Is there evidence for a 4.2ka B.P. event in the northern North Atlantic region?

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Abstract

We review paleoceanographic and paleoclimatic records from the northern North Atlantic to assess the nature of climatic conditions at 4.2ka BP, which has been identified as a time of exceptional climatic anomalies in many parts of the world. The northern North Atlantic region experienced relatively warm conditions in from 6-8ka B.P., followed by a general decline in temperatures after ~5ka B.P., which led to the onset of Neoglaciation. Over the last 5000 years, a series of multi-decadal to century scale fluctuations occurred, superimposed on an overall decline in temperature. Although a few records do show a glacial advance around 4.2ka B.P., because they are not widespread we interpret them as local events -- simply one glacial advance of many that occurred in response to the overall climatic deterioration that characterized the late Holocene.

1. Introduction

The North Atlantic is a key area in the global climate system because changes in atmospheric and oceanographic conditions in this region can have widespread effects on global climate. It is the core region for ventilation of the North Atlantic which drives the Atlantic Meridional Overturning Circulation (AMOC), with global teleconnections through the conveyor belt system of ocean currents. Detailed studies of two sediment cores in the North Atlantic (at ~65° and ~54°N) by Bond et al (1997) revealed quasi-periodic variations in the percentage of hematite-stained grains and Icelandic glass during the Holocene, which were interpreted as evidence for pulses of ice-rafting. They argued that during these episodes, "*cool, ice-bearing surface waters shifted across more than 5° of latitude, each time penetrating well into the core of the North Atlantic Current*". One of the 8 Holocene episodes (later dubbed "Bond events") occurred at ~4.2ka calendar years B.P. Given that this is the time at which exceptional climatic anomalies appear to have occurred in many parts of the world ("the 4.2ka B.P. event") it is important to re-assess the evidence for disruption of the North Atlantic Current at that time.

Bond et al. (2001) argued that the colder episodes they had identified were driven by a reduction in solar insolation (cf. Wanner and Bütikofer, 2008; Wanner et al., 2011), notwithstanding the fact that total solar irradiance did not vary by more than ±0.15% over this period (Vieira et al., 2011; Roth and Joos, 2013; Wu et al. 2018). Nevertheless, the literature is replete with studies that have tried to link diverse paleoclimatic records from around the world to the timing of Bond events (e.g. Fleitmann et al., 2003; Gupta et al., 2003; Wang et al., 2005; Pèlachs et al., 2011), despite the fact that other paleoceanographic studies have been unable to reproduce the record of ice-rafting reported in Bond et al., (1997) (e.g. Andrews et al., 2014). Here we review sedimentary records from the northern North Atlantic (north of 60°N) with a specific focus on whether there is evidence for an "event" around 4.2ka B.P. We do not focus on records from Iceland as these have been reviewed separately by Geirsdóttir et al. (2019).

The North Atlantic has a very distinct pattern of sea surface temperatures, reflecting the ocean currents that traverse the region (Figure 1). Warm sub-tropical water enters the region from the southwest via the Gulf Stream (North Atlantic Current) and this transfers heat to sub-polar latitudes north of Scandinavia by way of the Norwegian Atlantic and West Spitsbergen currents, as well as around the western and northwestern coast of Iceland via the Irminger current. In contrast, cold polar water exits the Arctic Ocean via the East Greenland current, which extends to the southern tip of Greenland. The region between these water masses is where deepwater formation occurs, driving the large-scale Atlantic Meridional Overturning Circulation (AMOC). On the timescale of the Holocene, there have been significant changes in the characteristics and position of these major oceanographic features, as recorded by various paleoceanographic proxies.

2. Paleoceanographic evidence

First, we consider a transect of sediment cores that are aligned along the axis of the main influx of Atlantic water entering the North Atlantic, from west of the UK to Svalbard (Figure 1). We focus on those studies that have provided estimates of paleo sea-surface temperatures. Effectively, this means only those that have analyzed alkenones and diatoms, which reflect conditions in the photic zone or mixed layer near the ocean surface. Figure 1 shows the location of all available Holocene alkenone-based paleotemperature estimates (Figure 2; see references in the caption). These indicate that SSTs were higher in the early Holocene, with the largest anomalies (relative to today) at high latitudes (that is, there was strong polar amplification of the warming) (Andersson et al., 2010). This early Holocene warming was a consequence of orbital forcing: June/July insolation was ~10% higher than today at the start of the Holocene in the northern parts of the region, but the peak warming was delayed due to the influence of the decaying Laurentide and Scandinavian Ice Sheets and associated icebergs and freshwater (Renssen et al., 2009, 2012; Zhang et al., 2016). Consequently, maximum temperatures were a few thousand years later than the peak insolation, punctuated by a short-lived cooling event around 8.2ka B.P. associated with the final major freshwater discharge event of the Laurentide Ice Sheet (Barber et al., 1999; Rohling and Pälike, 2005). Thereafter, as insolation declined so sea surface temperatures declined steadily, or by some estimates, in a more step-like manner (e.g. Calvo et al., 2002; Risebrobakken et al., 2010). For example, Birks and Koç (2002), Andersen et al. (2004) and Berner et al. (2011) all found that August SSTs at 67°N (core MD95-2011) were 4-5°C warmer than today from ~9000-6500 years B.P., then steadily declined. These analyses were based on diatoms, but similar results (albeit with a smaller change in temperature, ~2.5°C, perhaps reflecting a different seasonal bias) were obtained in a study of alkenones from the same core (Calvo et al., 2002). Studies further north, paint a similar picture (Sarnthein et al., 2003; Risebrobakken et al., 2003, 2010; Werner et al., 2014). This pattern of maximum SSTs in the first half of the Holocene and cooling thereafter is seen throughout the eastern North Atlantic, in all proxies that are indicative of conditions in the photic zone (Rimbu et al., 2003; Leduc et al., 2010; Sejrup et al., 2016). The timing of the onset of cooling varies, but cooling was well underway by~5.5ka B.P., in what some refer to as a "transition period" that subsequently led to much cooler conditions in the late Holocene (after 3.5ka B.P.) (e.g. Aagaard-Sorensen et al., 2014; Andersen et al., 2004; Leduc et al., 2010; Sejrup et al., 2016). Although there were short-lived cooling episodes superimposed on the overall first order pattern of temperature change (e.g. Werner et al., 2014), there is no evidence for quasi-periodic cooling episodes disrupting the northward flux of Atlantic water, as described by Bond et al (1997). Proxies of sub-surface conditions (below the mixed layer) – Mg/Ca ratios and oxygen isotopes in forams, as well as foram assemblage changes – generally do not show the same pattern of pan-Holocene cooling as the SST proxies, often indicating slight warming through the Holocene (e.g. Andersson et al., 2010; Sejrup et al., 2011). But these records also do not show a pattern of quasi-periodic cooling events. Could this be because of low resolution in sampling, or poor chronologies? This seems very unlikely as many of these records are from high-deposition rate sites, providing high resolution records that are generally well-dated (e.g. Berner et al., 2011). Indeed, one exceptionally well-dated, high resolution sediment core from the Storegga Slide region (90 AMS ^{14}C dates over 8000 calendar years) provides oxygen isotope data on planktonic forams at a resolution of ±20 years within the core of the Norwegian Atlantic Current at ~64°N. This clearly shows multi-decadal to century-scale variability throughout the last 8000 years, but none of the cold water flux episodes that one would expect to see, based on the work of Bond et al. (1997). We therefore conclude that there is no signal of a 4.2ka B.P. event in paleoceanographic proxies from regions influenced by the flux of warm water from the sub-tropical Atlantic into the Nordic Seas. Cooling of the sea surface had set in more than a millennium earlier in this region.

Next, we consider studies in the western part of the North Atlantic, north of Iceland on the Icelandic Shelf, and further to the west, near Denmark Strait. Here, many studies have examined, *inter alia*, foraminiferal assemblages, coccoliths, dinoflagellate cysts and sea-ice biomarkers and ice-rafted debris (IRD) reflecting transport of material in the cold East Greenland Current (e.g. Andrews et al., 1997; Jennings et al., 2002; Giraudeau et al., 2004; Solignac et al., 2006; Sicre et al., 2008; Justwan et al., 2008; Perner et al., 2015; Moossen et al., 2015; Cabedo-Sanz et al., 2016; Kolling et al., 2017). In this region, warmest conditions occurred around 6.0±1.5ka B.P. (the timing depending on location); these conditions were associated with minimal input of IRD, reflecting the recession of tidewater glaciers onto land along the eastern coast of Greenland, and a weak East Greenland Current, with minimal stratification of the water column at that time as the flux of warmer, more saline Irminger Current water increased (Justwan et al., 2008; Jennings et al., 2011; Werner et al., 2014; Telesinski et al., 2014; Perner et al., 2016). Conditions began to change by ~5.0±0.5ka B.P. (the timing varying geographically) when cold water diatoms and forams, sea-ice (as tracked by the biomarker index, IP_{25}) and IRD started to increase, and the water column became more stratified as the East Greenland Current strengthened (Moros et al., 2006; Telesinski et al., 2014; Perner et al., 2016; Kristiansdottir et al., 2017). These changes correspond to the re-advance of glaciers in East Greenland, part of the much more widespread onset of neoglaciation that is well-documented in many regions around the North Atlantic (Solomina et al., 2015). Warmer conditions (related to a strengthened Irminger Current) developed over the past 2000 years, but this period is also characterized by a series of minor fluctuations in the extent of ice in the region, with much colder conditions after ~1.0ka B.P. when the coldest conditions of the last 8000 years occurred, with abundant IRD and sea-ice in Denmark Strait and off the north coast of Iceland (Bendle and Rosell-Mele, 2007; Andresen et al., 2013; Cabedo-Sanz at al 2016; Kolling et al., 2017). None of these records show evidence of an unusual anomaly at 4.2ka B.P.; rather, the overall cooling of the late Holocene began 500-1000 years earlier (cf. Orme et al., 2018). Similar variability is also seen further south and southwest of Iceland, at ~59°N (Farmer et al., 2008; Moros et al., 2012; Orme et al., 2018) though there is evidence from dinocysts for an anomaly in the seasonality of SSTs at ~4.5ka B.P., perhaps related to a westward shift in the Sub-Polar Gyre, allowing warmer Atlantic water to influence the site (van Nieuwenhove et al., 2018).

This review of paleoceanographic studies extending from southern Greenland to Fram Strait, and from western Svalbard and the southern Barents Sea southward to 60°N, provides no evidence for a significant change in major oceanographic conditions that could be linked to the 4.2ka B.P. climate anomaly seen elsewhere. Rather, the evidence points to a more gradual change that was well under way by ~5ka B.P., from the relatively warm conditions of the early Holocene (driven by precessional forcing) to much colder conditions that have characterized the last 3 millennia.

3. Terrestrial records from around the North Atlantic

3.1 Eastern Greenland and the Greenland Ice sheet

Lake sediment records from sites along the coast of eastern Greenland provide a record of Holocene environmental conditions that generally reinforce the paleoceanographic evidence discussed earlier. A "Holocene Thermal Maximum" (characterized *inter alia* by longer ice-free conditions, higher levels of lacustrine productivity, increased evaporation, more tundra vegetation and higher levels of terrestrial plant material transferred to lakes) is clearly seen from ~8ka B.P. (or earlier) to ~5.0±0.5ka B.P (e.g. Kaplan et al., 2002; Andresen et al., 2004; Schmidt et al., 2011; Balascio et al., 2013; Wagner and Bennike, 2015; Axford et al., 2017; van der Bilt et al., 2018a). Thereafter, conditions became colder, often with a decline in vegetation cover, an increase in the flux of coarse-grained sediments, and a shift in the types of chironomids and diatoms present, towards species that thrive in cooler conditions. At the same time, in glacierized watersheds, the growth of glaciers led to an increase in the flux of minerogenic material which is a diagnostic signal of the onset of late Holocene neoglaciation across the region. In Kulusuk Lake (65°N on the coast of southeastern Greenland) this change occurred at ~4.2ka B.P., when there was an abrupt increase in clastic sediments from glaciers that had probably disappeared during the mid-Holocene warm period (Balascio et al., 2015). A similar transition is seen in sediments from nearby Ymer Lake, where a higher frequency of avalanches and a longer season with ice-cover is thought to have favored the transfer of coarser material into the lake after ~4ka B.P. (van der Bilt et al., 2018). At another site in the same region, the Holocene thermal maximum was identified (via the evaporative enrichment of δD in leaf wax n-alkanes) from 8.4 to 4.1ka B.P, followed by a decrease in evaporation as the open water season became shorter. At the same time, there was an increase in the flux of clastic sediments and terrestrial organic material into the lake as river runoff increased (Balascio et al., 2013). In all of these studies, it is clear that there was a fairly rapid transition from warm mid-Holocene conditions to the colder, wetter late Holocene that encompassed the 4.2ka B.P. interval of interest. In some cases, there is evidence for a short-lived "event" at around that time (e.g., at Kulusuk Lake; Balascio et al., 2015) but this appears to be simply a part of the overall deterioration in climate that led to ice growth across the region. There is currently no evidence for a more widespread glacial advance at 4.2ka B.P. Given that cooling was persistent over the last 5000 years, and the elevational threshold for glacierization is close to mountain tops across the region (declining in elevation poleward) it is understandable that different locations would have experienced the onset of neoglaciation at different times (cf. Geirsdottir et al., 2019). However, as the ELA continued to lower over the last 3-4 millennia, glaciers that had greatly diminished in size, or disappeared entirely, during the warmest period of the Holocene were eventually regenerated, with the exact timing varying across the region. In the case of Kulusuk Lake, it seems reasonable to conclude that the steady decline in temperatures and the specific hypsography of that basin led to a short-lived positive mass balance, with early ice growth and associated sediment input to the lake around 4.2ka B.P. This was the first of several advances within the Neoglacial period.

Ice cores from Greenland provide records of past climate variations from oxygen isotopes, glaciochemistry and physical characteristics, which are broadly consistent with those from coastal lake sediments. Alley and Anandakrishnan (1995) examined evidence for summer melting in the GISP2 ice core, as recorded by changes in the physical properties of the ice. Their analysis was at a relatively low resolution, but they showed maximum Holocene summer temperatures from ~7.5ka B.P., followed by a two-step transition to colder conditions, from ~6.5 to 5.5ka B.P., and ~4.5 to 4ka B.P., with persistently low summer temperatures (minimal melting) thereafter. After adjusting for ice thickness changes, Vinther et al. (2009) also showed that there was an overall decline in temperature at the Summit of the Greenland Ice Sheet (73°N, 3210 masl) over the last ~9,000 years (interpreted from changes in $\delta^{18}O$ in the GISP2 ice core). The mean temperature of the warmest and coldest millennia (7-8ka and 0-1ka b2k, respectively) differ by ~2.35°C (assuming no change in the seasonality of snowfall on the ice sheet). Superimposed on the long-term temperature decline there were multidecadal anomalies on the order of ±1°C. One of the largest of the negative anomalies after the well-known 8.2ka B.P. event began ~4400 b2k and reached a minimum at 4340 b2k, but by 4200 b2k, temperatures had sharply increased (Figure 3). In the Vinther et al. (2009) reconstruction (Figure 3a, which combines data from Renland and Agassiz Ice Caps), this appears to be driven mainly by the record from Agassiz Ice Cap on Ellesmere Island (Figure 3b); nothing comparable is seen in oxygen isotopic records from Summit or Renland (Figures 3b, 3c), or in the Summit temperature reconstruction of Kobashi et al. (2017), based on the differential diffusion of argon and nitrogen isotopes in firn prior to its densification into ice (Figure 3d). However, $\delta^{18}O$ in chironomid head capsules from a lake in northwest Greenland also recorded the highest values of the last ~6000 years at ~4.2ka B.P. (Lasher et al., 2017) and at Camp Century, there was a local isotopic maximum shortly before 4000 B.P. (Figure 3b). In Murray Lake (northeastern Ellesmere Island), relatively warm conditions at ~4.2ka B.P. were reconstructed from varve thickness (Cook et al.,2009). Similarly, Gkinis et al. (2014) found an abrupt increase in temperature in the NorthGRIP ice core at ~4200 b2k after deconvolving the isotopic record to take into account diffusion effects that have smoothed the signal. However, this technique is very sensitive to the assumptions made about the past accumulation rate, as diffusion is a function of both past accumulation and temperature. For example, a 15% reduction in accumulation would reduce an apparent temperature anomaly from 5°C to 3.5°C (Gkinis, pers. comm.). Under the assumption of no changes in accumulation rate, Gkinis et al. (2014) identify a warm period in the North GRIP core at 4.2ka B.P. and refer to this as the "*mid-Holocene optimum*". It will be interesting to see if this technique, when applied to other ice cores, reveals more details about short-term temperature fluctuations that may have been obscured by diffusion effects. But for now, only 3 records, from northwest Greenland and northern Ellesmere Island, point to short-lived warmer conditions at ~4.2ka B.P., in contrast to the majority of records that indicate temperatures were declining at that time.

3.2 Iceland

We did not undertake a review of the literature on the Holocene paleoclimatology of Iceland as that is well summarized by Geirsdóttir et al (2019). They conclude that Neoglaciation in Iceland had begun by 5ka B.P. but different topography and proximity to the ocean led to varying environmental effects across the island. Several step-like changes occurred during the last 5ka B.P., culminating in the most extensive glacier advances during the last millennium. One of the step-like changes occurred at ~4.5-4.0ka B.P, and they conclude that this is indistinguishable from a "4.2ka B.P. event". They note that the eruption of Hekla at 4.2ka deposited at ≥1cm of tephra over 80% of Iceland, so the direct effects on the landscape at that time complicate the detection of a signal that may be related to other forcing factors. Of the two lakes in NE Greenland that did not have a tephra in the sediments, one (Skoravatn) shows an abrupt change at 4.2ka B.P., while the other (Tröllkonuvatn) does not.

3.3 Svalbard

Lake sediment records from Svalbard record changes in climate at the northernmost limit of North Atlantic water (the West Spitsbergen Current). All studies describe a warm early Holocene phase when many of the glaciers seen today were small or absent (Farnsworth, 2018). On Amsterdamoya, at the northwestern edge of Svalbard, warm and dry conditions spanned the interval from 7.7 to 5ka B.P.; glaciers were small or absent by 8.4ka B.P., only re-forming in the late Holocene (Gjerde et al., 2018; de Wet et al., 2018). To the south, on the Mitrahalvoya Peninsula, there is also evidence that glaciers reached their minimum size by the mid-Holocene, but subsequently re-formed or re-advanced. Karlbreen began to grow around 3.5ka B.P. (Røthe et al., 2015) but in the neighboring watershed of Hajeren an abrupt increase in minerogenic sediments at 4.25 ka B.P. registered the onset of neoglaciation in that basin (van der Bilt et al., 2015). Paleotemperature estimates (from alkenones) in the same record indicate this advance was triggered by an abrupt drop in temperature at that time; thereafter, temperatures remained low (van der Bilt et al., 2018b). Other records from the region indicate that the first neoglacial advances of glaciers occurred around 4.6ka B.P. (e.g. Svendsen and Mangerud, 1997; Reusche et al., 2014).

3.4 Scandinavia

As most glaciers in Scandinavia had their largest areal extent during the "Little Ice Age" (~A.D. 1400-1850), information about past glaciers in Norway during the late Holocene is based on reconstructions from indirect evidence, mainly sediments deposited in distal glacier-fed lakes (e.g. Nesje 2009, Bakke et al., 2010; 2013). After several large glacier advances in the earliest Holocene, the climate was generally warm during the early Holocene (8.5-6.5ka B.P.) and most glaciers melted away completely (Nesje 2009) (Figure 4). Around 6 ka B.P. glaciers start to re-grow mainly as a function of decreasing summer insolation over the Northern Hemisphere (Wanner et al. 2008). The regrowth of glaciers follows a pattern of gradual increases in glacier size interrupted by smaller glacier advances (Bakke et al, 2010, 2013; Vasskog et al., 2012). Along a coastal south-north transect through Scandinavia, different locations have experienced the onset of neoglaciation at different times, mainly as a function of altitude (cf. Geirsdóttir et al., 2018). By 2ka B.P. many glaciers had reached present day size, but maximum glacier extent was in the 18th century, during the Little Ice Age (Nesje 2009). A review of more than 20 papers shows that none of them indicate any abrupt anomalous change in glacier extent connected to a perturbation of climate around 4.2 ka. (Bakke et al., 2005a; 2005b; 2008; 2010; 2013; Dahl and Nesje; 1992; 1994; 1996; Lauritzen 1996; Snowball and Sandgren, 1996; Seierstad et al., 2002; Lie et al., 2004; Nesje et al. 2009; Vasskog et al., 2011; 2012 Støren et al., 2008; Wittmeier et al., 2015; Shakesby et al., 2007; Kvisvik et al., 2015, Gjerde et al., 2016). Investigating this further, we examined other terrestrial evidence mainly pollen, macrofossil and diatom records derived from lake sediments (e.g. Bjune et al., 2005; Velle et al., 2005). They have a time resolution somewhat lower than the glacier reconstructions (typical 500 yr spacing) but they all reflect the general decrease in summer insolation over the northern hemisphere and no abrupt transition close to 4.2ka B.P. (Bjune, 2005; Bjune et al., 2004, 2006; Velle et al., 2005). The only terrestrial evidence from Scandinavia that shows a clear anomaly close to 4.2ka B.P. is a speleothem record of $\delta^{18}O$ from Northern Norway which records a short-lived temperature maximum (isotopic minimum) at ~4ka, before rapidly decreasing to much colder temperatures at ~3.7ka B.P. (Lauritzen and Lundberg 1999). However, a speleothem from a nearby cave (Okshola) does not show a comparable anomaly at this time (Linge et al., 2009).

4. Conclusions

A review of paleoceanographic and terrestrial paleoclimatic data from around the northern North Atlantic reveals no compelling evidence for a significant and widespread climatic anomaly at ~4.2ka B.P. (i.e., an "event") in most areas. In particular, there is no supporting evidence for "*cool, ice-bearing surface waters… penetrating well into the core of the North Atlantic Current"* at that time, as described by Bond et al., (2001). The region experienced relatively warm conditions from 6-8ka B.P. followed by a general decline in temperatures after ~5ka B.P., signaling the onset of Neoglaciation. Over the last 5000 years, a series of multi-decadal to century scale fluctuations occurred, superimposed on an overall decline in temperature. Against this background of declining temperatures, three records in northwest Greenland and Ellesmere Island show an unusual warm anomaly around 4.2ka B.P., and a few others (in SE Greenland, Iceland and western Svalbard) show a cold anomaly, associated with a glacial advance. We interpret these as local events -- simply one glacial advance of many that occurred in response to the overall climatic deterioration that characterized the late Holocene.

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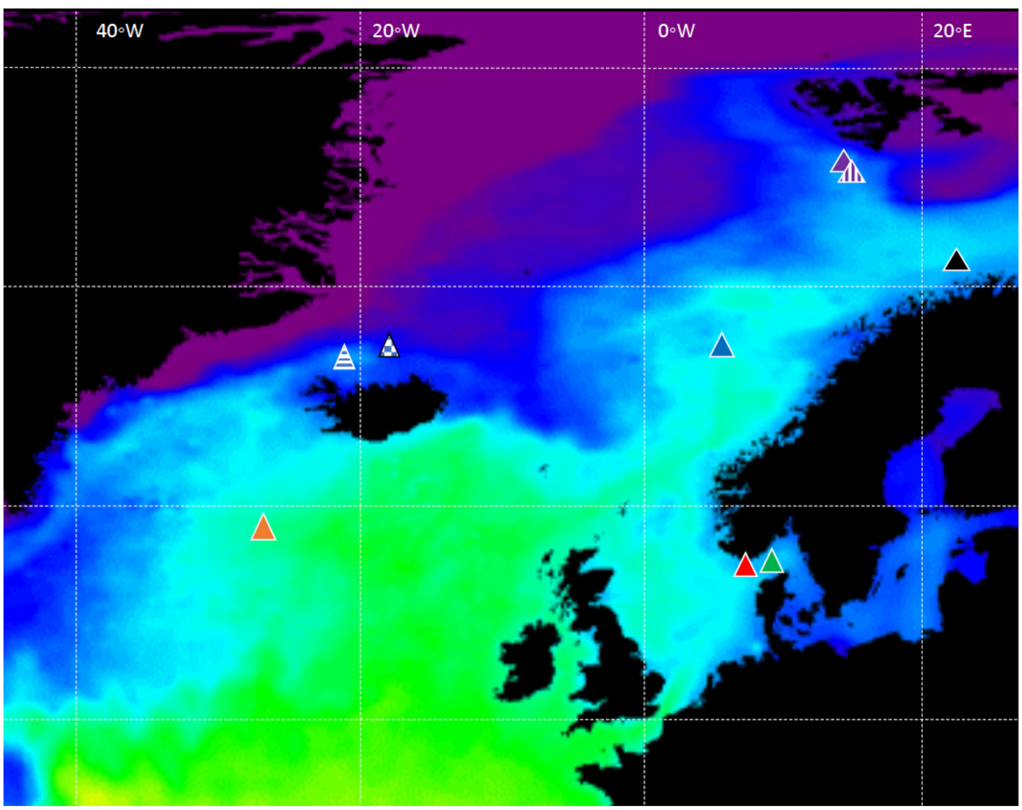
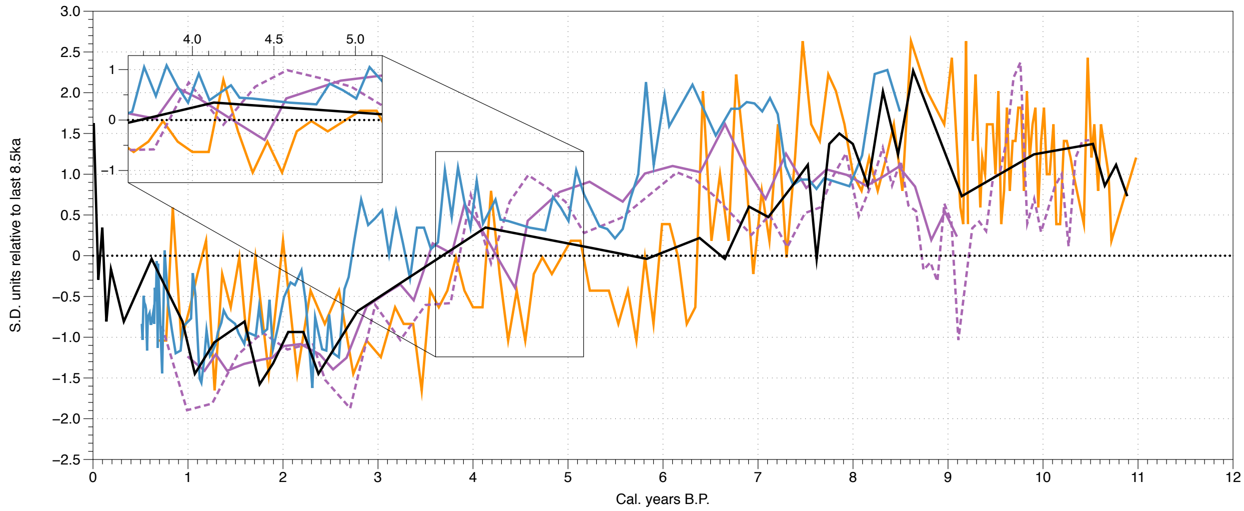


Figure 1. Location of sediment cores used to obtain the alkenone-based paleo SST estimates shown in Figure 2.



Black = Risebrobakken et al., 2010: core PSh-5159N, 71.35N, 22.63E
Purple = Marchal et al., 2002: core M23258-2, 75N, 13.97E
Dashed Purple = Rigual-Hernández et al 2017: core SV-04, 74.957, 13.899E
Orange = Marchal et al., 2002: core MD95-2015, 58.76N, 25.958W
Blue = Calvo et al., 2002: core MD95-2011, 66.97N, 7.633E

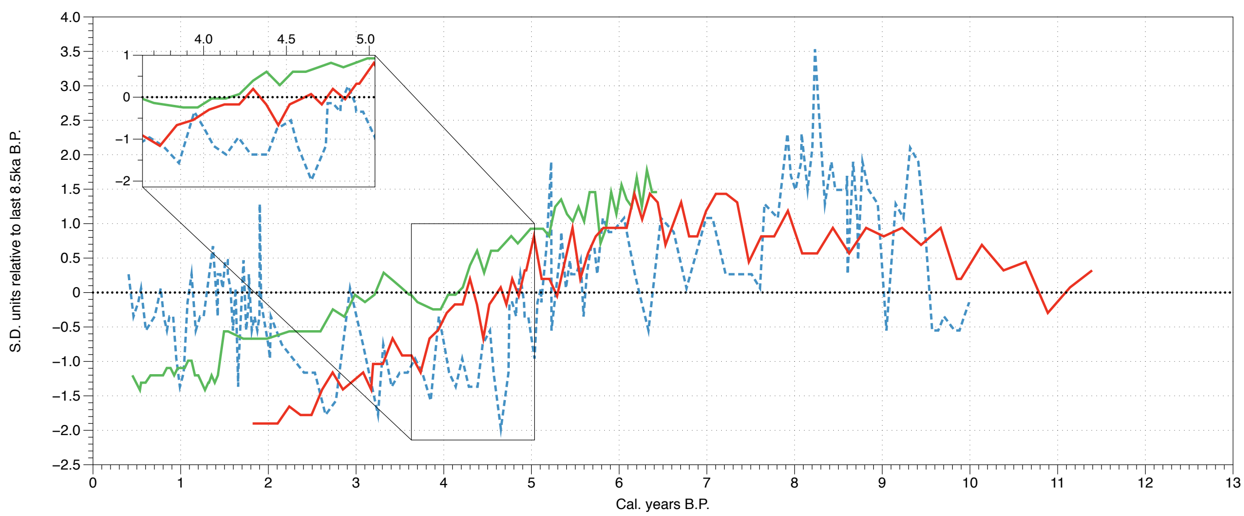
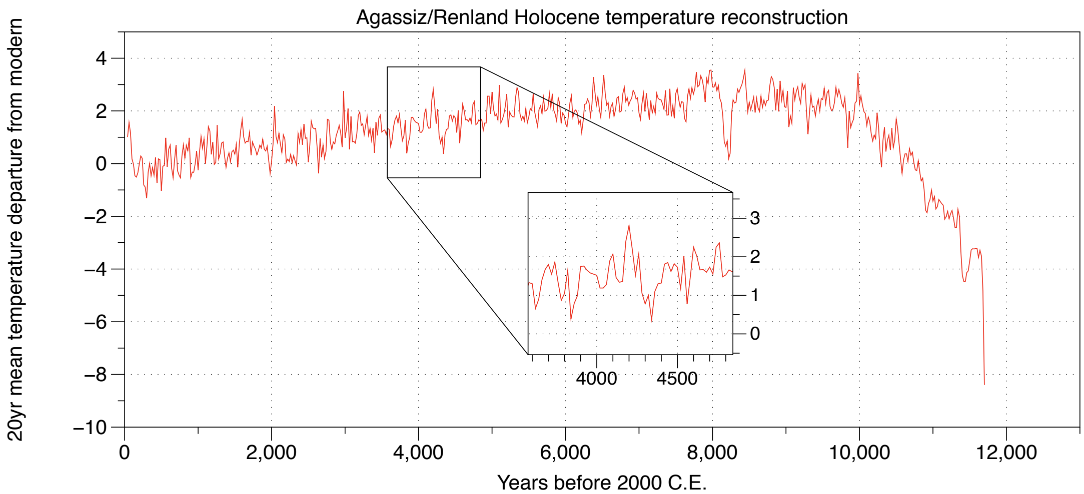


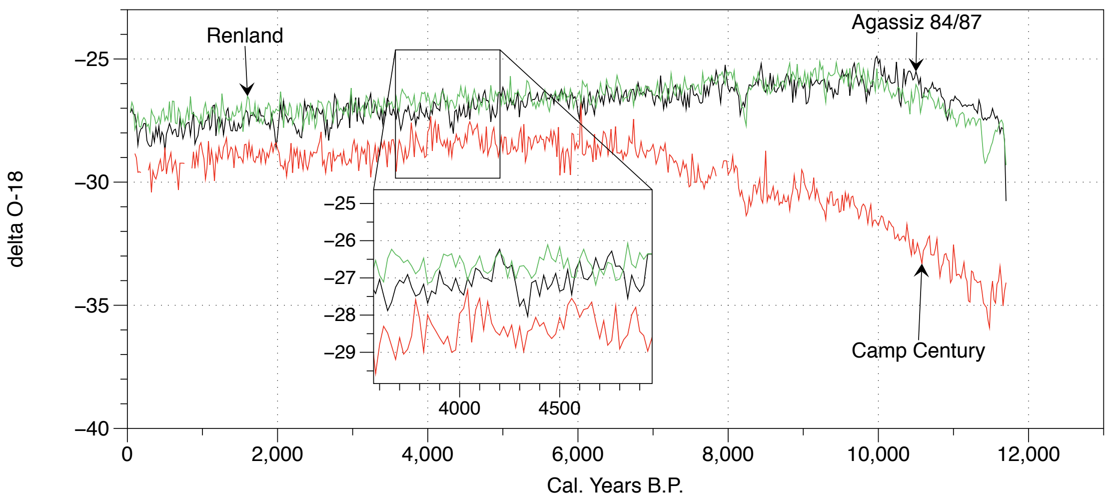
Figure 2. Alkenone-based paleo SST estimates from North Atlantic sites (shown in Figure 1).
Red = Emeis et al., 2003: core IOW 22517, 57.67N, 7.091E
Green = Emeis et al., 2003: core IOW 22514, 57.84N, 8.704E
Dashed Blue= Kristiansdottir et al., 2017: core MD2269, 66.63N, 20.85W

a)



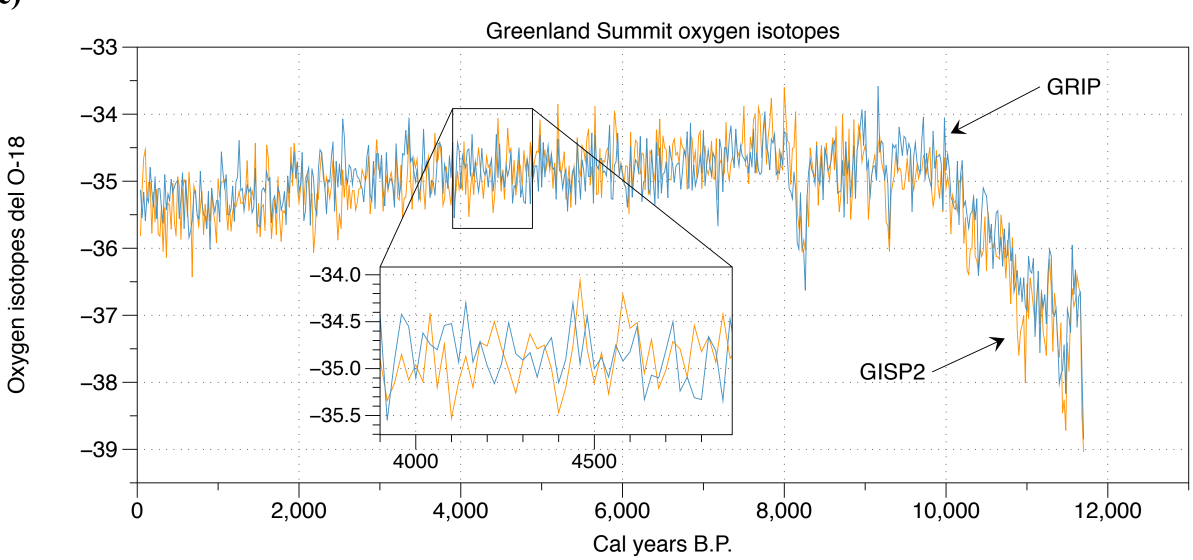
Agassiz/Renland Holocene temperature reconstruction
20yr mean temperature departure from modern
Years before 2000 C.E.

b)



Renland
Agassiz 84/87
Camp Century
delta O-18
Cal. Years B.P.

c)



Greenland Summit oxygen isotopes
GRIP
GISP2
Oxygen isotopes del O-18
Cal years B.P.

d)

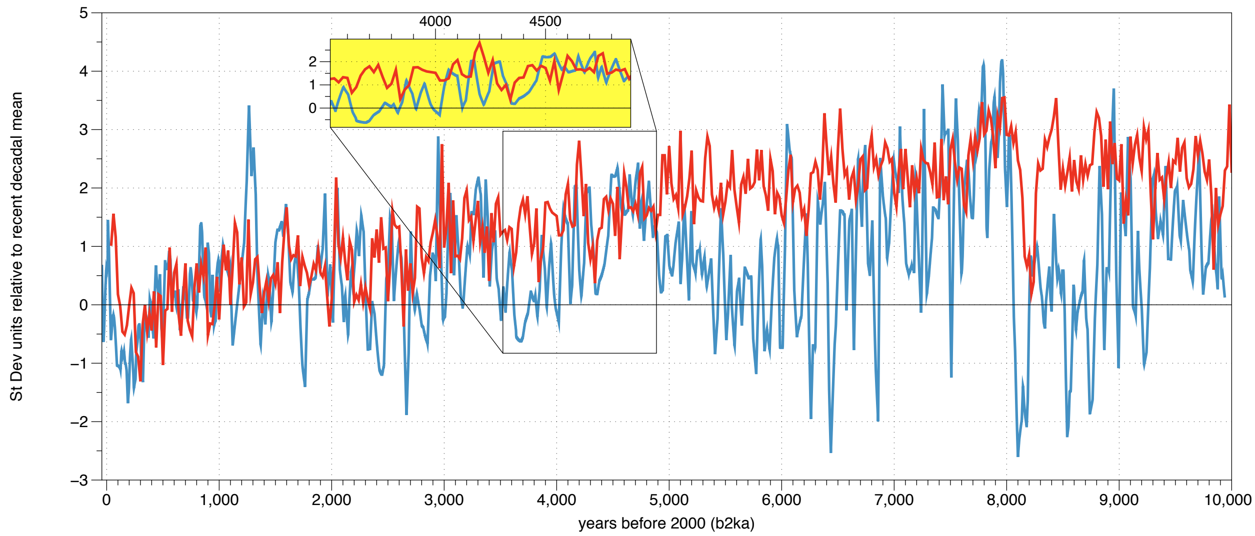


Figure 3. a) Temperature anomalies from the smoothed estimate of present temperatures in Greenland, based on oxygen isotope records from Renland Ice Cap (East Greenland) and Agassiz Ice Cap (Ellesmere Island). Timescale is in years b2k (before A.D. 2000). The interval around 4.2ka BP is enlarged in the box (Data source: Vinther et al., 2009).
b) Individual oxygen isotopic records from Renland and Agassiz Ice Caps (which were combined to create the record in Figure 3a), and from Camp Century
c) Individual oxygen isotopic records from GRIP and GISP2 at Summit, Greenland Ice Sheet
d) Paleotemperature estimates from argon and nitrogen isotopes (in blue) (from Kobashi et al., 2017) and from the Renland/Agassiz joint record (in red) as shown in Figure 3a (from Vinther et al., 2009).

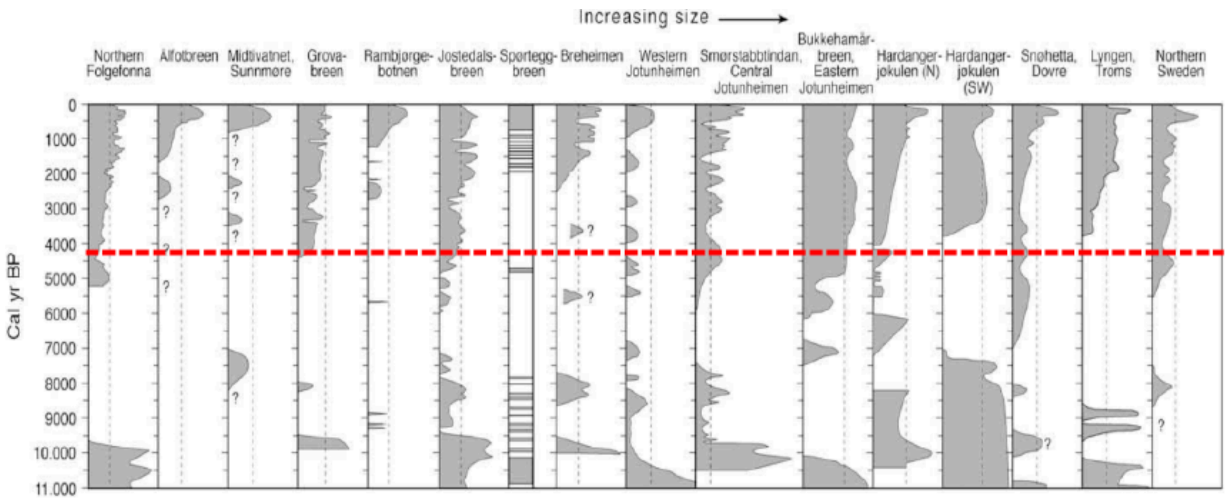


Figure 4. Summary of glacier extent in various regions of Scandinavia during the Holocene. 4.2ka B.P. is highlighted by the red dashed line (after Nesje, 2009).