Younger Dryas ice-margin retreat in Greenland, new evidence from Southwest Greenland

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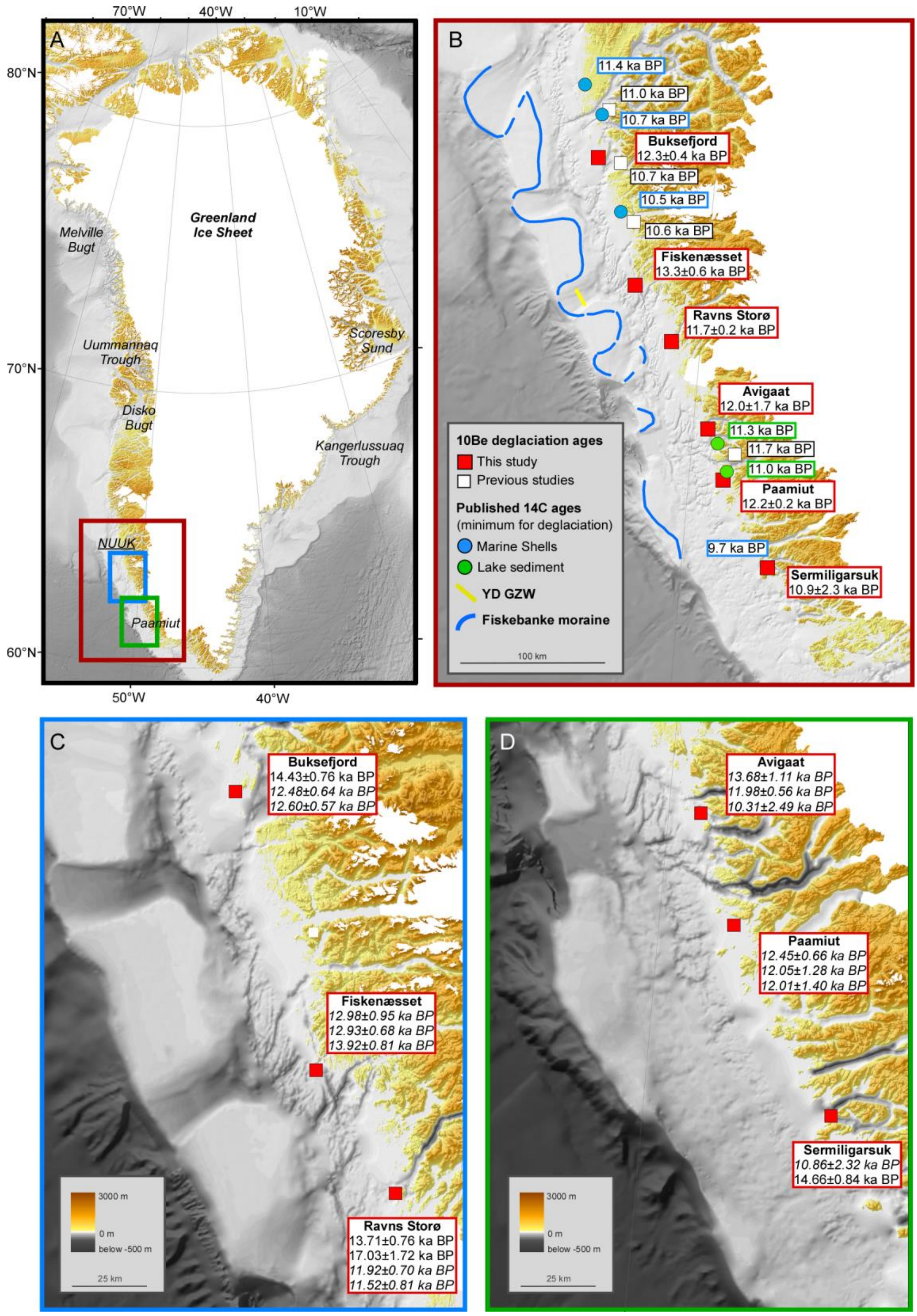
Abstract. To date the final stage in deglaciation of the Greenland shelf, when a contiguous ice sheet margin on the inner shelf transitioned to outlet glaciers in troughs with intervening ice-free areas, we generated cosmogenic ^{10}Be dates from bedrock knobs on six outlying islands along a stretch of 300 km of the Southwest Greenland coast. Despite ^{10}Be inheritance influencing some dates, the ages generally support a Greenland Ice Sheet (GrIS) margin that retreated off the inner shelf during the middle Younger Dryas (YD) period. Published ^{10}Be and ^{14}C-dated records show that this history of the GrIS margin is seen in other parts of Greenland, but with large variations in extent and speed of retreat, sometimes even between neighbouring areas. Areas with a chronology extending into the Allerød period show no marked ice margin change at the Allerød/YD transition except in northernmost Greenland. In contrast, landforms on the shelf (moraines and grounding zone wedges) have been suggested to indicate YD readvances or long-lasting ice-margin stillstands on the middle shelf. However, these features have been dated primarily by correlation with cold periods in the ice core temperature records. Ice-margin retreat during the middle and late YD is explained by advection of warm subsurface water at the ice-margin, and by increased seasonality. Our results therefore point to the complexity of the climate/ice-margin relation, and to the urgent need for direct dating of the early deglaciation history of Greenland.

Keywords: Younger Dryas, Greenland ice sheet, Climate change, Cosmogenic exposure dating.

1 Introduction

The Younger Dryas (YD) cold climate oscillation from 12.8 to 11.7 ka BP (thousand years Before Present) began after the Allerød warm period with a 200-yr-long period of cooling, and culminated with a 60-yr-long period of abrupt warming, as recorded in Greenland ice cores (Steffensen et al., 2008). Over the Greenland Ice Sheet (GrIS), annual mean temperatures dropped between 5 and 9°C (Buizert et al., 2014), when both summer insolation (65° N) and atmospheric CO_2 were increasing (e.g. Buizert et al., 2014). YD climate changes were especially concentrated around the north-eastern North Atlantic in the areas of Atlantic Meridional Overturning Circulation (AMOC) (Carlson, 2013). Similar to present climate change, the YD oscillation was a result of perturbations in the Earth's climate system, and, with a view to the future, it is of great interest to study the effect of these climate changes on the margin of the GrIS. During YD, it seems that the GrIS in most areas had its margin on the shelf, and earlier work has concentrated on the behaviour of ice streams in transverse troughs on the shelf (e.g. Larsen et al., 2016and newer references discussed below)

In this study, we present 18 new cosmogenic ^{10}Be exposure ages from six localities from the inner shelf spanning 300 km of Southwest Greenland. Our purpose is to shed light on ice-margin behaviour during the final phase of deglaciation of the shelf when a contiguous GrIS margin transformed into outlet glaciers in transverse troughs feeding the shelf (Fig. 1). Despite field observations that coastal islands experienced warm-based glacial scouring, ^{10}Be inheritance from episodes of earlier exposure influence some samples in our chronology. Still, clustered ages suggest that the GrIS margin generally retreated during the middle YD. A survey of previous studies elsewhere in Greenland show that these results are in line with other dated records from Greenland, also records going back to Allerød times, which show no evidence for response to YD cooling. Moraines and grounding zone wedges (GZW) on the shelf have earlier been interpreted as evidence for a YD ice margin readvance, but these are dated only by climatic inference. The apparent mismatch between temperature records and ice margin behaviour, as well as mechanisms which may overrule or mute the effect of temperature change in this environment are discussed.



Fgure 1: Location of study area (A), Estimated deglaciation ages and ^{14}C minimum constraints from this work and previous studie (B); Details of cosmogenic dates from this work, and shelf bathymetry. For references to previous results see text. Background map of Greenland and surrounding seas from BedMachine Greenland v.3 (Morlighem et al. 2017.

2. Background

2.1. Setting

The continental shelf in the study area in Southwest Greenland narrows from a width of c. 70 km in the north to c. 50 km in the south (Fig. 1). It is composed of an inner c. 25-km-wide and up to 500-m-deep trough running along the coast and dissected by glacial erosion in Proterozoic orthogneiss bedrock (Henriksen, 2008). On the outer shelf, a belt of shallow banks with an gently undulating surface are composed of younger stratified marine and fluvial sediments. These banks are dissected by 400- to 500-m-deep transverse troughs that are a continuation of the major fjords inland (Holtedahl, 1970; Henderson, 1975; Sommerhoff, 1975; Roksandic, 1979; Sommerhoff, 1981; Ryan et al., 2016). At a distance of 10-15 km beyond the coast, the inner trough forms an archipelago with a multitude of small glacially sculptured rocky islands and skerries, reflecting intensive, but uneven glacial erosion. From these rocky islands we collected our samples (Fig. 2).

2.2. Deglaciation history

Although there is little evidence for glacier overriding, it is likely that the ice sheet covered the rather narrow shelf during LGM based on evidence from a marine core in the Davis Strait outside the Fiskenæsset trough suggests that the ice-margin here stood at the shelf break until deglaciation began at c. 18.6 cal. ka BP (Winsor et al., 2015a). By c. 11 ka, the retreating ice-margin reached the present coastline, and the subsequent deglaciation of the fjords and land began, as summarised by Winsor et al. (2015b). This leaves a period of c. 7 ka with the ice-margin located on the shelf inland of the shelf break. Possible evidence for prolonged GrIS margin during this interval is a series of lobate moraines that run along the troughs and impinge on the inner side of the banks (Fig. 1) (Sommerhoff, 1975; Winsor et al., 2015a). From their setting, these moraines were correlated with the Fiskebanke moraine system to the north (Funder et al., 2011), where they were thought to date from a YD readvance on the shelf (van Tatenhove et al., 1996; Roberts et al., 2009). In our area a limited YD readvance on the inner shelf , Neria stade, was postulated by (Weidick et al., 2004) , based on weathering limits on coastal mountains. A YD readvance in this part of the GrIS was also suggested by modelling, which indicated that the ice-margin in SW Greenland retreated from the shelf edge to the present coastline in the Bølling-Allerød period, but then returned to the shelf during YD (Simpson et al., 2009; Lecavalier et al., 2014). A grounding zone wedge in the Fiskenæsset trough points to a stillstand or readvance of the glacier front at an unknown time during deglaciation (Fig. 1, Ryan et al., 2016). The significance of these features is discussed below in the light of our new chronology.

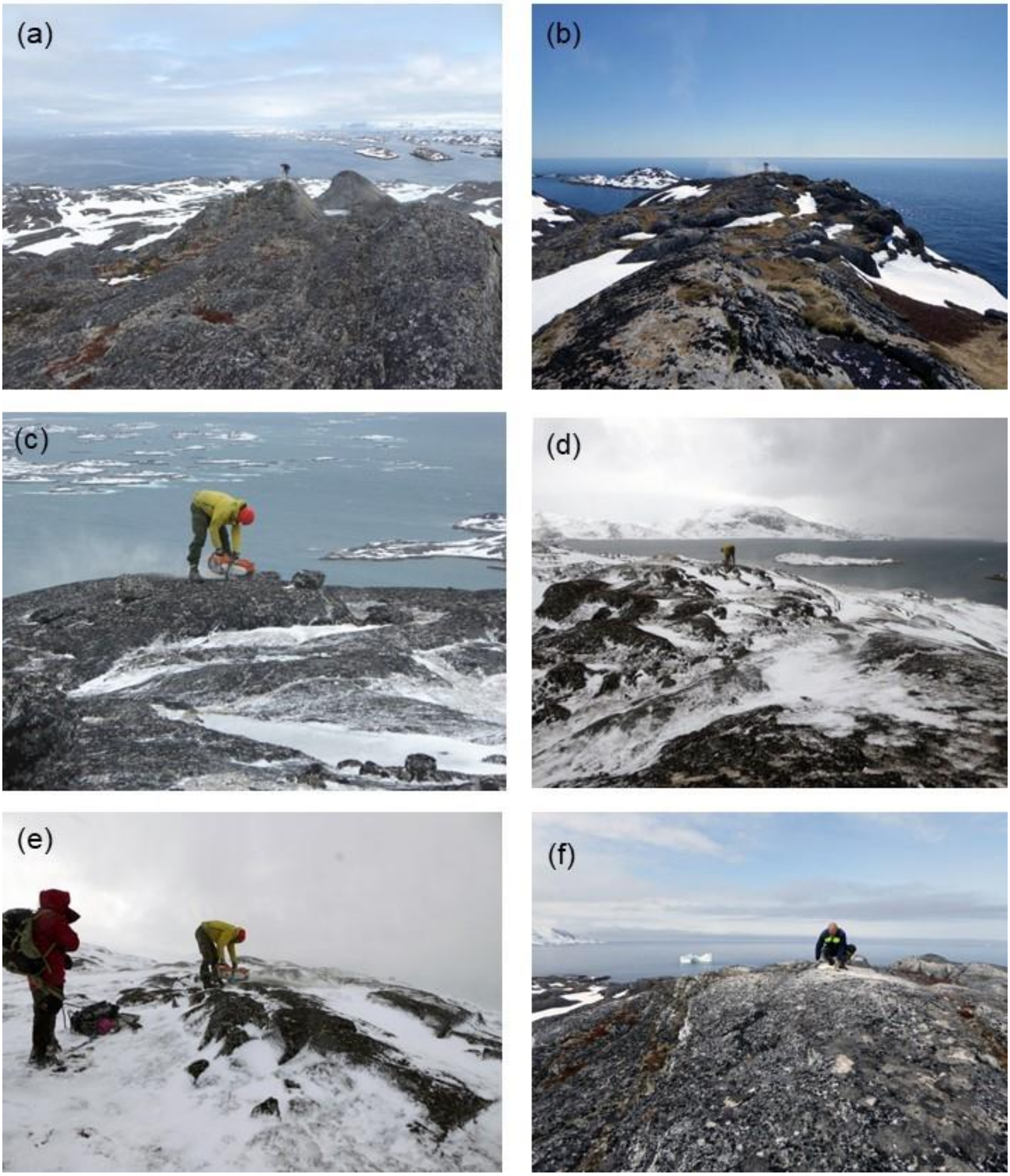


Figure 2: Sampling localities: (a) Buksefjord (Sample X1526, 12.0 ka), (b) Fiskenæsset (Sample X1521, 13.0 ka), (c) Ravns Storø (Sample X1520, 17.0 ka, inheritance), (d) Avigaat (Sample X1518, 10.3 ka), (e) Pamiut (Sample X1515, 12.0 ka), (f) Sermiligarssuk (Sample X1507, 10.9 ka).

3. Field and laboratory methods

Samples were collected from the summits of bedrock knobs in glacially sculptured islands along the inner shelf in Southwest Greenland. Unfortunately, erratic boulders on the bare rock surfaces were largely absent. This potentially represents a problem because while boulders were ideally incorporated in the ice in a pristine condition without previous exposure to cosmic radiation, the glacial erosion of the bedrock surface may not have been deep enough to remove inherited isotopes from older exposures, which may result in overestimation of the age (Briner et al., 2006; Corbett et al., 2013; Larsen et al., 2014). To minimize the risk for inheritance, we selected sites in the lowland where the overlying ice would have been thickest and most erosive, but above the marine limit to avoid the risk of shielding of the rock surface by the sea. From each site we collected 3-4 samples within a radius of 100 m to be sure that all samples from each locality had been deglaciated at the same time.

Contrary to inheritance, other geologic processes may yield young ages that are younger than true time of exposure. This may occur if a surface has been partially shielded from cosmic radiation by vegetation or snow cover for some time (Gosse and Phillips, 2001). However, it is unlikely that the rocky summits were vegetated, as soil would have been washed into the depressions of the glacial sculpture, which was not osbserved (Fig. 2). Also, long-lasting and deep snow cover over sample sites is unlikely in the stormy and maritime climate at the outer coast. Indeed, we experienced heavy snowfall during the sampling, with thick snow accumulating in hollows, while the tops were left free of snow (Fig. 2). Topographic shielding from nearby mountains was checked with a clinometer in the field, but in all cases was non-existent.

The laboratory work comprised sample preparation at the University of Buffalo and measurement of ^{10}Be-concentrations at the AMS facility at Aarhus University. The laboratory procedure for the preparation followed the University at Buffalo's protocols (Briner, 2015). Samples were crushed and sieved to 250–500 μm, then exposed to a magnetic separator to remove the more magnetic minerals and facilitate the subsequent froth flotation. In addition to flotation, some samples (X1509, X1513, X1521) had to undergo heavy mineral separation to obtain sufficient amounts of quartz. Before the next step the samples were examined under a microscope to see if they had been substantially purified. Finally, the samples were etched by hydrogen chloride (HCl) and a mixture of hydrofluoric and nitric acid (HF/NHO_3) in order to further isolate pure quartz from remaining minerals. Quartz purity was then verified by inductively coupled plasma optical emission spectroscopy at the University of Colorado. Pure quartz samples were fully dissolved with a ^{9}Be carrier and $Be(OH)_2$ was isolated through column separation. The $^{10}Be/^{9}Be$ ratios were measured at Aarhus AMS Centre (AARAMS) and all samples were blank corrected (Olsen et al., 2016). Nuclide concentrations were normalized to the Beryllium standard 07KNSTD (Nishiizumi et al., 2007).

The ages were calculated with the CRONUS-Earth online calculator (Balco et al., 2008), using the $^{10}Be/^{9}Be$-ratio measured by the AMS subtracted the processed blank ratio. The processed blank ratio was 2.10×10^{-15} and the blank-corrected sample ratios ranged from 0.76×10^{-13} to 2.58×10^{-13}. The Arctic ^{10}Be production rate (Young et al. (2013) and the time-invariant scaling scheme for spallation processes given by (Lal, 1991) and Stone (2000) were applied. The time-invariant scaling scheme does not incorporate variations in past geomagnetic field strength, but these usually only affect younger samples, at c. 10 ka, by 1% (Nishiizumi et al., 2007). The maximum deviation between different scaling schemes in this material is c. 1%, so they generally provide consistent ages and do not affect the relative chronology. We used a rock density of 2.65 g cm^3 and made no correction for potential surface erosion or snow/vegetation cover. The study area has undergone glacioisostatic uplift since the deglaciation, and this may potentially influence the ^{10}Be ages. However, as the production rate calibration dataset probably experienced a similar uplift history at our sample sites, no correction for glacioisostatic uplift is applied (cf. Young et al. 2020). Accordingly, we present ^{10}Be ages without correcting for glacioisostatic uplift, similar to most other ^{10}Be studies from Greenland. Individual ^{10}Be ages are presented with their 1-sigma analytical uncertainties, which include the uncertainty in the blank correction, the "internal" uncertainty (Table 1). When we compare our ^{10}Be ages with ^{14}C ages or climate records we include the production rate uncertainty, the "external" uncertainty (Balco et al., 2008).

Previously published ^{14}C ages have been re-calibrated using the Intcal20 calibration programme (Reimer et al., 2013). Following the procedure adopted for dates on marine shells from Greenland, ages on marine shells have been corrected with a ΔR of 0 for western Greenland and with a ΔR of -150 yr for eastern Greenland, based on dating modern pre-bomb shells (e.g. Mörner and Funder, 1990), acknowledging that significant, but unknown, changes in the reservoir effect may potentially have affected the ages especially in the turbulent millennia during the early deglaciation phases (e.g. Andrews et al., 2018).

Table 1. Summary of ^{10}Be data from South Greenland.

Sample ID	Latitude (N)	Longitude (W)	Elevation (m a.s.l.)	Sample type†	Shielding factor	Thickness (cm)	Quartz (g)	^{9}Be carrier weight (g)	^{10}Be conc. (atoms/ g)×10^4	^{10}Be uncert. (atoms/ g)×10^4	^{10}Be age (ka) internal (external) uncertainties††
Buksefjord											
X1524	63.83957	51.73826	118	bedrock	1	4.5	40.45	0.6067	6.88	0.36	14.43 ± 0.76 (1.03)
X1525	63.83970	51.73851	117	bedrock	1	5.5	33.19	0.6082	5.90	0.30	*12.48 ± 0.64 (0.88)*
X1526	63.83967	51.73839	102	bedrock	1	6	40.13	0.6086	5.59	0.26	*12.06 ± 0.57 (0.82)*
							Calculated average (number of samples out of total)				***12.3 ±0.4 (n=2/3)***
Fiskenæsset											
X1521	63.04961	50.99505	76	bedrock	0.999962	4.5	21.16	0.6068	5.93	0.43	*12.98 ± 0.95 (1.14)*
X1522	63.05008	50.99449	75	bedrock	0.999969	5.5	26.75	0.6074	5.85	0.31	*12.93 ± 0.68 (0.93)*
X1523	63.05016	50.99454	76	bedrock	0.999969	5.5	36.35	0.6049	6.31	0.36	*13.92 ± 0.81 (1.05)*
							Calculated average (number of samples out of total)				***13.3 ±0.6 (n=3/3)***
Ravns Storø											
X1519	62.71573	50.40947	193	bedrock	0.999986	7	35.09	0.6074	6.95	0.38	13.71 ± 0.76 (1.01)
X1520	62.71573	50.40947	189	bedrock	0.999986	6	45.21	0.6083	8.66	0.87	17.03 ± 1.72 (1.91)
X9364	62.71799	50.41719	209	bedrock	1	4.5	34.47	0.6092	6.26	0.37	*11.91 ± 0.70 (0.91)*
X9365	62.71770	50.41629	208	boulder	1	4.5	39.87	0.613	6.05	0.43	*11.52 ± 0.81 (0.99)*
							Calculated average (number of samples out of total)				***11.7 ±0.4 (n=2/4)***
Avigaat											
X1516	62.17882	49.80153	47	bedrock	1	7	45.06	0.6062	5.94	0.48	*13.68 ± 1.11 (1.29)*
X1517	62.17888	49.80107	44	bedrock	1	6	45.08	0.6089	5.23	0.24	*11.98 ± 0.56 (0.81)*
X1518	62.17894	49.80064	42	bedrock	1	4.5	45.26	0.608	4.55	1.10	*10.31 ± 2.49 (2.54)*
							Calculated average (number of samples out of total)				***12.0 ±1.7 (n=3/3)***
Paamiut											
X1513	61.85744	49.53121	65	bedrock	1	6	32.31	0.6111	5.57	0.29	*12.45 ± 0.66 (0.89)*
X1514	61.85734	49.53098	61	bedrock	1	6.5	25.48	0.6086	5.34	0.56	*12.05 ± 1.28 (1.40)*
X1515	61.85708	49.53045	60	bedrock	1	5.5	35.44	0.607	5.36	0.62	*12.01 ± 1.40 (1.52)*
							Calculated average (number of samples out of total)				***12.2 ±0.3 (n=3/3)***
Sermiligarsuk											
X1507	61.32122	48.86104	57	boulder	0.999672	6	33.35	0.6086	4.81	1.02	*10.86 ± 2.32 (2.38)*
X1509	61.32136	48.86013	61	bedrock	0.999704	5.5	24.76	0.5672	6.55	0.37	14.66 ± 0.84 (1.10)
									Best estimate for deglaciation age		***10.9 ±2.3 (n=1/1)***

†: All samples are coarse grained orthogneiss

†† : Italics: used in average/best estimate (see text)

^{10}Be ages were calculated using the online exposure age calculator formerly known as the CRONUS-Earthonline exposure calculator v.3 (Balco et al., 2008), the Baffin Bay production rate of 4.04 ± 0.07 at g^{-1} a^{-1} (regional SLHL) (Young et al., 2013), and the St scaling scheme (Lal, 1991; Stone, 2000) under standard atm
A rock density of 2.65 g cm^{-3} was used and we assumed zero erosion. Samples were measured using the Beryllium standard 07KNSTD (Nishiizumi et al., 2007).
Italics: ages used for average calculations

4. Results and Interpretations

As noted above a problem in using bedrock samples is the risk of inheritance in the rock surface. The knobby terrain is evidence that basal glacial erosion was not uniform, but varies in intensity even over small distances. Therefore, the amount of inheritance may also vary over short distances, resulting in a spread of old ages (Corbett et al., 2013). We therefore consider a spread of old ages as "inheritance outliers", while the mean of clustered younger ages gives the most reliable deglaciation age. Where there is no overlap between the uncertainties, we regard the youngest age as a maximum age for deglaciation. At each site our new ages are compared to previously published cosmogenic dates of deglaciation or thinning of ice streams at fjord mouths. In addition, we also show ^{14}C results on dating marine molluscs or onset of organic sedimentation in coastal lakes. Although not dating deglaciation, these dates serve as minimum constraints for local deglaciation. Much of this information has recently been reviewed by Sinclair et al. (2016). The six sites are described below, and the results are shown in Table 1 and Fig. 3.

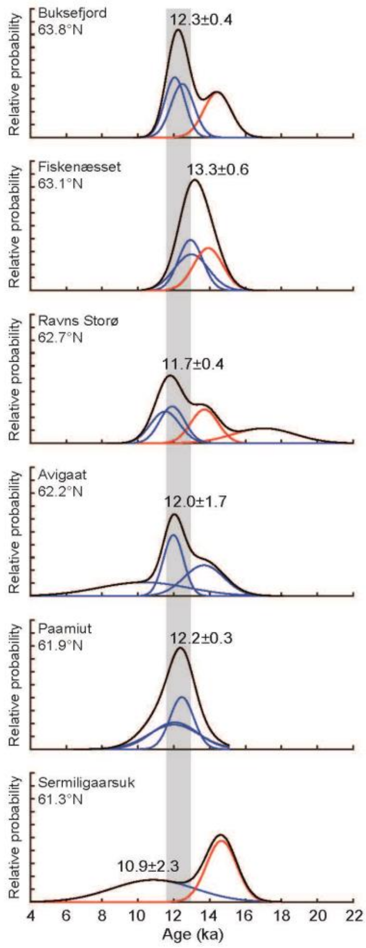


Figure 3: Normal kernel density plots for the 10Be ages from six coastal sites in SW Greenland. The mean age for each site is calculated after excluding statistical outliers (see table 1).

4.1 Buksefjord

This site is located at the outer margin of the strandflat, c. 15 km from the main coastline and midway between the mouth of Ameralikfjorden and Buksefjorden (Fig. x). The three bedrock samples from this locality were collected between 102 and 118 m a.s.l. and yielded ages of 14.4 ± 0.8 ka (X1524), 12.5 ± 0.6 ka (X1525) and 12.1 ± 0.6 ka (X1526). We interpret the oldest age as an outlier. The two youngest ages have overlapping internal uncertainty and average 12.3 ± 0.2 ka, which we interpret as the time of deglaciation at this site.

On the coastal mountains c. 10 km to the east, ^{10}Be dates of boulders from between 82 and 360 m a.s.l. gave an average deglaciation age of 10.7 ± 0.6 ka (Larsen et al., 2014). At the mouth of the Nuuk Fjord Complex, 30 km to the north, marine shells on the outer coast gave a minimum constraint for the deglaciation of 11.4 cal. ka BP (Weidick, 1976a), while ^{10}Be ages close to Nuuk showed deglaciation at c. 11 ka (Winsor et al., 2015b). This may imply that our dates here are some centuries too old, although the dates from around Nuuk Fjord indicate that the outer coast became ice free while an ice stream still occupied the Nuuk trough.

4.2. Fiskenæsset

Three bedrock samples were collected in the outer archipelago c. 6 km from the coast, from a small ice-scoured island c. 15 kilometres west of the Fiskenæsset settlement (Fig. x). The samples were collected from 75-76 m a.s.l and yielded ages of 13.0 ±1.0 ka (X1521), 12.9 ± 0.7 ka (X1522), and 13.9±0.8 ka (X1523). The average, 13.3 ±0.5 ka, is the oldest deglaciation date of our sites (Fig. 1).

These ages imply that the GrIS margin here was close to the coast prior to the YD. This result should be substantiated by other sources, but there is no available evidence from the adjacent coast to support or oppose this timing of deglaciation. Farther north, at Sermilik Fjord, ^{14}C dates of marine molluscs show that the initial marine transgression and retreat of the GrIS from the outer coast probably did not begin until a short time before 10.5 cal ka BP, and on coastal mountains nearby, ^{10}Be ages from 450 m a.s.l. show that the GrIS surface had thinned at c. 10.6 ka (Larsen et al. 2014). Even though these results come from a different trough, the difference in dates on deglaciation of the coast of 2000 years warrants confirmation. However, it should be noted that Weidick (1976b) considered the Sermilik glacier to be the last to retreat from the shelf in this part of Greenland, while the ice sheet margin both to the north and south had ice free for several millennia. In areas to the south; although no direct chronology exists to support this idea.

4.3. Ravns Storø

Four samples were collected on the island of Ravns Storø, in the middle of the archipelago, c. 5 km from the coast (Fig. x). The samples were collected between 189 and 209 m a.s.l. within a radius of 200 m. The ages show a spread of more than 5000 years: 13.7 ±0.8 ka (X1519), 17.0 ±1.7 ka (X1520), 11.9 ±0.7 ka (X9364) and 11.5 ±0.8 ka (X9365). The two youngest ages, including our only boulder sample (sample X9365), have overlapping internal uncertainties, and we consider their average, 11.7±0.2 ka, as the best estimate for the time of deglaciation at this site, while the oldest ages are outliers. From this area there is no supporting information on deglaciation history.

4.3. Avigaat

Three samples from the bedrock surface were collected from an islet in the inner archipelago, c. 3 km from the coast and the abandoned Avigaat settlement (Fig. x). The samples were taken between 42 and 47 m a.s.l. and yield ages of 13.7 ± 1.1 ka (X1516), 12.0 ± 0.5 ka (X1517) and 10.3 ± 2.5 ka (X1518). The variable ages and their uncertainties are large. Because these ages overlap, we consider the average of 12.0 ± 1.6 as the best estimate for deglaciation at this site. Some support that this might be generally correct comes from a ^{14}C age of c. 11.3 cal ka BP for basal gyttja in a lake in coastal Nerutussoq fjord to the south (Fig. x), giving a minimum age for deglaciation at this site (Kelly and Funder, 1974).

4.4. Paamiut

This site is located on a small ice-scoured island on the inner archipelago, c. 5 km from the coast and close to the mouth of Kuanersoq fjord and the town of Paamiut (Fig. x). Here, three bedrock samples between 60 and 65 m a.s.l, are dated to 12.5 ±0.7 X1513), 12.1 ±1.3 (X1514) and 12.0 ±1.4 ka (X1515). All three ages overlap within the internal uncertainty and average 12.2 ± 0.2 ka.

Around Paamiut and Kuanersoq several studies have supplied both ^{10}Be and ^{14}C deglaciation dates for the outer fjord. As could be expected, these deglaciation dates from farther inland are somewhat younger than ours. ^{10}Be dates from Kuanersoq indicate thinning of the ice margin in the fjord beginning c. 11.7 ka, and, by extrapolation, retreat from the fjord mouth at c.11.2 ka (Winsor et al., 2015b). From a ^{14}C age of 11.0 cal ka BP for basal gyttja in a lake 8 km from our samples and well below the local marine limit, Woodroffe et al. (2014) suggested that deglaciation could not have been much earlier than c. 11 ka. These results from nearby coastal localities therefore indicate deglaciation c. 1000 years later than at our site. Much of this work concerned the ice stream in Kuanersoq, while our samples come from the open coast to the south, and we suggest that an ice stream in the Kuanersoq trough remained at the inner shelf while the adjacent coastal areas became ice free.

4.5. Sermiligaarsuk

From a small island in the inner archipelago, c. 2 km from the coast and 12 km south of the Sermiligaarsuk fjord, we collected two samples from 56 and 61 m a.s.l. One sample is from bedrock (X1509) and one is from a 1-m-diameter boulder (X1507). The two samples have widely scattered ages of 10.9 ±2.3 ka (X1507, boulder) and 14.7 ±0.8 ka (X1509, bedrock) with a large uncertainty, particularly in the boulder age. The oldest age is unrealistic for deglaciation and interpreted as an inheritance outlier. The age of 10.9 ±2.3 ka (sample X1507), one of our few boulder dates, is interpreted as a closer approximation for deglaciation at this site. This age is the youngest for deglaciation of the inner shelf, but the island is also closer to the coast than any of the other sites.

Marine shells below the marine limit in the nearby outer Sermiligaarsuk Fjord have an ^{14}C age of 9.7 cal ka BP, providing a minimum for deglaciation at this site (Weidick et al., 2004).

5. Discussion

A complication in using bedrock samples in exposure dating is the chance of inheritance in the rock surface. The knobby terrain in SW Greenland is evidence that glacial erosion was not uniform, but varied in intensity even over small distances. Therefore, the amount of inheritance may also vary over short distances, resulting in an age dataset that includes old ages (Corbett et al., 2013). We therefore consider anomalously old ages as “inheritance outliers”, while the mean of clustered younger ages gives a reliable deglaciation age.

5.1. Overview of results from Southwest Greenland

According to the criteria outlined above, two of the sites, Paamiut 12.2 ± 0.2 ka; n=3) and Fiskenæsset (13.3 ±0.5 ka; n=3), contain no obvious outliers, hence no obvious inheritance, implying deglaciation ages during the middle YD (Paamiut) and prior to the YD (Fiskenæsset). Also, Avigaat has overlapping uncertainties, indicating deglaciation in late

YD times, but with a large uncertainty. At two sites, Buksefjord and Ravns Storø, one or two samples in each are interpreted as being affected by inheritance, but the remaining clusters indicate deglaciation during middle to late YD. Finally, at Sermiligarssuk only one sample is considered free of inheritance, yielding a best estimate for deglaciation in the Early Holocene. From this, the results, although affected by inheritance, would point to deglaciation on the inner shelf in this part of Greenland at varying times between the late Allerød and the early Holocene. However, at some sites the ages are significantly older than expected from a comparison with previous dating of deglaciation at the adjacent coast.

At Paamiut, Fiskenæsset and Buksefjord, our ages are up to 2 ka older than deglaciation ages obtained at nearby fjord mouths. A possible explanation may be that while the coastal areas became ice free, ice lingered in the major troughs, not reaching the inner shelf until the Early Holocene, as shown previously for ice streams in Disko Bugt (Jennings et al., 2014) and suggested by Weidick (1976b) for our area in SW Greenland.

However, the offset in deglaciation ages, especially the oldest from Fiskenæsset, could also be impacted by a small amount of uniform inheritance in the bedrock, as demonstrated at Utsira, Norway (Briner et al., 2016). Uniform inheritance may influence the mean age from a cluster, meaning that several bedrock samples from adjacent sites could all have experienced a similar amount of inheritance. This effect is considered to particularly affect bedrock and boulders in areas that experienced long ice-free periods between brief maximum glacial phases (Briner et al. 2016). The landscape in the coastal archipelago is the result of intense erosion by warm based ice, probably back through several ice ages and during the better part of the last Ice Age (e.g. Nielsen and Kuijpers, 2013; Seidenkrantz et al., 2019), and we consider the type of deep, uniform inheritance as described by Briner et al. (2016) to be unlikely in our samples.

A possibly more likely type of uniform inheritance could be if the ice margin readvanced, but failed to erode the bedrock deeply enough to remove the ^{10}Be signal from previous exposure. ,This could have happened during a YD readvance, as suggested for this area by Weidick et al. (2004) and Lecavalier et al. (2014). However, independent dating is required to show if any of these potential errors have affected the ages, especially those from Fiskenæsset.

In summary, the ^{10}Be dates from the coast of SW Greenland – although affected by inheritance – suggest that the ice sheet margin retreated on the inner shelf close to the coast at least since mid/late YD times. Some dates are very old, compared to deglaciation dates on the coast. This could be due to differential ice margin behaviour in and away from troughs or – in the case of the oldest age - to ice margin readvance over ice free land.

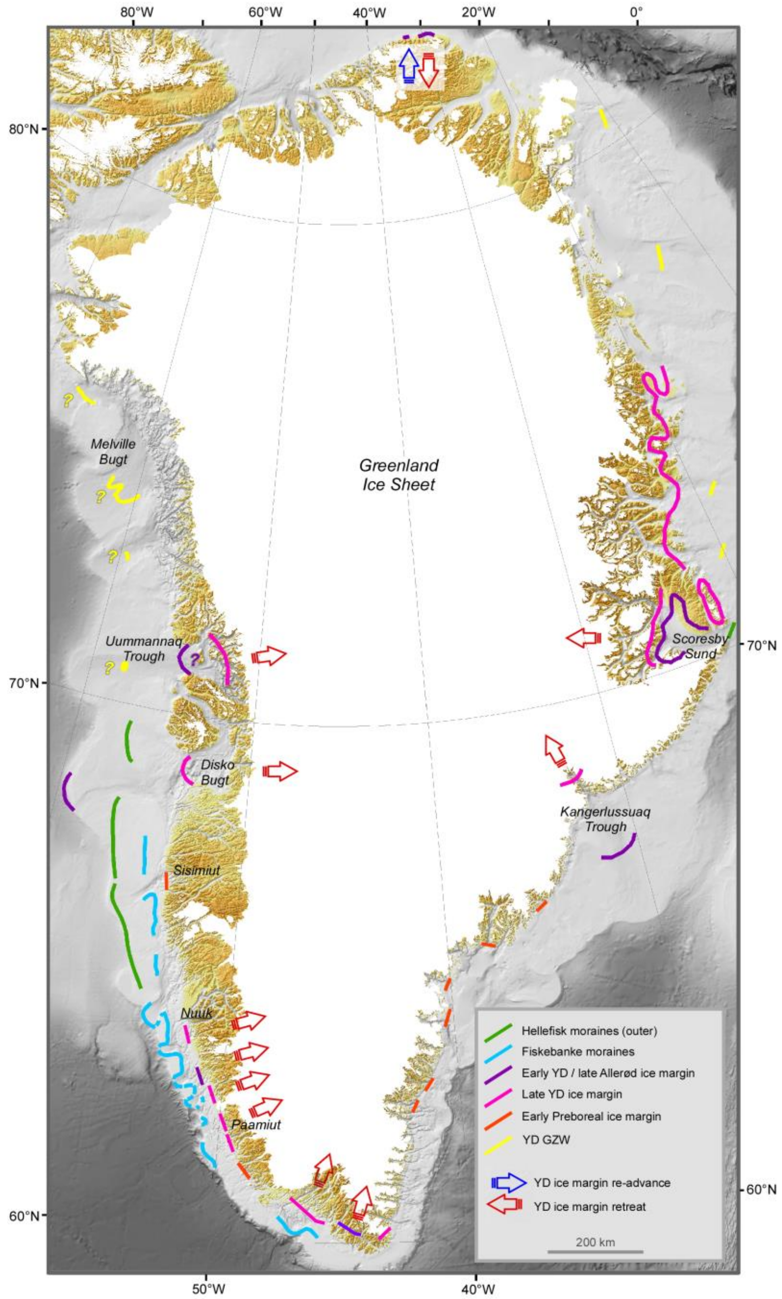


Figure 4: Deglacial ice-margin features in Greenland discussed in the text. (Question marks at GZWs apply to age and come from the original literature). Sources Larsen et al. (2016) and references discussed in the text. Background map of Greenland and surrounding seas from BedMachine Greenland v.3 (Morlighem et al. 2017)

5.2. YD ice-margins in Greenland

From a recent review of YD ice-margins in Greenland, Larsen et al. (2016) concluded that ice-margin retreat indeed characterised most areas with a dated record going back through or at least into the YD. This is well constrained by ^{10}Be dating in coastal areas and ^{14}C-dated marine sediment cores in and outside major cross-shelf troughs, and applies to areas in western, eastern and southernmost Greenland (Fig 4). More recently, GrIS retreat during the YD has been corroborated from the Disko Bugt shelf (Hogan et al., 2016; Oksman et al., 2017), and from east and south Greenland (Levy et al., 2016; Andrews et al., 2018; Dyke et al., 2018; Rainsley et al., 2018).

In available records, the most dramatic retreat occurred on the shelf at Disko Bugt, where the ice stream apparently retreated over more than 200 km from an Allerød-Early YD position near the shelf break (Fig. 4), but the retreat was punctuated by periods of topographically conditioned stillstand and a spectacular, but brief, non-climatic readvance (e.g. O'Cofaigh et al., 2013; Hogan et al., 2016). A similar history is recorded on the east side of Greenland (Fig. 4), in Kangerlussuaq trough and fjord, where not only the shelf, but also most of the fjord became ice free during YD (e.g. Andrews et al., 2018). It is noteworthy that in these areas, as well as other areas with a record going back to the Allerød (Scoresby Sund, southernmost Greenland), there is no evidence for marked ice margin response to the initial YD cooling. In these two areas, the GrIS had retreated behind the present coastline before YD (e.g. Björck et al., 2002; Larsen et al., 2016; Levy et al., 2016; Levy et al., 2020). Only on the north coast of Greenland did glaciers apparently advance/retreat at the beginning/end of YD (Larsen et al., 2016).

In areas where the dated record goes back only to the middle YD, such as most of our area, or areas where the ice remained on the shelf until early Preboreal times, as shown by recent cosmogenic dates from coastal southeast Greenland and the Sisimiut area (Fig. 4) (Dyke et al., 2018; Rainsley et al., 2018; Levy et al., 2020; Young et al., 2020), the ice margin retreat may be seen as a response to the slow warming in the latter portion of the YD (Vacco et al., 2009; Buizert et al., 2018). Therefore, while the dated records going back to Allerød times do not show evidence of ice margin readvance/stillstand at the initial YD cooling, a readvance may have occurred in other areas, considering the large variation in YD ice margin behaviour seen in the dated records. Below we discuss evidence, landforms on the shelf, which has been attributed to such a YD ice margin readvance/stillstand.

5.3 YD readvance on the shelf? – moraines and GZWs

Since their first discovery, the Hellefisk and Fiskebanke moraines on the west Greenland shelf have played a prominent role in the discussion of early deglaciation history (Fig.4) (e.g. Kelly, 1985; Funder et al., 2011; Hogan et al., 2016). The outermost and oldest, the Hellefisk moraine system, runs along the shelf break for 200 km at a depth of c. 200 m, c. 120 km from the coast and consists of swarms of up to 100-m-high ridges (Brett and Zarudski, 1979). To the east of this, halfway towards the coast, the younger Fiskebanke moraines impinge on the inner side of the fishing banks c. 40 km from the coast. These are composed of single ridges, which occur on and off, on the inner banks and along the sides of transverse troughs for a distance of c. 500 km along the coast (Fig. 4). Although undated, the two moraine systems have generally been regarded as climate signals for two distinct periods of cooling – either Saalian and LGM (Funder et al., 2011) or LGM and YD (van Tatenhove et al., 1996; Roberts et al., 2009; Simpson et al., 2009; Lecavalier et al., 2014).

Recently, for the first time, absolute ages have been supplied for parts of the Hellefisk moraines, where Hogan et al. (2016), from ^{14}C dates in marine cores, found that in their study area south of Disko Bugt, these moraines represented a deglacial topographically controlled calving bay, dated to c. 12.2 ka BP. In contrast to this, from a marine core at the shelf edge off Nuuk, Seidenkrantz et al. (2019) found that the outer Hellefisk moraine here dated to c. 60 ka. These results imply that the outer Hellefisk moraines are not synchronous, but have widely different ages, and are to some extent controlled not by climate but by topography. Extending for 500 km along the coast the Fiskebanke moraines are the most compelling evidence for ice margin response to YD cooling in Greenland – if the can be dated to YD. The results from Hellefisk moraines may also cast some doubt on the climatic significance of the Fiskebanke moraines. Their affinity to the inner shelf trough and transverse troughs could indicate topographic, rather than climatic, control. As noted above our dates from Fiskenæsset, well behind the moraines, could be interpreted either in favour of or against a YD readvance to the moraines. This stresses the need for climate-independent dating of this important event.

Other geomorphic evidence for YD readvance or long-lasting stillstand has recently been suggested from several major transect troughs. This is based on high-resolution multibeam bathymetry, revealing a large variety of glacial landforms that formed during retreat of major ice streams. Notably, the occurrence of large GZWs has been suggested to reflect long lasting stillstand on mid-shelf (see references below). GZWs are wedge-shaped sediment accumulations deposited at the front of an ice stream during a period of stability (e.g. Dowdeswell and Fugelli, 2012). GZWs have now been observed in most investigated troughs around Greenland, and, although not dated, prominent GZWs have tentatively been assigned to the YD based on the assumption that they correlate with cold periods in the ice core temperature record (Sheldon et al., 2016; Slabon et al., 2016; Arndt et al., 2017; Newton et al., 2017; Arndt, 2018).

In the Uummannaq trough deglaciation of the shelf began before 15 ka and by c. 11.5 cal. ka BP the large ice stream in the trough had disintegrated into fjord glaciers with their front close to the present ice-margin (e.g. Jennings et al., 2017). However, there are two very different views on what happened in the intervening 3.5 millennia. Based on exposure ages on coastal mountains and ^{14}C dates in the fjords Roberts et al. (2013) found that the large Uummannaq ice stream had retreated from the trough and into the fjords during YD, controlled by topography and bathymetry (Fig. 4). In contrast, Sheldon et al. (2016), from a series of marine cores and a prominent GZW in the transect trough, suggested that the ice stream was stabilised for 2 ka since Allerød times on the outer shelf, 150 km further away from the coast (Fig. 4). This was based on extrapolation from a ^{14}C age and correlation with the ice core temperature record.

An even larger discrepancy between the two dating approaches is seen in northeast and east Greenland, where Arndt et al. (2017) and Arndt (2018), from multibeam bathymetry interpreted lineaments, mid shelf GZWs and a moraine at the mouth of Scoresby Sund as evidence for readvance of fast flowing ice streams in major troughs along the northern east coast of Greenland (Fig. 4). These features were dated to YD by climatic inference. This overlooks earlier work from land, especially in Scoresby Sund. Here Greenland's highest concentration of ^{10}Be and ^{14}C dates show that the outlet glaciers in this fjord system had retreated into the fjord in Allerød times, forming a swarm of moraines from Allerød through YD and into the Preboreal (Denton et al., 2005; Kelly et al., 2008; Hall et al., 2010; Vasskog et al., 2015; Levy et al., 2016), ruling out the possibility of a major YD readvance to or on the shelf,

Similar outbursts of fast flowing ice streams, reaching mid shelf GZWs were recorded also further north in transect troughs at Kong Oscar Fjord, Kejser Franz Joseph Fjord and the wide shelf of Northeast Greenland, and dated by the same means to YD (Arndt et al., 2017; Arndt, 2018). Also here it has been overlooked that mid fjord moraines in these fjords, 100 km behind the mid shelf GZWs, previously have been dated to late YD/earliest Preboreal – after calibration of the ^{14}C ages (Hjort, 1979). Reconciling these two datasets would imply an extraordinarily dynamic behaviour of the ice-margin along the East Greenland seaboard, with both advances and retreats of more than 100 km within YD, in a period with increased sea ice along the coast (Flückiger et al., 2008; Buizert et al., 2018).,

In Northwest Greenland mid shelf GZWs with a length of more than 100 km have been recorded in transect troughs on the shelf in Melville Bugt (Slabon et al., 2016; Newton et al., 2017). From analogy with the "climate-correlated" Uummannaq GZW they were tentatively referred to YD, although non-climatic, bathymetric conditions may also have determined their position (Newton et al., 2017).

In summary, some of the landforms on the shelf, which, on climatic grounds, have been attributed to YD ice margin readvance/stillstand apparently do not date from YD, and surprisingly, none of the ^{14}C and ^{10}Be dated records show evidence for ice margin response to initial YD cooling (Fig. 4). This highlights the need for climate-independent dating of the submarine landforms and GZWs, to exploit this rich source of information, and get a better understanding of the ice sheet/climate relation.

5.5. Ice-margin retreat during the YD cold oscillation?

To explain the mismatch between YD cooling, and apparent ice-margin retreat, two agents have especially been called on: advection of warm oceanic subsurface water to the ice-margin, and increased climatic seasonality.

In both cases the sequence of events begins with increased production of meltwater around the North Atlantic during the Allerød warm period. The fresher and lighter water eventually sealed off the Atlantic surface circulation from the atmosphere and impeded AMOC. However, warm water from the subtropical areas was still driven into the North Atlantic, but now as subsurface currents (Marcott et al., 2011; Ezat et al., 2014). The subsurface water followed the path of the present North Atlantic surface circulation in the Irminger Current running south along Southeast Greenland, continuing around Greenland's southern tip, and heading northwards as the West Greenland Current (Fig. 1). Along the Greenland shelf the warm Atlantic subsurface water was present and caused ice-margin retreat in Southeast and West Greenland at 15-16 cal. ka BP, and it was continuously present at the southeast Greenland shelf edge through Bølling-Allerød and YD times (Kuijpers et al., 2003; Knutz et al., 2011; Jennings et al., 2017; Andrews et al., 2018).

Today warm subsurface water from these currents, below a cap of fresher water, causes extensive melting of floating outlet glaciers in Greenland (e.g. Mayer et al., 2000; Motyka et al., 2011), and during the early phase of deglaciation when the GrIS had its entire margin on the shelf it was especially sensitive to the advection of warm subsurface water, causing ice-margin retreat even when temperatures were dropping as discussed extensively in the literature (Kuijpers et al., 2003; Jennings et al., 2006; Knutz et al., 2011; Rinterknecht et al., 2014; Winsor et al., 2015b; Sheldon et al., 2016; Sinclair et al., 2016; Jennings et al., 2017; Oksman et al., 2017; Andrews et al., 2018; Dyke et al., 2018; Rainsley et al., 2018).

Crucial to the impact of the warm water is the depth of the grounding line at the ice-margin, and the accessibility for the warm subsurface water. This is again dependent on local bathymetry and – in the troughs – of the type of connection to the open ocean, which in each area may control the impact of the warm water on the ice-margin, as well as the impact from changing sea level and presence or absence of buttressing sea ice. This may explain why the deglaciation of the shelf and troughs had such a different character even between neighbouring troughs, such as between the rapid deglaciation of the Kangerlussuaq trough and the much slower deglaciation along the adjacent coast to the south (Dyke et al., 2018). It may also explain the differences in the timing of onset of deglaciation in the neighbouring Nuussuaq and Disko troughs in West Greenland as discussed by Jennings et al. (2017).

Also increased seasonality owes to the meltwater cap over the North Atlantic, and reduced AMOC. The fresher meltwater-diluted water seals off the ocean surface and, especially in winter, cuts off the ocean/atmosphere exchange (Denton et al., 2005; Hall et al., 2008; Vacco et al., 2009; 2010; Buizert et al., 2014; Levy et al., 2016; Buizert et al., 2018). This results in very cold and arid winters and increase in extent and duration of sea ice, while summer temperatures – which primarily determines a glacier's mass balance - are less affected and may even warm up (Björck et al., 2002).

Vacco et al. (2009) presented a glaciological model for warm based glacires on land, tuned with a temperature record from a Greenland ice core, showing that for areas with high amplitude change between Allerød and YD moraine deposition would be expected in the beginning of YD followed by recession of the ice margin in the mid and late YD. This may explain why major YD moraines in Greenland have been seen only in northernmost Greenland and Scoresby Sund – two areas where the glaciers had become landlocked already before YD, while the ice margin in most other areas retreated on the shelf, controlled to a large extent by warm subsurface water.

The importance of changing seasonality has recently been investigated in a model where the deglacial ice-core temperature records in three ice cores are combined with simulated seasonal air temperatures for the whole of Greenland, enabling assessment of variations in seasonality in time and space (Buizert et al., 2018). The varied seasonality model deviates from the previous Huy models (e.g. Lecavalier et al., 2014), indicating that the dramatic temperature changes in the ice core records at the beginning and end of YD, are muted in the varied seasonality model.

In summary, the apparent contradiction between ice core temperature records, where temperatures dropped dramatically at the onset of YD, and the dated glacial record where glaciers in many parts retreated, may be explained by the effect of warm subsurface water on the ice-margin, which was, all around Greenland, located on the shelf during the early phase of deglaciation, while local topographic and bathymetric conditions controlled the access of warm water to the ice-margin. Increased seasonality owes to increase in distribution and duration of winter sea ice, and the YD temperature drop in the ice cores was due to a large extent to lowering of winter temperatures with little impact on the ice-margin, but a large effect on distribution, thickness and duration of sea ice. This may also explain why neither the rapid initial YD cooling nor the abrupt warming at the end left a clear track in the dated records.

6. Conclusions

^{10}Be dates on bedrock surfaces in the glacially eroded archipelago on the inner shelf of southwestern Greenland are affected by ^{10}Be inherited from earlier exposure, but clustering of ages from each site suggest that the ice-margin here was retreating and close to the coast at least since mid YD times.

A survey of ^{10}Be and ^{14}C dated records, going back through YD elsewhere in Greenland - south, east and west - shows that also here the ice-margin was retreating, but with large differences in speed and extent even between neighbouring basins, probably controlled by local topography and trough geometry.

While the retreat in our and in other part dates from mid or late YD and may be seen as retreat from an initial YD readvance, areas with a record that goes back into the Allerød show no evidence for ice margin readvance at the initial YD transition. Only in northernmost Greenland did glaciers from a local ice cap apparently advance/retreat at the beginning/end of YD.

Moraines and GZWs on the shelf, which have been attributed to YD readvance, are dated by climatic inference, and need climate-independent dating to distinguish climate- from non climatic factors.

The apparent mismatch between the ice core temperature record and the ice-margin may be explained by the circumstance that during LGM the GrIS, contrary to other large ice sheets, around the whole perimeter was standing on the shelf and especially sensitive to changes in ocean currents, sea level, and sea ice distribution and thickness.

Recently, high resolution bathymetry has supplied a wealth of data on ice stream dynamics during deglaciation. However, this evidence is essentially dated only by climate-inference. To tap this rich source of information and get a better understanding of the ice sheet/climate relation, climate-independent dating of the submarine features is badly needed.

We subscribe to the contention by Andrews et al. (2018, p. 16): "The use of the GrIS's isotopic records as a one-to-one template for coeval changes in glacier and ocean response potentially ignores the different response timescales between the atmosphere, oceans and cryosphere" – with a bearing also on the future.

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