations (Fig. 2c) whereas in most paleoceanographic records the sampling resolution is decadal at best, hence we should talk about NAO/AO-like intervals. Several recent papers have focused on conditions along the Iceland shelf over the last 1000–2000 cal yr BP using a variety of proxies (Andrews et al., 2009a; Axford et al., 2011; Jennings et al., 2001; Knudsen et al., 2004; Masse et al., 2008; Sicre et al., 2008), but there has been little effort to compare the proxies across the Denmark Strait (Jennings et al., 2002a; Andresen et al., 2013). The goal of this paper is to present proxy data from four areas, namely East Greenland fjord mouths, East Greenland mid-shelf, NW/N Iceland, and SW Iceland (Fig. 1), and assess whether there is any evidence for significant oscillations in our marine proxies over the last 2000 cal yr BP. In particular, changes in the strength and extent of the Irminger current (IC) and the East Iceland current (EIC) (Fig. 2) modulate changes in nutrient availability and in the presence/absence of drift ice. SW Iceland is always under the influence of the IC and is only very rarely affected by sea ice. SW Iceland shelf sediments contain little or no quartz, no presence of the sea-ice biomarker IP25 (Belt et al., 2007), and have high carbonate content (Axford et al., 2011). The opposite end member is represented by the inner NE Greenland shelf, which is always influenced by the East Greenland current (EGC) and extensive landfast and drifting sea ice (Reeh, 2004). During the Little Ice Age (LIA), the northeast Greenland fjords may have been covered by permanent landfast sea ice (Funder et al., 2011; Reeh et al., 2001), thus restricting both sea-ice and iceberg drift and probably curtailing the bloom of sea-ice diatoms (Muller et al., 2011). In such an environment, ice-rafted debris (IRD) is restricted, laminated sediments replace bioturbated muds (Dowdeswell et al., 2000; Jennings and Weiner, 1996), and marine carbonate production would be extremely limited. The IC is present within the deep cross-shelf troughs as an intermediate water mass, which under present-day conditions is being transported into the SE Greenland fjords causing retreat of tidewater glaciers (Andresen et al., 2012; Syvitski et al., 1996; Straneo et al., 2010). Hence the most sensitive area to NAO/AO-like variability is represented by the outer E Greenland shelf and NW/N Iceland (Fig. 3) (Jennings et al., 2011; Olafsdottir et al., 2010; Andresen et al., 2013), where drift ice can impinge on the shelf during NAO positive circulation (Fig. 2c), and thus variable values in IRD and IP25 proxies can be expected (Andrews et al., 2009a).

2 The oceanographic, climatic, and physical background

The present-day velocity of bottom currents through the Denmark Strait is high (Jochumsen et al., 2012) and little sediment accumulation takes place. Core-top dates are no younger than 8000 cal yr BP (Andrews and Cartee-

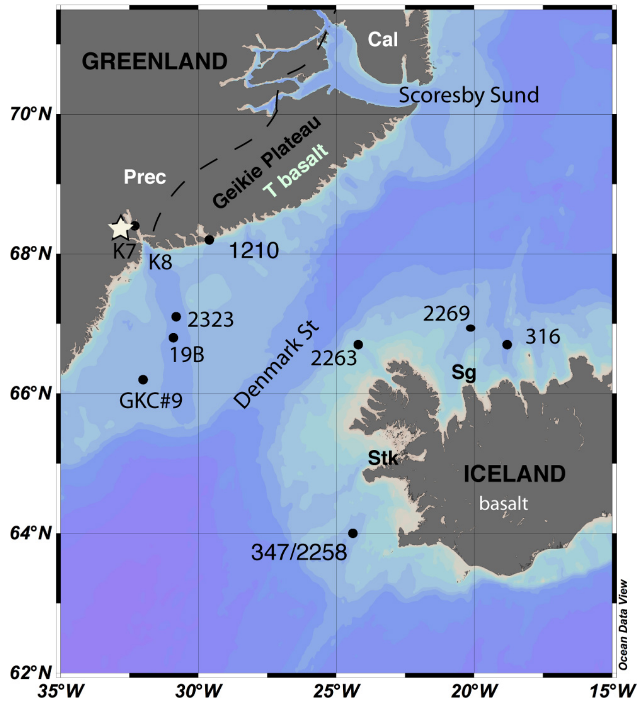

Fig. 1. Bathymetry and core locations – see Table 1 for details. The basalt outcrop of the Geikie Plateau is delimited by a dashed line. The letters Stk refer to the location of the Stykksholmur climate station (Fig. 3a) and Sg to the Siglunes transect (Fig. 3a). Core K14 (see text) is at the same location as core 1210. The large white star in Kangerlussuaq Fjord shows the location of the Tertiary felsic intrusion. North of Scoresby Sund is an extensive outcrop of Caledonide (Cal) sediments and igneous rocks (Higgins et al., 2008). Elsewhere the bedrock is dominated by Precambrian (Prec) basement.

Schoofield, 2003), thus limiting late Holocene proxy records to the adjacent shelves rather than from the Denmark Strait itself.

The East Greenland current (EGC) (Fig. 2) transports polar water, sea ice, and icebergs along the E Greenland margin (Rudels et al., 2012). The East Iceland current (EIC) branches from the main EGC and flows toward the N Iceland shelf where it comes in contact with the North Iceland Irminger current (NIIC), a northward flowing branch of the Irminger current (IC). The IC flows northward along the W Iceland shelf and a branch is directed westward where it moves landward in the Kangerlussuaq Trough as an intermediate water mass (Jennings et al., 2002a, 2011) although the main branch moves southward along the shelf break.

Instrumental and observational data are available from either side of the Denmark Strait for the last one or two centuries (Fig. 3), whereas incursions of sea ice onto the N Iceland coast have a longer record (Wallevik and Sigurjonsson, 1998) (Fig. 3d). The mean annual temperature (MAT) records from opposite sides of the Denmark Strait (Stykksholmur: W Iceland and Ammassalik: SE Greenland) share significant similarities with r^2 values (1895 AD to

Table 1. Core locations (see Fig. 1).

	Core type	Length (cm) 2 cal ka BP	Long.	Lat.	Water depth (m)	Pb210	SAR (mm yr^{-1})	Sampling resolution
				E Greenland				
JM96-1210	GC	240	−29.60	68.20	452	yes	1	10 yr
BS1191-K14	GC	> 162	−29.60	68.19	459	no	1.2	
BS1191-K7	GC	> 241	−32.10	68.26	862		3.2	∼ 15 yr
BS1191-K8	GC	> 234	−31.86	68.13	872		2.6	15 yr
MD99-2322	PC	230	−30.9	66.8	714	no	1.56	17 yr
HU93-019B	BC	48.5	−30.80	67.10	713	yes	1.94	5 yr
PO175GKC#9	GKC	20	−32.00	66.20	313	yes	0.63	16 yr
				NW/N ICELAND				
B997-316	PC and SGC	269 and 20	−18.80	66.70	658	no	2.47	20 yr
MD99-2269	PC	410	−20.85	66.63	365	no	2.03	20 yr
MD99-2263	BC	45.5	−24.20	66.70	235	yes		
				SW ICELAND				
B997-347	PC	34	−24.2	63.2	321	no	0.17	100 yr
MD99-2258	BC	24	−24.4	64	355	yes	0.62	15 yr
BC	box core							
GC and GKC	gravity core							
PC	piston core							
SGC	short gravity core							

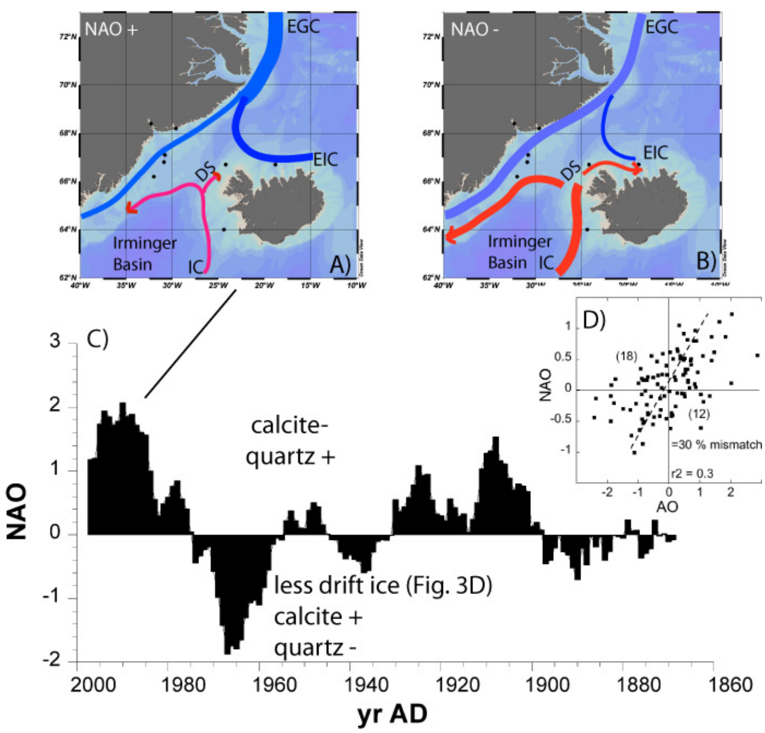

Fig. 2. EGC = East Greenland current; EIC = East Iceland current; IC = Irminger current. Core sites (Fig. 1) are shown as small dots. **(A)** Schematic plot of surface currents in the area of Denmark Strait (DS) under the influence of a NAO positive regime, and **(B)** NAO negative regime (after Blindheim et al., 2001). **(C)** Variations in the NAO winter index – data smoothed with a 10 yr running mean. **(D)** Plot of the annual indexes for the AO and NAO (1899–2000 AD).

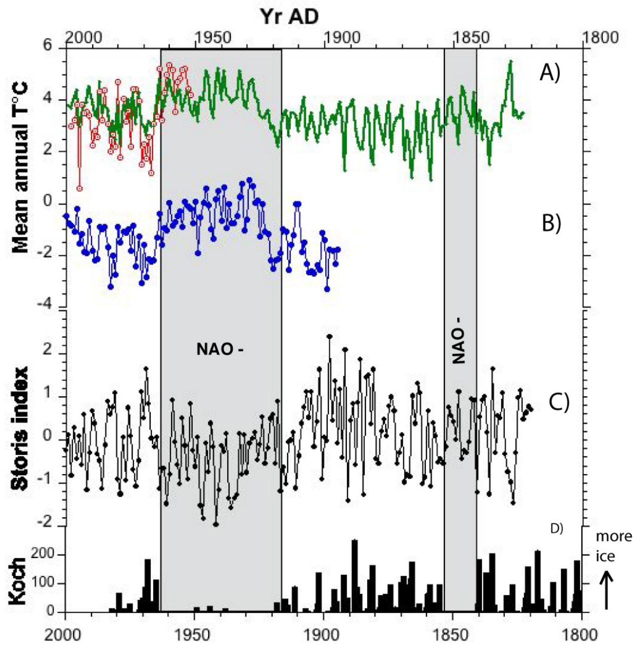

Fig. 3. Historical data from the area of Denmark Strait: **(A)** mean annual temperature data from W Iceland (green), and the Siglunes (red) marine record (Fig. 1); **(B)** MAT (blue) from SE Greenland; **(C)** index of storis (Schmith and Hanssen, 2003); **(D)** revised Koch sea-ice index. Grey shaded areas represent intervals of reduced drift ice during inferred NAO negative circulation (Fig. 2).

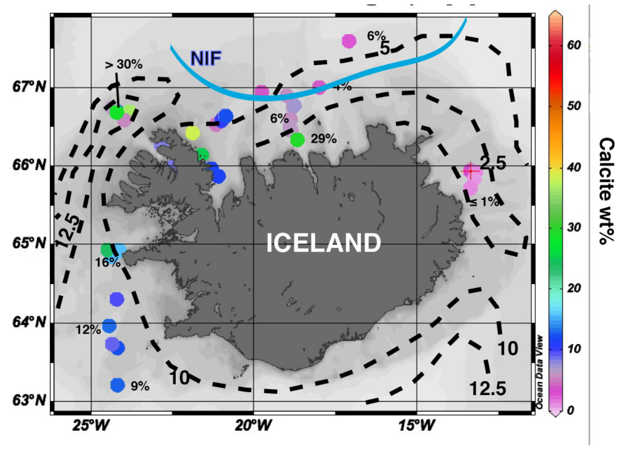

Fig. 4. Total carbonate weight% in the > 2 mm seafloor sediments around Iceland (colored circles and black numbers) (i.e. 12 %), and the index of vertical mixing dashed black lines show the (Stefansson and Olafsson, 1991) (units see text) means for 20 May–15 June, 1972–1984. The heavy blue line shows the average position of the North Iceland Front (NFI)(Belkin et al., 2009).

the present) of 0.65. There is a tight coupling between the land and marine temperatures as can be seen by the parallel trends in the Stykksholmur MAT and the average water column temperatures from the Siglunes hydrographic section (Olafsson, 1999) (Fig. 3); however, direct yearly correlations between the NAO index and the MAT data are low due to the complex underlying causes of the NAO (Zhang and Vallis, 2006). Positive values of the storis index (Schmith and Hanssen, 2003) indicate heavy ice years off SW Greenland, fed by the EGC. Zhang and Vallis (2006) noted lagged correlations between low-frequency data of Great Salinity Anomalies (GSA) (Belkin et al., 1998; Dickson et al., 1988), Iceland sea-ice extent, Labrador Sea sea surface temperature (SST) anomalies, and the NAO index. Indeed they noted (Zhang and Vallis, 2006, p. 476) that over the last century positive Iceland sea-ice anomalies lead (forces?) the positive NAO by 7 yr. Jennings et al. (2002a, p. 56) also drew attention to the "NAO paradox" (Dickson et al., 2000), noting that "[...]an extreme example of the lack of correspondence between the NAO and sea-ice flux is the Great Salinity Anomaly" because this occurred during a NAO-mode. Moreover, Lamb (1979) suggested that the climate generated by the 1969 GSA could be an analog for conditions during the LIA in the North Atlantic.

Stefansson and Olafsson (1991) provided a link between Icelandic oceanography and marine environmental proxies by synthesizing nutrient data, collected on the regular hydrographic cruises, from the Iceland Marine Research Institute (Stefansson, 1968). A key argument was the importance of water stratification as a control on productivity. They computed an index of vertical mixing (IVM), the reciprocal of the water column stability, and defined IVM as: $10^{-2}[D\sigma_t/Dz]-1$. Stefansson and Olafsson (1991) noted that the IVM has statistically significant positive correlations with nutrient variables. Maps of IVM values shows a tongue of high values extending northward along the W/NW Iceland shelf corresponding to the flow of the Irminger current (Fig. 4). The contours on the IVM parallel the calcite weight% of surface sediment from the seafloor, with a tongue of increased values heading NNE along the west coast terminating in the Djupall Trough, where calcite weight% (wt%) exceeds 30 wt% (Fig. 4). This broad trend supports the hypothesis that calcite wt% is a measure of water stratification and temperature (Andrews et al., 2001; Mahadevan et al., 2012; Thórdardóttir, 1984). There is a statistically significant relationship ($n = 26$, $p = 0.0002$) between 50 m water temperature and the weight% of carbonate for all sites, but excluding Djupall, which for its temperature, has anomalously high carbonate values. The relationship indicates an increase in carbonate of 1.8 ± 0.4 % for a 1 °C increase. The temperature data specifically relate to late July 1997 (Helgadottir, 1997).

The major component of the biogenic calcite is the Coccolithophorids, which bloom along the Iceland and East Greenland margins (Balestra et al., 2004; Thórdardóttir, 1984; Dylmer et al., 2013) – an exception is NW Iceland where foraminifera are a significant component. Icebergs may also act to disturb the water column, break down stratification and increase nutrient supply (Burton et al., 2012). Increased stratification is linked to excursions of fresh polar/Arctic waters onto the Iceland shelf (Olafsson, 1999), which in turn is associated with NAO+ conditions (Fig. 2), and thus changes in the calcite wt%, or ideally the flux, might serve as an index to changes in the NAO/AO or the Atlantic Multidecadal Oscillation (AMO) (Hakkinen, 1999; Mauritzen and Hakkinen, 1997).

Changes in sea-ice extent are considered a major climatic forcing for changes in the Holocene (Jennings et al., 2002a; Smith et al., 2003) and modelling indicates that such changes might be forced by volcanism (Schleussner and Feulner, 2013; Zhong et al., 2011). Sea ice is a pervasive factor along the East Greenland margin (Melling, 2012). Sea-ice charts (www.dmi.dk/dmi/index/gronland/iskort/iskort_-_ostgronland.htm) for the last few years show a narrow band of fast ice exists from 67° N northward and extends into the fjords, whereas offshore from this ice concentrations range from $< 9/10$th to the ice marginal zone ($< 20\,\%$), which from December through the spring lies close to the shelf break (Alonso-Garcia et al., 2013). In NAO/AO + years, drift ice is frequently transported in the EIC toward Iceland; 20th Century observations and historical sources indicate significant variability on decadal timescales (Fig. 3c, d) (Polyakova et al., 2006).

Kangerlussuaq Fjord, the Geikie Plateau, and Scoresby Sund all contain fast-flowing tidewater glaciers that contribute icebergs from their tidewater margins (Dwyer, 1993; Nuttall, 1993; Seale et al., 2011). Several of the fast-flowing ice streams are buttressed by a sikussuaq, a mélange of sea ice, bergy bits, and icebergs (Mugford and Dowdeswell, 2010; Reeh, 2004; Reeh et al., 2001; Syvitski et al., 1996). Sikussuaq and landfast ice within the fjord, restrict the movement of icebergs, and consequentially impact their ability to transport and release ice-rafted debris (IRD) to the adjacent shelf. Icebergs are frequent in the area south of Scoresby Sund (Dowdeswell et al., 1992) and can cause scouring and remobilization of sediment on the sea floor (Syvitski et al., 2001). Retreat of the tidewater glaciers has been linked to the introduction of "warm" Atlantic Intermediate water into the fjords (Andresen et al., 2011; Straneo et al., 2010; Howat et al., 2008; Joughin et al., 2008; Syvitski et al., 1996). Coarse (> 2 mm) IRD decreases rapidly seaward from the E Greenland fjords following a power law (Andrews et al., 1997) and is thus only rarely noted in late Holocene Icelandic marine sediments.

The bedrock geology map (www.geus.dk) for Nansen Fjord shows Achaean gneisses and granites overlain in places by K–TK shales, siltstones and sandstones, which are capped by an extensive early Tertiary basalt outcrop forming the Geikie Plateau (Larsen et al., 1999) (Fig. 1). The bedrock geology of Kangerlussuaq Fjord is complex (Map sheet 13, 1 : 250 000 Greenland Geological Survey) and differs from the Nansen Fjord in that outcrops of KT sediments and flood basalts are rare. Archean gneisses and granites dominate the bedrock geology. A massive felsic Tertiary intrusion outcrops on the south flank of the fjord (Fig. 1). North of Scoresby Sund there is an extensive coast-parallel outcrop of Caledonian sediments and igneous rocks (Higgins et al., 2008), which have a distinctive mineral signature in their fjord and shelf sediments (Andrews et al., 2010), although this signature is severely muted south of Scoresby Sund. The bedrock of Iceland consists of late Tertiary and Quaternary basalts and other volcanic facies with minor sediment inclusions (Hardarson et al., 2008). Quartz is virtually absent in the bedrock and is transported to Iceland in drift-ice (Eiriksson et al., 2000; Moros et al., 2006).

3 A working model for detecting climate change in the Denmark Strait region

Malmberg (1969) suggested that the reduction in the salinity of surface (upper 200 m) waters during the time of the GSA would increase the stability of the water column, hence reduce convection. This process would also increase the formation of sea ice. This thesis was critically evaluated by Marsden et al. (1991) who showed that in the Greenland and Labrador Seas between 1953 and 1980, sea-ice anomalies indeed lagged salinity anomalies. We link incursions of sea ice to an increase in the transport of "foreign" minerals onto the Iceland shelf and concomitantly to a decrease in marine productivity. These changes may thus provide a proxy for the Thermohaline Circulation (THC) (Broecker, 1997). Axford et al. (2011) compiled marine and lake climate proxies from Iceland at a 100 yr sampling resolution for AD 300–1900 cal yr. Further analysis (in our paper) of 18 variables from the 5 marine cores by Principal Component Analysis (PCA) (Davis, 1986) indicated that the 1st PC axis explained 68 % of the variance, with changes in quartz, the sea-ice biomarker IP25, and δ^{18}O of foraminifera being negatively correlated with changes in calcite wt%. The scores on the 1st PC (see later) represent a 1st-order climate signal for the eastern side of the Denmark Strait. This "seesaw" of proxies fits well with the schematic picture of NAO variations depicted on Fig. 2, such that with incursions of sea ice and reduction in the IC during positive NAO/AO conditions off Iceland, we would expect an increase in quartz, an increase in the sea-ice biomarker IP25, and a reduction in calcite production.

We employ three major proxies to capture changes in marine conditions. These are (1) changes in sediment mineralogy, especially the variations in quartz weight% (wt%), which represents an external input to the Iceland bedrock signatures (Andrews et al., 2009b; Eiriksson et al., 2000; Moros et al., 2006); (2) variations in IRD, defined as clasts > 2 mm or very coarse sand; and (3) variations in the calcite content. In terms of our proxies, changes in stratification would result in changes in coccolithophorid accumulation rates associated with changes in the timing of the extent and nature of the sea-ice margin (Balestra et al., 2004; Giraudeau et al., 2004, 2010).

3.1 Cores, chronology, and methods

We present data from cores (Table 1, Fig. 1) retrieved during cruises in 1991, 1993, 1996, 1997, and 1999 (e.g. Helgadottir, 1997; Labeyrie et al., 2003). All the radiocarbon dates on the cores have been published in INSTAAR Date Lists (Dunhill et al., 2004; Quillmann et al., 2009; Smith and Licht, 2000), apart from new dates obtained on cores HU93030-19B, MD99-2322, JM96-1210GGC, the three PO175GKC cores (Table 2), and 4 dates from B997-316PC (Jonsdottir, 2001). ^{210}Pb and ^{137}Cs measurements have been carried out (Table 2) on a GKC core and several box cores (Alonso-Garcia et al., 2013; Andrews et al., 2009a; Smith et al., 2002).

An important issue in radiocarbon-based chronologies is the value(s) for the ocean reservoir correction on either side of the Denmark Strait (Eiriksson et al., 2010; Hjort, 1973; Jennings et al., 2002b, 2011). The presence of well-dated tephras in many cores enables some insights into the problem (Jennings et al., 2002b), but no corrections have been determined for NW/W Iceland over the last 2000 cal yr. Our value for ΔR of 0, and usage in radiocarbon calibration programs (Reimer et al., 2002) means that our ^{14}C age calibrated ages may be too old. OxCal was used to calibrate the radiocarbon dates (Blockley et al., 2007). Core lengths accumulated over the last 2000 cal yr, sediment accumulation rates (SAR), and sampling intervals are listed on Table 1. These data show that there is the potential to reconstruct conditions across the Denmark Strait at multidecadal to millennial resolution, although the impact of bioturbation is always an issue (Anderson, 2001). However, the high SARs and the preservation of discrete tephra events in several of the cores (e.g. Jennings et al., 2002b) suggest only a limited impact. The age scales on subsequent figures are based on interpolation of the depth–age relationships between each date in a core (Table 1), rather than fitting them to a mathematical model. As noted in Andrews et al. (1999) the errors on interpolated ages are $\geq$ the error estimates on the bounding dates (i.e. Table 2). We do this with the clear knowledge that "All age-depth models are wrong: but how badly?" (Telford et al., 2003). Thus, as we will discuss later, potential errors in depth–age models makes it uncertain as to whether there are leads and lags in responses between sites, or whether the differences are associated with imperfect age models. These issues can be mitigated by wiggle matching, based on a high density of dates (e.g. Telford et al., 2003; Sejrup et al., 2010), but are not financially practical for a large regional study such as ours.

Our proxies have been measured using methodologies that have been well documented in the literature. The mineralogy of the sediments has been estimated by quantitative X-ray diffraction analysis using ZnO as a calibration standard (Eberl, 2003; McCarty, 2002; Omotoso et al., 2006) on the < 2 mm sediment matrix (Andrews et al., 2010). We use the downcore variations in mineralogy to seek an understanding of changes in sediment provenance using an Excel macro program called SedUnMix (Andrews and Eberl, 2012). Calcite and total carbonate (as there is little/no dolomite in the samples calcite $\sim$ carbonate%s) have been measured by XRD, coulometer, or by a WHOI carbonate device – comparisons of duplicate runs indicated excellent reproducibility between the different methods. In addition to sediment samples, large IRD clasts from the JM96-1210, GKC#9, and MD99-2262 (W Iceland) cores were pulverized for qXRD in order to gain further insights into sediment source mineralogy. We have taken surface samples and cores from Kangerlussuaq Fjord in 1988, 1991, and 1993 (Smith and Andrews, 2000; Syvitski et al., 1996), and surface samples were also taken on the Sir James Clark Ross cruise JR106 in 2004. It is appropriate at this stage to note that we are often dealing with percentage data, i.e. the non-clay and clay mineral weight% (wt%) sum to 100, and this poses problems in the interpretation of changes in percentage data (Aitchison, 1986). On X-radiographs, clasts > 2 mm were counted continuously in 2 cm depth increments (Andrews et al., 1997; Grobe, 1987; Pirrung et al., 2002). If such data are not available we use the percentage of sand-size grains in fractions > 1 mm as an index of iceberg rafting (Jennings et al., 2011), as sea ice derived from the Arctic Ocean contains predominantly silt and clay-size particles (Dethleff, 2005; Dethleff and Kuhlmann, 2010).

4 Results and interpretation

We present data from our study areas beginning with the heavily glaciated fjords of E Greenland and ending with the the ice-free waters off SW Iceland – the impact of low-frequency variations will vary along such a transect. The least affected areas will be the ice-free SW Iceland shelf, whereas the impact on the numerous E Greenland tidewater glaciers will be very significant if the NAO-mode (Fig. 2) is sufficiently persistent to increase the flow of Atlantic water into the fjords (Straneo et al., 2010; Andresen et al., 2012). This would increase glacier calving and IRD deposition in the fjords and inner shelf, but the increased rate of iceberg melting would limit the area over which IRD could be delivered.

4.1 Fjords, East Greenland

4.1.1 Geikie Plateau

Core JM96-1210GGC was taken near the mouth of Nansen Fjord close to the site of BS1191-K14 (Jennings and Weiner, 1996) (Fig. 1). Local IRD basalt clasts contain some quartz from crustal contamination. The Christian IV Gletscher at the head of the fjord has an estimated calving rate of $\sim$ 2 km^3 yr^{-1} (Andrews et al., 1994). CTD casts in 1996 indicated that water temperatures below 150 m were between 1.5 and 2 °C, thus ensuring rapid melt of icebergs exiting the fjord. The depth–age model is based on 3 ^{14}C dates (Table 2) and ^{210}Pb and ^{137}Cs measurements (Smith et al., 2002); the sediment accumulation rate (SAR) averaged 0.11 cm yr^{-1}.

The average sampling interval over the last 2000 cal yr is 20 yr sample^{-1}.

The grain-size spectra (Fig. 5a), multi-modal peaks in the > 2 mm fraction, are clearly ice-rafted (Fig. 5b). Sand content is relatively uniform (Fig. 5b) but the largest fraction is medium to coarse silt reflecting glacial abrasion. The calcite wt% is at the detection limit; an interpretation is complicated by the presence of marine mudstones within the drainage (Larsen et al., 1999). To express the qXRD data in terms of provenance, the minerals have been grouped (e.g. Na and Ca-feldspars grouped into plagioclase) and smoothed with a 5-point moving average. The 2000 yr record of changes in mineralogy in part reflects changes in grain size with a fining-upward trend being mimicked by a decrease in quartz. The plagioclase and quartz percentages tend to parallel each other and both show significant oscillations. In Fig. 5d we have used SedUnMix to estimate the downcore variability of the mineral signal by comparing each sample with the uppermost 5 samples in the core (i.e. the recent sediment sources) (Andrews and Eberl, 2012). As might be expected there is not a great deal of variation (red line, Fig. 5d) and the average absolute difference in mineralogy between the observed and SedUnMix calculated wt% is generally < 1 wt% (Fig. 5d, blue line). When compared with the foraminifera-based climate divisions from the near by core BS1191-K14 (Table 1) (Jennings and Weiner, 1996) there are some associations and the Little Ice Age (LIA) appears to be represented by an increase in quartz and plagioclase but larger peaks occur between the LIA and the Medieval Warm Period (MWP). Wanner et al. (2011) suggested two cold intervals within the last 2000 cal yr at 500 and 1500 AD – the former event (the so-called Dark Ages cold event) is marked by maximum wt% values of quartz and plagioclase and relatively large deviations in mineralogy.

4.1.2 Kangerlussuaq Fjord

Relative to Nansen Fjord sediments from Kangerlussuaq Fjord are enriched in quartz and the K- and Na-feldspars, but depleted in the Ca-feldspars, pyroxene, maghemite, saponite, and Fe-chlorite. The lithofacies, grain-size and IRD (> 2 mm) records from cores within Kangerlussuaq Fjord were presented earlier (Smith, 1997; Smith and Andrews, 2000). The sediments consisted predominantly of weakly stratified fine-grained muds with IRD (Smith, 1997, p. 78). Here we reproduce the IRD records and new qXRD data for cores BS1191-K7 and K8 (Table 1, Figs. 1 and 7), which although poorly dated (Table 2) cover the last 800 yr, and include the local Little Ice Age (Geirsdottir et al., 2000). The SAR for K8 is 0.26 and 0.11 cm yr^{-1} for K7 – using ^{210}Pb, Smith et al. (2002) determined a SAR over the last 100 yr of 0.42 cm yr^{-1} at K8. Thus the interpolated age estimates for K7 and K8 have unknown errors on the age estimates but are probably in the range of ±150 yr; for this reason we do not correlate specific multidecadal events between the fjord and trough proxies. The IRD records (Fig. 6) show pulses of ice-rafting with a notable absence ∼ 1500 AD. In K7 a low persistent level of IRD is evident in the last 400 yr, whereas in K8, closer to the fjord mouth, no visible IRD > 2 unitmm was captured in the core for the last 300 yr. In K7 higher levels of quartz in the < 2 mm fraction appear to coincide with reduced IRD.

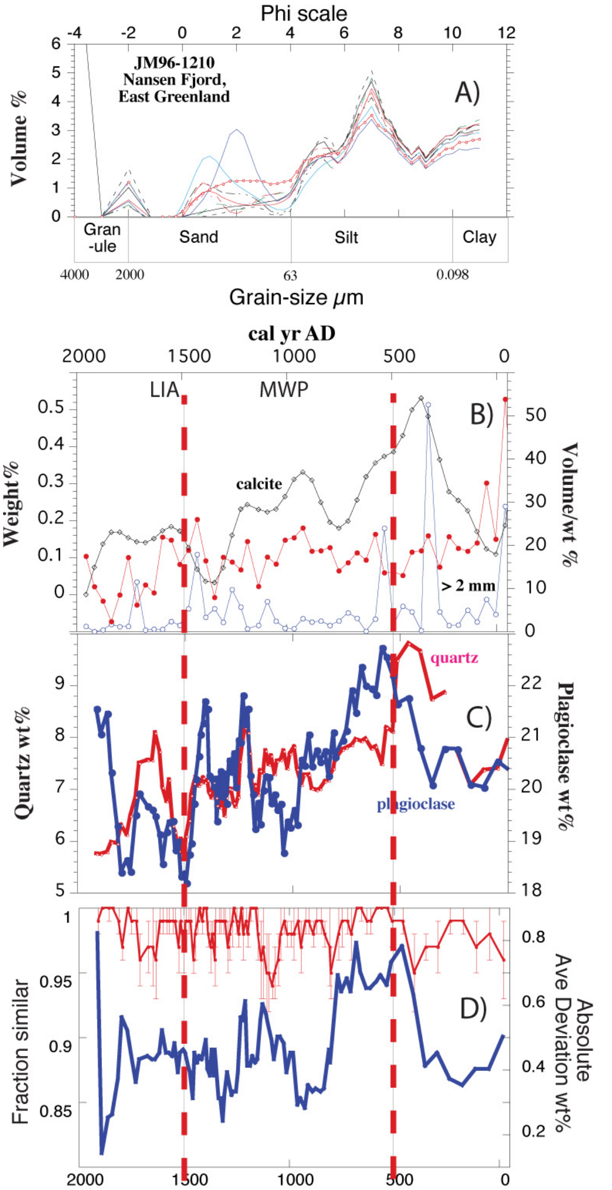

Fig. 5. Data from JM96-1210 and BS1191-K14, Nansen Fjord (Fig. 1). **(A)** Grain-size spectra from JM96-1210; **(B)** plots of the > 2 mm sediment (blue), sand (red), and wt% calcite (smoothed); **(C)** smoothed wt% quartz and plagioclase; **(D)** downcore variations and error in the estimated similarity from the uppermost 5 qXRD sample mineralogy (red) and the absolute average deviation (wt%) from the 9 mineral groups. The two vertical dashed red lines represent the timing of the latest Holocene cold events (Wanner et al., 2011).

Table 2. Radiocarbon dates.

Core ID	Depth (cm)	^{14}C date	Error	From	To	Median cal BP	Sigma
JM96-1210GGC	49	785	15	486	374	438	20
	149.5	1275	15	893	756	821	37
	332	3050	80	3051	2685	2833	94
BS1191-K14	50	855	60	521	305	436	58
	115.5	1440	70	1107	771	938	82
BS1191K7	262.5	1310	60	920	680	805	64
BS1191-K8	107.5	1155	56	768	550	664	53
	226	1391	55	1008	745	881	64
HU93030-019B	55.5	835	20	489	359	436	32
MD99-2322	2.3	675	30	359	145	273	42
	34	693	38	413	148	290	51
	101.5	1267	44	875	670	758	54
	150	1627	46	1254	1021	1137	59
	232.5	2415	20	2081	1908	1987	42
MD99-2263	7	595	15	288	145	253	37
	10	600	15	294	146	257	34
	25	850	15	514	445	482	17
	38.5	1620	15	1253	1127	1192	32
	45	2165	15	1824	1690	1759	35
MD99-2269	1	72	37	55	10	36	11
	42.5	680	30	413	257	328	41
	131	1010	30	646	530	589	31
	177.5	1226	25	858	688	762	41
	265.5	1693	42	1334	1159	1250	43
	412	2396	47	2150	1890	2028	67
B997-316	18.5	402	38	121		44	38
	41.75	755	35	471	305	396	46
	118	1020	40	660	525	595	36
	219.25	1090	90	830	505	655	81
	249.5	1550	120	1342	862	1102	121
MD99-2258	27.5	725	15	429	301	367	35
	31.25	1937	36	1587	1380	1482	52
	70.5	3180	20	3068	2877	2975	48
B997-347	3	770	65	504	285	403	61
	61	3110	45	3020	2760	2886	66

5 Kangerlussuaq Trough

Icebergs with keel depths < 400 m can exit the fjord and icebergs of this size have been observed (Dowdeswell et al., 1992) and inferred based on iceberg scours on the adjacent seafloor (Syvitski et al., 2001). High-resolution seismic surveys of the inner Kangerlussuaq Trough in 1993 and 1996 (Jennings et al., 2002b, 2006, 2011) indicates that cores HU93030-019B and MD99-2322 (henceforth 2322) (Fig. 1) are stratigraphically from the same site and are well below the limit for iceberg scouring. The ^{210}Pb profile and ^{14}C dates from 019B (Smith et al., 2002; Table 2) are conformable in contrast to the results from PO175GKC#9 (Alonso-Garcia et al., 2013), farther down the trough but in only 300 m of water, where Cold Room storage (calcite dissolution) and possible reworking led to the ^{14}C dates being rejected. The calibrated dates from the base of 19B box core and the 2322 Calypso core overlap (Table 2). The SAR for 19B is ~0.127 and 0.148 cm yr^{-1} for 2322; 19B was processed in 1 cm intervals and 2322 averaged 2 cm intervals. The average resolution is one sample every 15 yr.

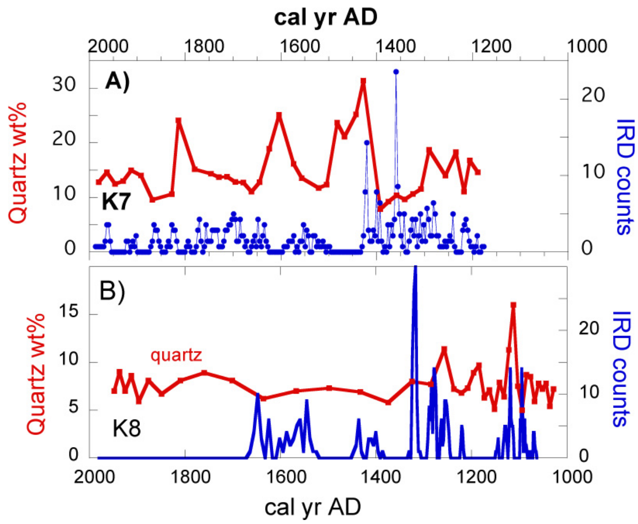

Fig. 6. IRD (blue) and quartz wt% data (red) from BS1191-K7 **(A)** and -K8, outer Kangerlussuaq Fjord **(B)** (Tables 1 and 2).

The sediment in the upper 2000 yr of 2322 is generally silty clay with scattered IRD clasts and coarse sand (Jennings et al., 2011). The mineralogy of the < 2 mm fraction was determined by qXRD and the sampling for the last 2000 cal yr was increased from an earlier study (Andrews et al., 2010). The inner Kangerlussuaq Trough is positioned to receive sediment from the Kangerlussuaq Fjord tidewater glaciers, and from the numerous tidewater glaciers that drain the Geikie Plateau (Fig. 1). A key question is whether the sediment in Kangerlussuaq Trough is being contributed from the erosion of felsic rocks (e.g. granites, gneisses, felsic intrusions – Kangerlussuaq Fjord), from erosion of the complex basalt outcrop (Geikie Plateau), or some fraction from each. A 5-point moving average has been applied to the data to enhance the signal (Burroughs, 2003).

The SedUnMix program allows for 5 samples per end member or source with a maximum of six sources; details of the program are given in Andrews and Eberl (2012). We initially evaluate downcore variability in sediment composition by comparing the downcore sediment samples with the 5 uppermost samples from 019B and computing the standard error based on 10 iterations of the SedUnMix algorithm. The results indicate that in the last ~500 yr there has been little variation in sediment provenance, although commencing ca. 1500 AD there were consistent deviations (Fig. 7a).

The next question is whether we can assign sediment sources and unmix the sediment composition. We represent the Tertiary basalt source by the qXRD mineralogy from JM96-1210 and the Kangerlussuaq Fjord source by surface samples from core BS1191-K7 (Alonso-Garcia et al., 2013; Andrews et al., 2010). However, we note that interpretation is complicated by the closed array problem (Chayes, 1971) so that as the contribution from one source increases there is a tendency ($r = -0.58$) for the other source to decrease

(Fig. 7). The average difference between the measured and calculated weight% for each of the non-clay and clay minerals used in the analyses is 3.3 ± 0.3 wt%, indicating a reasonably good fit between the measured (sources) and calculated (2322) mineral wt%s. The results (Fig. 7b) indicate that Geikie Plateau has been the dominant sediment source with an average of 45 ± 7 % of the sediment being assigned to a Nansen Fjord-like source, compared to 22 ± 11 % from Kangerlussuaq Fjord. There is thus an unaccounted source(s) averaging ∼33 % of the composition (Fig. 7c). However, our estimates also include a calculation of the standard deviations of the estimate based on iterations of the 2×5 matrix (Andrews and Eberl, 2012); thus, we could account for nearly 100 % of the sediment composition by specific sample selection from the two source areas. Additional sediment sources would certainly be derived from icebergs that originated in Scoresby Sund and northeast Greenland and sediment entrained in sea ice from the Arctic Ocean (Andrews, 2011; Bigg, 1999; Darby and Bischof, 2004; Darby et al., 2011; Seale et al., 2011). Over the 100 yr interval ending ca. 1993 AD, the relative contribution from the Geikie Plateau has increased, whereas the data indicate a decrease in the more felsic source(s) associated with Kangerlussuaq Fjord (Fig. 7b).

The last 2000 yr covers several important climate events, such as the Medieval Warm Period (MWP) and Little Ice Age (LIA) (Fig. 8) although their time boundaries are probably not globally synchronous nor agreed on (Hughes and Diaz, 1994; Broecker, 2001: Mann et al., 2009). For the purpose of this paper we plot the boundaries for the MWP between 950 and 1250 AD, and the LIA from 1250 to 1910 AD. A 15 yr equi-spaced integrated time series was generated for 2322 data using AnalySeries (Paillard et al., 1996). Figure 8a graphs the PC 1 scores from the Iceland marine data set (Axford et al., 2011) versus the calcite wt% from 2322. The 2322 calcite data parallel the lower-resolution PC scores, suggesting a broadly parallel climate evolution on both sides of the Denmark Strait, in keeping with the climate data for the last 100–150 yr (Fig. 3). We use a regime shift indicator (RSI) algorithm (Rodionov, 2004, 2006) to ascertain whether the records show marked regime shifts (Fig. 8b and c), and whether these shifts have any correspondence with the 14C probability distributions of vegetation kill/ice growth on Baffin Island, Canada (Miller et al., 2012) (Fig. 8b). Such a comparison is appropriate, as the changes in the small Baffin Island ice caps have been linked to intervals of persistent volcanism which force variations in the extent of sea ice (Miller et al., 2012; Schleussner and Feulner, 2013). The peaks in the ^{14}C probability distribution reflect intervals when vegetation was able to grow on the uplands of Baffin Island, and the absence of dates represents intervals of persistent ice/snow cover (Williams, 1978). Lowell et al. (2013) reported dates on dead vegetation with similar ages to the north of our sites (Istorvet Ice Cap ∼71° N) (Miller et al., 2013). Particularly noteworthy in our data is the calcite minimum ∼1600 AD

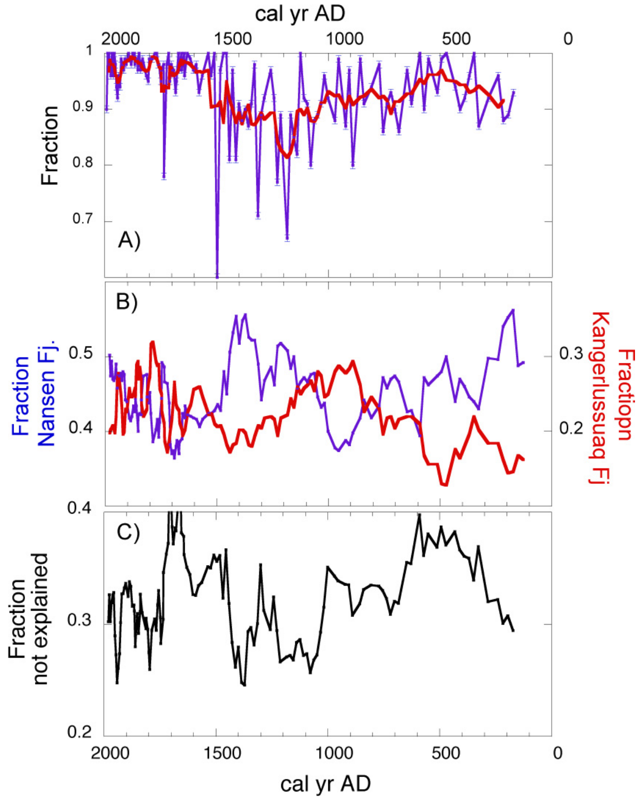

Fig. 7. (A) Changes in the fraction of sediment with a similar mineralogy to the uppermost 5 samples (last 100 yr) at site 2322 (blue) (Fig. 1) and 5 point running mean (red). **(B)** Estimated fractions of sediment to site 2322 from the Nansen Fjord (blue) or Kangerlussuaq Fjord (red) based on mineralogy. **(C)** Changes in the average fraction of sediment not attributable to either source (see Fig. 7b).

which coincides with the abrupt reduction of dates on vegetation emerging from the retreating ice patches, although lagging the vegetation reduction by ∼100 yr, whereas the older interval, which terminated ∼1000 AD, occurred ca. 150 yr prior to the regime shift in calcite (Fig. 8b).

The quartz wt% data has a trend opposite to the calcite wt% (Fig. 8c) with an increase starting ca. 800 AD and continuing until 1500 AD when there is a decrease, although there are sharp peaks, which project above the regime shift mean values (Fig. 8c).

5.1 N/NW Iceland: none to variable sea ice, variable IRD

N/NW Iceland sites (Figs. 1 and 2) should be sensitive to low-frequency multidecadal/-century changes, as indeed was demonstrated by Axford et al. (2011) and Andresen et al. (2005). X-radiographs from the numerous cores from the N Iceland shelf indicate that IRD clasts > 2 mm are extremely rare or absent, but quartz is present in the < 2 mm sediment fraction (Moros et al., 2006). The quartz and quartz + k-feldspar data from 16 cores from the Iceland shelf have

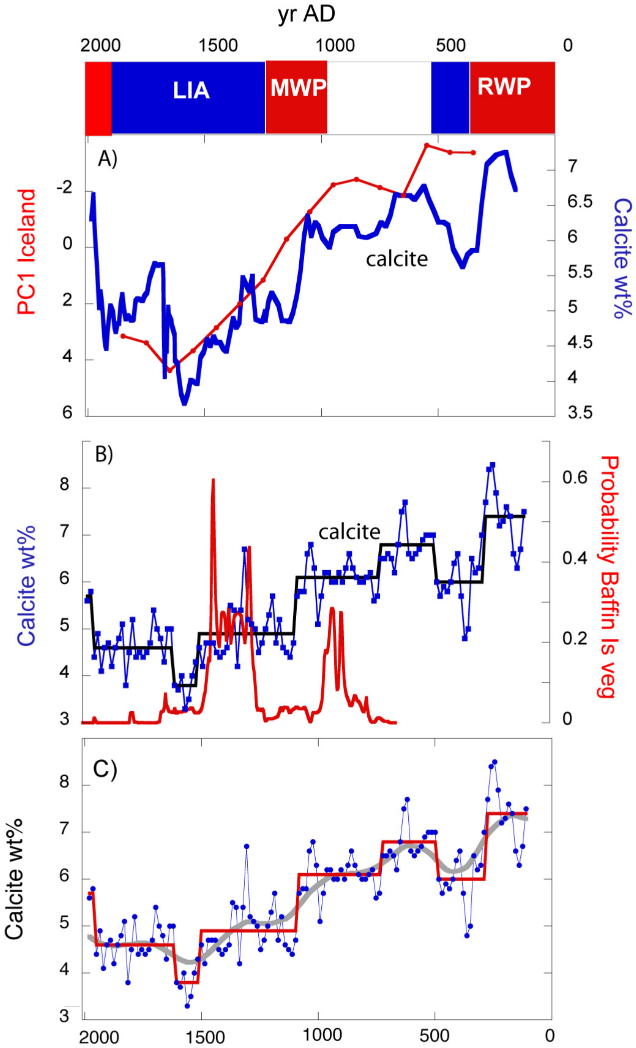

Fig. 8. (A) Climate intervals of the last 2000 yr and a graph of the 1st PC scores from the N Iceland proxy data (red) (Axford et al., 2011) versus the smoothed (5-point average) weight% calcite (blue) from the combined HU93030-019B and MD99-2322; **(B)** 15 yr calcite wt% (blue) and regime shift mean values versus the radiocarbon calibrated probability distribution for Baffin Island vegetation (Miller et al., 2012) (red); **(C)** quartz wt% (grey line with filled circles), regime shifts (black line) and the trend (heavy grey line).

been presented at 250 and 100 yr resolution (Andrews, 2009; Andrews et al., 2009b). We focus on three high-resolution records that span the last 2000 cal yr, namely MD99-2263 (Andrews et al., 2009a; Olafsdottir et al., 2010), MD99-2269 (Andrews et al., 2003; Moros et al., 2006; Stoner et al., 2007), and B997-316PC3 (Jonsdottir, 2001). Olafsdottir et al. (2010, p. 116) noted, "The characteristics for the LIA in both cores (MD99-2256, SW Iceland and MD99-2263/2264) are high amplitude fluctuations, not at all a stable continuous cold bottom water period". The three cores span an oceanographic gradient (Smith et al., 2005; Stefansson, 1962) with MD99-2263 dominated by the NIIC, whereas B997-316PC3 (Jonsdottir, 2001) occurs much further offshore at or near the North Iceland Marine Front (Fig. 4) (Belkin et al., 2009) with bottom water temperatures $< 0°$. This variation in oceanographic setting raises the question of differences in the ocean reservoir correction during the late Holocene (Eiriksson et al., 2010); however, the radiocarbon-based interpolated age at 105.5 cm in B997-316PC3 is 1433 AD, compared to an age for the Veiðovötn tephra of 1477 AD, identified at this depth (Jonsdottir, 2001; Larsen, 2005), indicates that our suggested reservoir correction is not unreasonable.

The calcite data (Fig. 9a) show large differences in wt% with very high values in MD99-2263 and lower values to the north and east, consistent with the suggestion of Stefansson and Olafsson (1991) and the distribution of calcite in Icelandic seafloor sediments (Fig. 4). There are considerable variations in wt% at each site and there is a marked decrease during the Iceland LIA (Geirsdottir et al., 2009; Grove, 2001; Ogilvie and Jónsson, 2001) and a consistent rise in values in recent decades. In detail (e.g. Fig. 9c), there is a remarkable similarity between the calcite wt% data from N Iceland (316PC3) and our data from the Greenland side of the Denmark Strait (Fig. 8) over the last 2000 yr. This will be examined more fully in Sect. 5 (below).

The quartz wt% time series (Fig. 9b) have values <wt 2 % for 2263, partly because of the very high calcite values at this site, compared to 2269 and 316PC3. The results for 2263 and 316PC show highly variable inputs over the last 1100 cal yr superimposed on a general increase toward the present day. This variability is in keeping with IRD deposition, which by its very nature is spatially and temporally "noisy".

5.2 SW Iceland: no sea ice and no IRD

The clockwise pattern of drift ice around the Iceland coast implies that the SW shelf is only occasionally impacted by drift ice (Divine and Dick, 2006), although Jennings et al. (2001) suggested that polar water may have impacted the area. Calcite is a major component of the sediment and reflects significant biological productivity (Stefansson and Olafsson, 1991). In Axford et al. (2011) it was shown that quartz and the sea-ice biomarker IP25 have values close to the detection limits.

Studies of MD99-2258 (box core) and the MD99-2259 piston core indicates a hiatus between ~1600 and 500 AD (Table 2). However, nearby core B997-347 (Fig. 1, Table 1) has an estimated age of the top-most sample ~1480 AD. A combined plot of calcite wt% from 2258 and 347 (Fig. 9d) shows sharp oscillations in calcite over the last 2000 yr. When compared with data from Kangerlussuaq Trough (Fig. 9d), the SW Iceland record leads by several decades. It is unclear whether this is an artifact of dating uncertainties or not, especially with the regional issue of the ocean reservoir correction (Sejrup et al., 2010). The quartz wt% in 347PC is close to zero, and it only rises above 1 % in 2258 on one occasion (~1790 AD) (Fig. 9e).

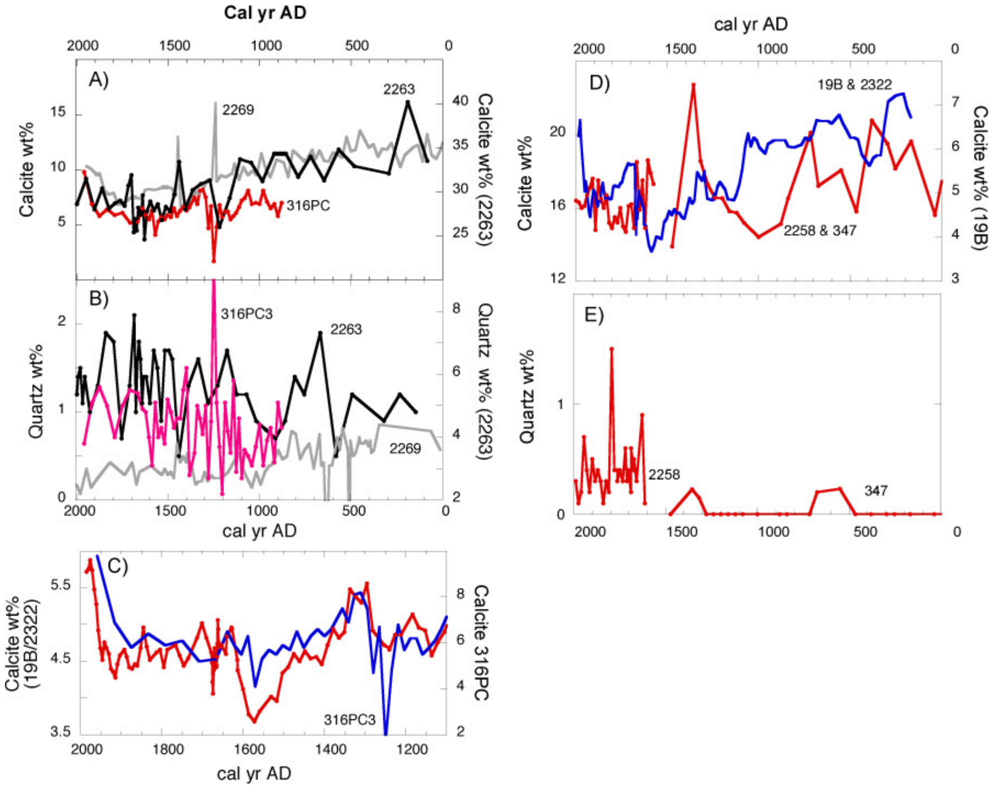

Fig. 9. (A) Calcite wt% data for sites MD99-2263 (left y axis), MD99-2269, and B997-316PC3 (right y axis) from N Iceland (Fig. 1). **(B)** Quartz wt% from same sites and same y axes. Note the difference in y axis values from the left to right axes. **(C)** Calcite wt% data from 316PC versus 19B/2322 over the last 1300 yr. **(D)** Calcite wt% data for cores from SW Iceland (Fig. 1) compared to 2322. **(E)** Quartz wt% data from SW Iceland.

6 Discussion

The common thread in our sites is provided by the variations in flow of the Irminger current (IC) and its conceptual impact on changes in calcite and quartz (Fig. 2) – the IC is a surface current off SW Iceland, a surface or intermediate current off N Iceland, and an intermediate water mass in the deep troughs and fjords of E and SE Greenland – variations in the latter are a prime cause for changes in the tidewater glacier dynamics of the area (Syvitski et al., 1996; Andresen et al., 2011). A number of Arctic/North Atlantic climate reconstructions have been developed, including a 2000 yr estimate of Arctic summer temperatures (Kaufman et al., 2009) (Fig. 11a), 1400 yr Arctic temperature (Shi et al., 2012a, b) and Arctic sea-ice records (Kinnard et al., 2011), and a 700 yr long AMO reconstruction (Mann et al., 2009; Gray et al., 2004). There is a strong association between the Arctic summer temperature time series and the 2322 calcite wt% data (Fig. 10a). On their initial chronologies the calcite and quartz data and the AMO record (Fig. 10b) are only poorly correlated. However, the fit between the AMO and calcite can easily be improved using AnalySeries (Paillard et al., 1996) (7 tie-points) to improve the association with $r = 0.76$ (Fig. 10c). The correction to the calcite time series was such that an increase in the ocean reservoir correction from 0 to ±30 yr would accommodate much of the suggested correction. Such a correction is certainly reasonable (Sejrup et al., 2010) but difficult to validate. The adjusted quartz time series had no obvious association with the AMO (not shown).

These data raise the question as to whether there are common trends and periodicities in calcite and quartz across Denmark Strait over the last 2000 cal yr. Stoner et al. (2007) constructed a composite Holocene age model from either side of the Denmark Strait (MD99-2269 and -2322, 440 km apart, Fig. 1) based on paleomagnetic secular variations (PSV), and an age control based on 47 ^{14}C dates. This PSV record has been shown to have regional chronological application (Olafsdottir et al., 2013). The composite chronology showed that the carbonate data from these two sites were remarkably similar (Stoner et al., 2007). We pursue this issue by examining these sites over the last 2000 yr. We do this by producing 30 yr equi-spaced data (Paillard et al., 1996) from 110 to 1970 AD for the two sites (Fig. 11) and 15 yr spaced data for 2322. Trends in the data were detected by using Singular Spectra Analysis (SSA) (Ghil et al., 2002), and the residuals from the trends were processed using SSA and Multi-Taper Method (MTM) analyses (Mann and Lees, 1996; Kondrashov et al., 2005). Given the warnings about age models

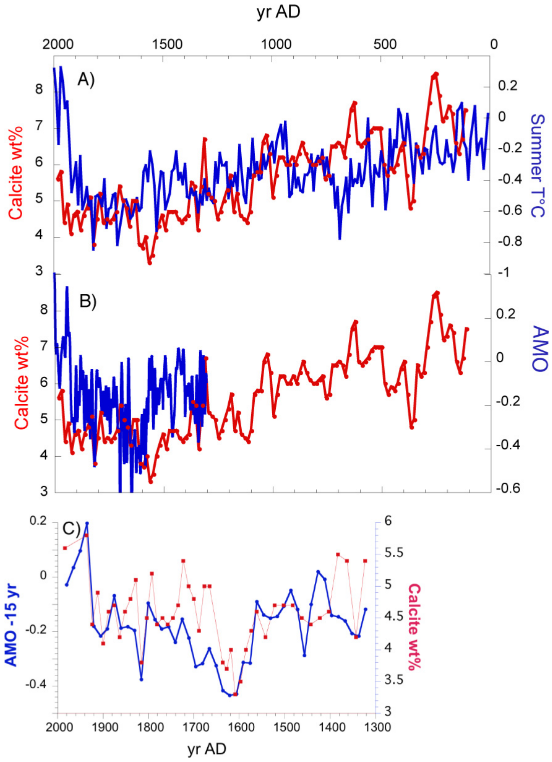

Fig. 10. (A) Plot of the calcite wt% from 19B/2322 (red) versus the summer temperature estimates (Kaufman et al., 2009) (blue). **(B)** Plot of the calcite wt% from 2322 (red) versus the AMO estimates for the last 700 yr (blue) (Mann et al., 2009). **(C)** Plot of the AMO estimates (Man et al., 2009) (blue) and corrected calcite wt% chronology (red).

(Telford et al., 2003) the 30 yr equi-spaced time series are at the limit of acceptance, although the Nyquist frequency essentially restricts the recognition of meaningful periodicities to a century or more. The 15 yr resolution series for 2322 is cautiously accepted given the high and nearly monotonic SAR but any significant period in the series will be multidecadal to century in length.

The trends in calcite from the two sites (Fig. 11a) explain 80.2 and 87.5 % of the data and are very similar to each other ($r^2 = 0.85$) on their respective time series. Both series show a progressive decrease in calcite commencing 500–600 AD, which reaches a minimum between 1550 and 1750 AD, and which is then followed by a modest increase. If this is indeed a reliable index of water stratification (Marsden al., 1991) then it suggests a persistent freshening of the surface water from Roman times on, and by inference a slow down in the THC. There is no obvious break in the data that can be associated with the MWP.

The residuals from the trend for 2322 show significant (95 %) oscillatory modes (Vautard and Ghil, 1989; Kondrashov et al., 2005) at 192 yr (175 yr MTM) and weaker modes at 110 and 76 yr. The reconstructed time series based on these modes (Fig. 11b) explains 82 % of the residual variance. A oscillatory mode (Vautard and Ghil, 1989) was also dominant on the 2269 calcite residuals (Fig. 11c), with a period of 114 yr (128 yr MTM) and weaker modes at 292 yr (370 yr MTM), and 67 yr (86 yr MTM). The reconstructed time series based on these modes explains 91 % of their variance. The oscillatory modes at these two sites, some 440 km apart (Fig. 1), are similar but not identical; however, the reconstructed time series from the two sites (Fig. 11a and b) are not correlated ($r^2 = 0.02$). The 15 yr spaced calcite and quartz data from 2322 were detrended which delimited two significant multidecadal oscillatory modes with periods of 90, and 62 yr compared to a single mode of 68 yr in quartz. The oscillations lie within range usually ascribed to the AMO.

In contrast, the trends in the quartz wt% across Denmark Strait (Fig. 11d) are very different and explain 44 % (2322) and 72 % (2269) of the variance. The 2269 trend is to a degree a mirror image of the calcite data (Fig. 11a) with quartz wt% increasing during the LIA with a low interval during the MWP. On the East Greenland shelf the quartz wt% reaches a maximum ca. 1200 AD and has a distinct low during the LIA (Fig. 11d). This may indicate that the IRD export of sediment from the quartz-rich areas on either side of the Geikie Plateau (Fig. 1) was restricted by pervasive landfast sea ice during the LIA (Reeh, 2004). Residuals from the SSA East Greenland quartz trend (Fig. 11e) had two dominant oscillatory modes with periods of 95 and 190 yr (104 and 296 yr MTM), which explained 78 % of the residual variance, whereas 50 % of the residual variance in the 2269 quartz data are associated with a mode of 155 yr (128 yr MTM). The coherence between the residual calcite and quartz records from these two sites (Fig. 1) was evaluated using MTM; this resulted in somewhat similar multicentury periods of 440 and 133 yr at 2322 compared to 416 and 181 yr for 2269.

The link between the NAO and ocean circulation and conditions in the Denmark Strait region (Figs. 2 and 3) suggests it is important to see if our records (Figs. 5–9) have any obvious association with a NAO reconstruction (Cook et al., 2002; Trouet et al., 2009; Kinnard et al., 2011). Departures from the mean for the period 1990–1100 AD were derived for the calcite and quartz wt% data for 2322 and 2269, respectively, and compared with the integrated 15 yr NAO index (Fig. 12c). Between 1100 and 1450 AD a prevailing low frequency NAO/AO-like positive index is associated with largely positive departures for both quartz and calcite off East Greenland, whereas on the N Iceland shelf the departures are positive for calcite and negative for quartz (Fig. 12d and e). At around 1450 AD there is a fundamental switch in all records and the reconstructed oscillatory NAO signal is matched by the calcite departures at 2322 (Fig. 12a) whereas the other records (Fig. 12b, d, and e) switch sign with a return to positive calcite departures in the last several decades. The reduction in quartz over the last 900 yr at 2322 may reflect a reduction in iceberg drift because of the development of pervasive land-fast sea ice along the NE Greenland shelf

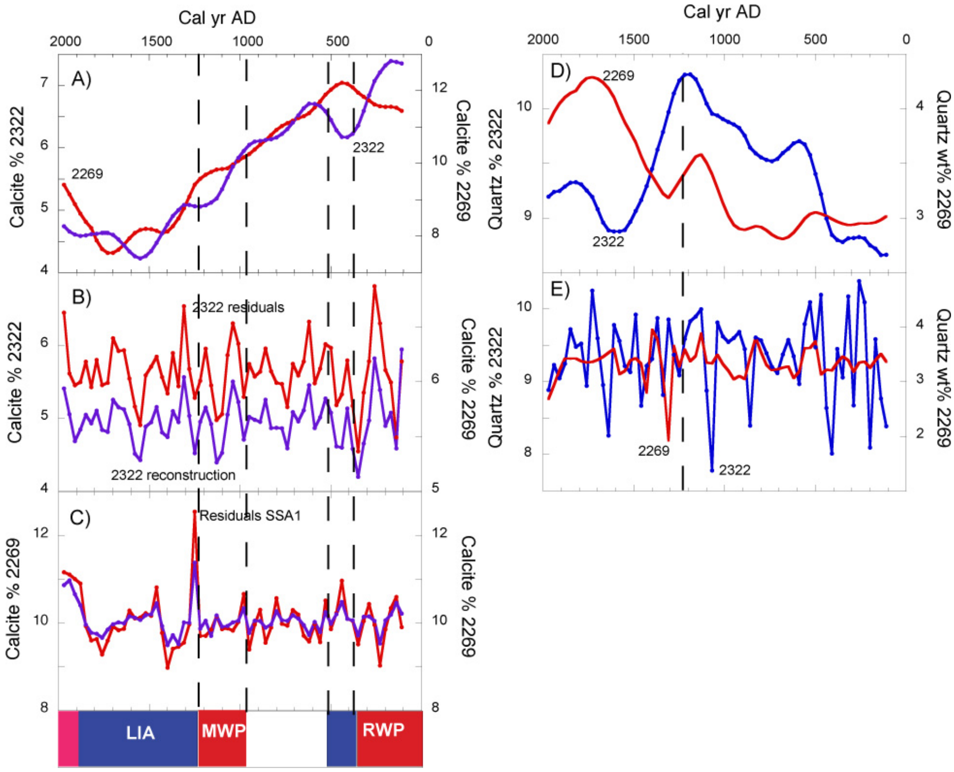

Fig. 11. (A) The SSA1 detrended calcite data for sites MD99-2322 and -2269 (Fig. 1), some 440 km apart. **(B)** Residuals from the SSA1 trend for MD99-2322 (red) and the reconstructed time series (blue). **(C)** Residuals from the SSA1 trend for MD99-2269 (red) and the reconstructed time series (blue). **(D)** SSA trends on quartz wt%. **(E)** residuals from the quartz trends (Fig. 11e). The lower left panel shows the same climate intervals as in Fig. 8.

(Reeh, 2004; Funder et al., 2011). For the interval 1100–1450 the reconstructed NAO (Fig. 12c) (Cook et al., 2002; Trouet et al., 2009; Kinnard et al., 2011) does not accord with our conceptual NAO model (Fig. 2), which links positive calcite departures with a NAO negative circulation. We are unsure as to why this is the case (see also Lehner et al., 2012).

7 Conclusions

Initial and reviewed drafts of this paper strongly focused on "NAO-like" variations as an explanation for our proxies. There is, however, the question as to how far a multi-year atmospheric forcing is a reasonable analog for multi-century to millienial variations (e.g. Figs. 10–12). This might be especially so for our region which lies at the northern margin of the influence of the Icelandic Low. A reviewer suggested that our records might be better considered as representing changes in surface water salinities, which implies a link with variations in the flux of sea ice through Fram Strait (Kwok and Rothrock, 1999; Schmith and Hanssen, 2003), hence the association between the NAO and AO (Fig. 2d) needs to be considered. Several recent papers have evaluated a variety of high-resolution marine proxies from off Iceland and adjacent areas and associated them with known climatic modes, such as the NAO, AO, and AMO (Jennings et al., 2002a; Andrews et al., 2003; Darby and Bischof, 2004; Jiang et al., 2005; Knudsen et al., 2011; Miettinen et al., 2011; Darby et al., 2012; Alonso-Garcia et al., 2013; Zhang et al., 2007), although such oscillations are superimposed on both long-term insolation trends and abrupt events, such as volcanic eruptions. The underlying controls on our proxies are the variations in the landfast and drift ice and the extent and presence of Atlantic water transported via the Irminger current (Fig. 2). Despite potential problems associated with dissolution the calcite content, cores on both sides of the Denmark Strait share a considerable number of features in common (Figs. 8–12) – this is true not only for the last 2000 cal yr BP but for the Holocene (Andrews et al., 2001; Stoner et al., 2007). The trough of the LIA is well defined (Fig. 11a) but there is no distinct MWP. On the 2000 yr timescale the calcite data show a similar long-term trend to the reconstructed Arctic summer temperatures (Kaufman et al., 2009) (Fig. 10a) and have an association with the AMO (Fig. 10c). The abrupt changes in calcite wt% are temporally associated (Figs. 8b, 12a, and d) with the expansion of small ice caps on Baffin Island and East Greenland (Miller et al., 2012; Lowell et al., 2013). Zhong et al. (2011) modeled the climatic impact of a sustained interval of volcanism in the late 13th and middle 15th centuries and indicated that such a history could

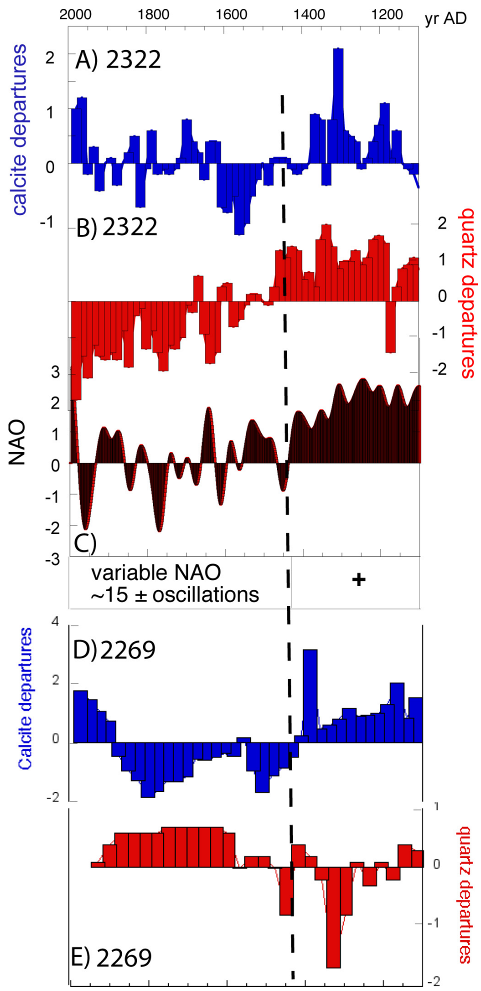

Fig. 12. Departures from the mean for the last 900 yr for calcite (**A** and **D**) and quartz (**A** and **D**) and quartz (**B** and **E**) at sites MD99-2322 and MD99-2269 (Fig. 1) against the NAO index **(C)** (Trouet et al., 2009; Kinnaid et al., 2011). The vertical dashed line ∼ 1450 AD represents the time of a major transition.

result in an expansion of sea ice in the Arctic Ocean and its marginal seas. A notable feature of the LIA reconstructions is the highly variable NAO (Fig. 12c), which is matched in the calcite data from 2322. Our simple conceptual model of the inverse association between the variations in calcite and quartz, based on data from Iceland (Axford et al., 2011) (Fig. 12d and e), is not mimicked by the quartz record on the East Greenland side of the Denmark Strait (Fig. 12a and b). Neither does our current understanding of NAO variations (Fig. 2) and associated changes in our proxies match a reconstructed persistent positive NAO-like scenario between 1100 and 1450 AD (Fig. 12c).

We conclude by noting that there is a fundamental "uncertainty principle" to the chronologies of high-resolution Holocene marine records and the delimitation of episodes such as the MWP and LIA (Fig. 8). This uncertainty is based on: the ± errors on the radiocarbon dates, the calibration to sidereal years, changes in the rate of sediment accumulation, and the ocean reservoir correction. Thus it is difficult to know whether the offset between the calcite data and AMO is a true lag or associated with the dating (Fig. 10c). The problem could be addressed by a very high density of radiocarbon dates (Sejrup et al., 2010) but this is not practical (because of the cost) in most situations.

Acknowledgements. Support for involvement in cruises in 1991, 1993, 1996, 1997, and 1999 was provided to J. T. Andrews and A. E. Jennings by grants from the National Science Foundation. This paper is a contribution to grants NSF-OCE-0823535 and ANS-1107761. J. T. Andrews thanks C.A.S.E. and Jacques Giraudeau for involving him in the project. The calcite and quartz XRD data for MD99-2269 were provided by Matthias Moros (available in NOAA data base). Mike Mann and Anne de Vernal kindly provided data from published climate reconstructions. We greatly appreciate the comments and advice advanced by three reviewers, and which have led to significant changes in the final version, including a change in the title and the suggestion to add more weight to changes in salinity. We appreciate the Editorial advice and guidance of Hans Renssen. The data generated for this paper will be submitted to the NOAA Paleoclimate Data Base for archiving http://www.ncdc.noaa.gov/paleo/paleo.html.

Edited by: H. Renssen

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