Quaternary Science Reviews 258 (2021) 106883

Contents lists available at ScienceDirect

Quaternary Science Reviews

journal homepage: www.elsevier.com/locate/quascirev

Holocene glacial history of Renland Ice Cap, East Greenland, reconstructed from lake sediments

Aaron K. Medford [a], Brenda L. Hall [a, *], Thomas V. Lowell [b], Meredith A. Kelly [c], Laura B. Levy [d], Paul S. Wilcox [e], Yarrow Axford [f]

[a] School of Earth and Climate Sciences and the Climate Change Institute, University of Maine, 5790 Bryand Global Sciences Center, Orono, ME, 04469, United States
[b] Department of Geology, University of Cincinnati, 500 Geophysics Building, Cincinnati, OH, 45221, United States
[c] Department of Earth Sciences, Dartmouth College, HB 6105 Fairchild Hall, Hanover, NH, 03755, United States
[d] Humboldt State University, 1 Harpst Street, Arcata, CA95521, United States
[e] Institute of Geology, University of Innsbruck, 52 Innrain, Innsbruck, Austria, 6020, United States
[f] Department of Earth and Planetary Science, Northwestern University, 2145 Sheridan Road, Evanston, IL, 60208, United States

ARTICLE INFO

Article history:
Received 2 December 2020
Received in revised form
23 February 2021
Accepted 1 March 2021
Available online xxx

Handling Editor: Dr C. O'Cofaigh

Keywords:
Greenland
Renland ice cap
Holocene
Glacier variations
Lake sediments

ABSTRACT

Shrinking glaciers, melting permafrost, and reduced sea ice all indicate rapid contraction of the Arctic cryosphere in response to present-day climate warming, a trajectory that is expected to continue, if not accelerate. The reaction of the Arctic cryosphere to past periods of climate variation can afford insight into its present and future behavior. Here, we examine a ~12,000 year record of glacier fluctuations and meltwater variation associated with the Renland Ice Cap, East Greenland, that extends from the early Holocene thermal optimum through the cooling of the Little Ice Age to present. Sediment records from glacially fed lakes indicate rapid early Holocene deglaciation, with ice extent likely slightly smaller than at present by ~9500 yr BP. Glacial activity resulted in occasional deposition of rock flour in the studied lakes in the early Holocene until at least ~7500 yr BP. Rock flour is absent for much of the period ~7000-4000 yr BP, suggesting ice extent generally was smaller than at present. However, thin layers of blue-gray clay throughout this period may indicate millennial-scale ice expansions, with Renland Ice Cap briefly reaching extents during cold phases that may have been similar to today. Glacial sediment deposition occurred again in the late Holocene at ~3200–3400 yr BP and was followed by a brief glacial episode at ~1340 yr BP and then a major event beginning shortly after ~1050 yr BP. We infer that rock flour deposition in the lakes in the last millennium corresponds with advance of Renland glaciers to their Little Ice Age positions, marked by a fresh, gray drift limit. Radiocarbon dates of *in situ* plant remains adjacent to the present ice cap indicate a short relatively warm period ~500 yr ago, when ice was within its AD 2011 limit, followed by glacier readvance. The general pattern of ice fluctuations in Renland is similar to that at other ice caps in the region, but also has important differences, including the preservation of a possible mid-Holocene record at times when lower-elevation ice caps in the Scoresby Sund region may have been absent. This finding reinforces the concept that examination of multiple geographic and geomorphologic settings is necessary for a full understanding of ice variations in a region.

1. Introduction

Rising temperatures, coupled with multiple positive feedbacks (Serreze and Barry, 2011), have caused the Arctic cryosphere to undergo contraction in recent decades (Jorgenson et al., 2006; Maslanik et al., 2007; Kwok et al., 2009; Velicogna, 2009). Prediction of the response of this region to future climate change requires not only an understanding of the impact of anthropogenic forcing, but also knowledge of the frequency, magnitude, and geographic expression of natural climate cycles and their effect on the cryosphere on both short and long timescales. Insight into this natural climate variability can come from high-resolution records of past climate change, which also will allow us to assess whether present-day changes are unique. The Holocene affords a valuable analogue

* Corresponding author.
E-mail address: brendah@maine.edu (B.L. Hall).

https://doi.org/10.1016/j.quascirev.2021.106883

for future climate and cryosphere behavior, because some periods were warmer than present in the Arctic (Kaufman et al., 2004; Axford et al., 2020). Outstanding questions include the nature and origin of a proposed millennial-scale pulsebeat in climate change (Denton and Karlén, 1973), whether or not there is synchrony among climate variations in different parts of the Arctic (Kaufman et al., 2004), and the degree to which current rapid ice loss was preconditioned by an extensive advance that culminated in the late 19th century.

Small, local ice caps are sensitive recorders of past summer air temperature, making them ideal for studying natural climate variability (Oerlemans, 2001; Alley, 2003; Zemp et al., 2015). Changes in ice extent, inferred either from landforms or properties of sediments in glacially fed lakes, reflect snowline variations and afford at least qualitative, and in some places quantitative, estimates of past temperatures (Dahl et al., 2003). Here, in order to document past ice fluctuations and to afford data with which to test hypotheses concerning the spatial and temporal pattern of climate cycles in the Arctic, we reconstructed variations in the past extent of the southern margin of Renland Ice Cap, East Greenland, using a combination of glacial mapping and analysis of sediments from glacially fed lakes (Dahl et al., 2003; Bakke et al., 2005; Briner et al., 2010; Levy et al., 2014). By combining our data with other glacial reconstructions from East Greenland (Wagner and Melles, 2002; Lowell et al., 2013; Levy et al., 2014; Balascio et al., 2015; Lusas et al., 2017; Adamson et al., 2019), as well as with nearby paleoclimate proxies from ice cores and lake sediments (Johnsen et al., 1992; Vinther et al., 2008; Axford et al., 2017; Simonsen et al., 2019), we provide a regional view of the timing, pattern, and spatial extent of Holocene climate and glacial events along the western boundary of the North Atlantic Ocean.

2. Study area

Scoresby Sund (~69–72°N, 21–30°W), the largest fjord system in East Greenland, drains the Greenland Ice Sheet (GIS) through some of the deepest fjords in the world (Fig. 1). Its location adjacent to the North Atlantic, as well as to the cold East Greenland Current, is ideal for examining the role of the ocean in Holocene climate variation (e.g., Bond et al., 2001; Denton and Broecker, 2008). An additional advantage of this field area is that geomorphological and sedimentological records of former glacier extent can be compared directly to climate data from the nearby Renland Ice Cap (Vinther et al., 2008; Simonsen et al., 2019), as well as from the Summit ice cores (O'Brien et al., 1995) on the GIS. Comparison of different types of proxies has the potential to yield information about seasonality during past climate variations (Denton et al., 2005; Buizert et al., 2014).

The Scoresby Sund region today displays strong gradients in elevation, temperature, and precipitation. The East Greenland Current keeps the climate cool and damp along the coast, where mean-annual temperature at Ittoqqortormiit near the mouth of the sound (~200 km from the field area) is -9 °C (http://www.worldclimateguide.co.uk/guides/greenland/ittoqqortoormiit/). Climate inland is warmer and drier (Funder, 1978). Small, independent ice caps and alpine glaciers occur throughout the region in mountains and on plateaus between fjords. The focus of this study, the southwest Renland plateau (~70.5–71.5°N, 24–28°W; Fig. 2), has gently undulating topography which reaches 2340 m elevation. Renland Ice Cap (~1200 km^2) covers much of the plateau at present.

3. Methods

We cored three lakes near the southwest margin of the ice cap (Bunny, Rapids, and Raven, all informal names; Table S1; Fig. 2), two of which are glacially fed at present and one of which is not. The latter is a 'control lake' (i.e., Dahl et al., 2003), the purpose of which is to document any inorganic sedimentary layers produced by non-glacial means (e.g., by severe precipitation events or widespread cryoturbation). The existence of such layers in the control lake would alert us that similar inorganic sediments in the glacial lakes

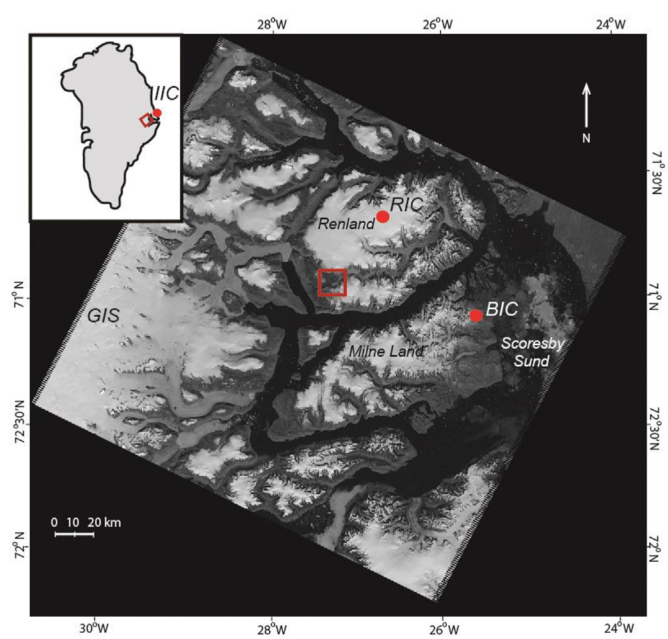

Fig. 1. ASTER image of western Scoresby Sund. Color circles show the locations of the Renland ice cap (RIC), the Bregne ice cap (BIC) in eastern Milne Land, and Istorvet ice cap (IIC) in Liverpool Land (on Greenland inset map). Box on main map outlines the Renland field area. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

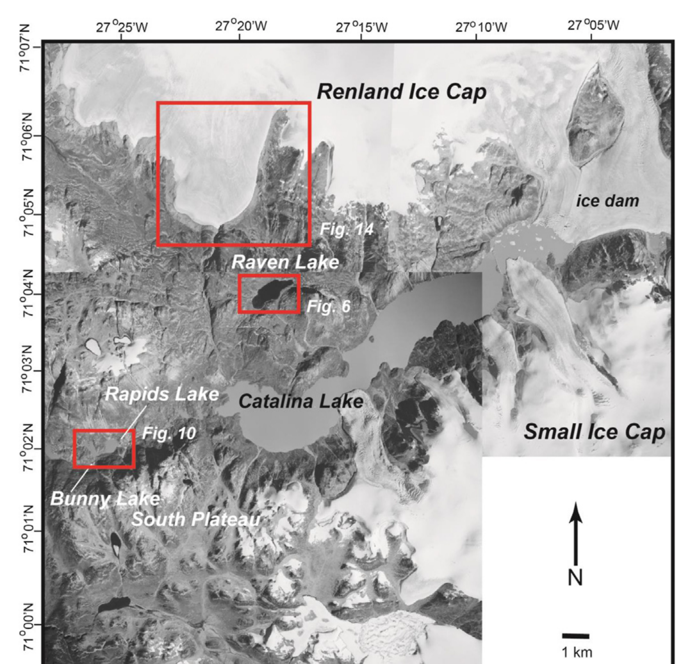

Fig. 2. Satellite image of the Renland field area, showing the relationship between local ice caps and Catalina Lake, as well as lakes cored as part of this study.

cannot be interpreted simply as a result of glacial activity. We constructed bathymetric maps using a Humminbird depth sounder and deployed a modified Bolivian piston coring system from a raft to collect multiple cores from each lake.

We split and cleaned each core and measured magnetic susceptibility (MS) with a Geotek MSCL-XYZ Core Scanner and performed initial stratigraphic analysis at 0.5 cm resolution. Stratigraphic information for all cores are in the Supplemental Information. We selected the longest and/or most complete stratigraphic record in each basin to serve as a master core. For these master cores, we measured percent loss on ignition (LOI) by burning at 550 °C (Bengtsson and Enell, 1986) and grain size using a Coulter counter.

We prepared 40 samples for radiocarbon dating, preferentially selecting terrestrial plant macrofossils. If insufficient macrofossils were visible (such as in some of the gyttja), we washed the minimum amount of organic-rich sediment through a 63 μm sieve necessary to isolate sufficient organic fragments for dating. Inspection suggests most were of terrestrial material, although those from Raven Lake were too small to be identified. We did not observe any contaminants, either surficial contaminants (i.e., bacterial growth) or reworked carbon-rich materials (i.e., coal). All samples were washed with ultrapure water, dried, and sent to the National Ocean Sciences Accelerator Mass Spectrometry Facility for analysis. We calibrated dates using CALIB 8.2 (Stuiver and Reimer, 1993; Stuiver et al., 2021) and the INTCAL20 dataset (Reimer et al., 2020). All dates in text are quoted in calendar years with a 2-sigma error. We did not include a reservoir correction, and none is anticipated. Bedrock surrounding the lakes lacks carbonate material and other ancient carbon sources. Moreover, water in Bunny and Rapids Lakes has a low residence time due to river through-flow. Lusas et al. (2017) obtained a modern age from aquatic algae at the sediment-water interface from a similar lake near the mouth of Scoresby Sund (Fig. 1), suggesting little to no reservoir effect. In addition to the samples from the lake cores, we also dated 15 samples of *in situ* plant materials, all terrestrial moss or *Salix*, that were buried by glacial advance near the present Renland Ice Cap margin and re-exposed by melting ice just prior to 2011.

4. Local geography

The geography of the field area (Figs. 1 and 2) is critical for interpreting sediments in the studied lakes. Meltwater from Renland Ice Cap flows south into a large east-west trending ice-dammed lake (Catalina Lake #1, hereafter referred to simply as 'Catalina Lake, ' informal name) that occupies Catalinadal, a fault-controlled trough. This trough drops sharply in elevation to the east and, in multiple places, is crossed by outlet glaciers from both Renland and adjacent ice caps. These glaciers today dam two other lakes (Catalina Lakes #2 and #3, from west to east). The easternmost ice dam, from a lobe of Edward Bailey Glacier, dams Catalina Lake #3, which is known to drain catastrophically to the east when the dam is compromised (Fig. 3; Grinsted et al., 2017).

At present, an outlet of Renland Ice Cap crosses Catalinadal and dams the east end of Catalina Lake, preventing its natural overflow to the east. Thus, drainage today is to the west across a sill and subsequently through a chain of small lakes, including two of our studied lakes, Rapids and Bunny. Because of this geographic situation, any time Catalina Lake drained to the west in the past, ice extent must have been similar to or greater than that of today. Shrinkage of Renland Ice Cap to a position slightly smaller than at present would allow resumption of the eastern drainage. This resumption could be gradual or catastrophic. The third lake in our study, Raven, is on a shoulder of Catalinadal (Fig. 2). It does not receive glacial meltwater at present.

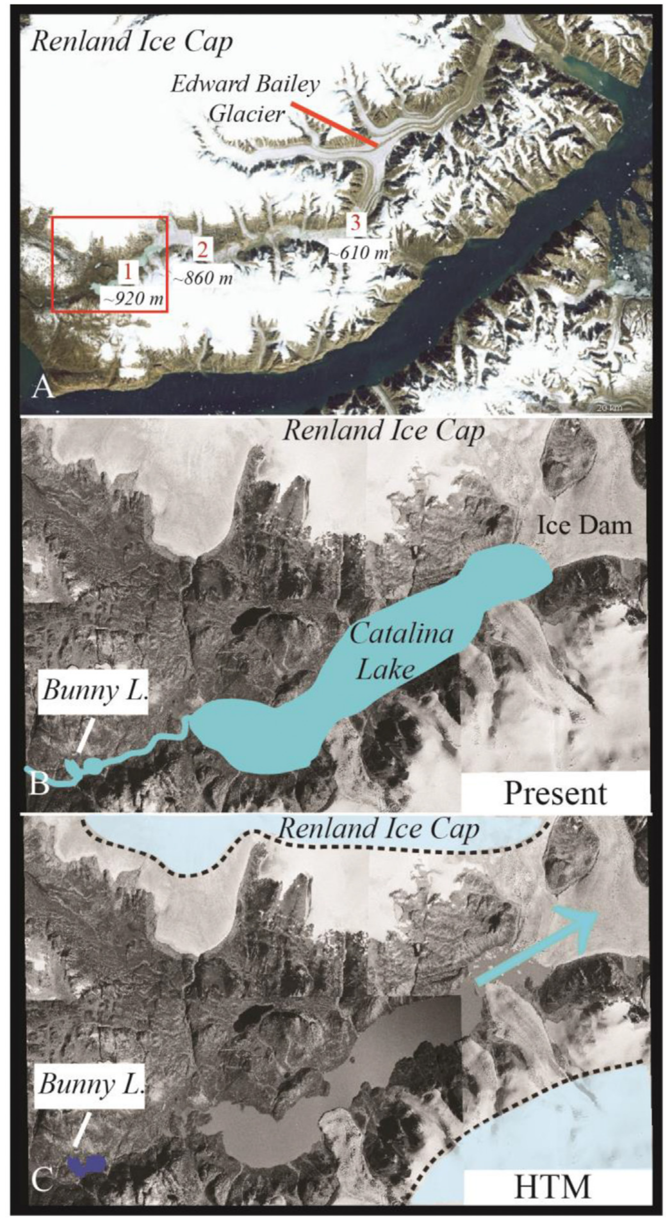

Fig. 3. Google Earth imagery showing geographic setting of the field area and a schematic representation of Catalina Lake drainage. A. Location of the field area (box) with the position of Catalina Lake (1) marked, as well as two other lakes (2, 3) in the same drainage and their approximate elevations (2021). Failure or retreat of ice dams along this drainage periodically results in outburst floods to the east (Grinsted et al., 2017). B. Current situation in the field area, where ice dams Catalina Lake, promoting westward drainage through Bunny Lake and deposition of rock flour. C. Proposed situation during the Holocene thermal maximum (HTM) with loss of the ice dam resulting in drainage of Catalina Lake to the east and isolation of Bunny Lake (dark blue). HTM ice margin (dashed) is schematic only, as the actual margin configuration at that time is unknown. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

Terrain within the field area consists of rugged bedrock draped primarily with thin, lichen-covered till. However, within 200–500 m of the present-day glaciers, the drift is gray, unweathered, and lichen-free (Fig. 4A and B); by correlation with deposits elsewhere in the region (Hall et al., 2008a; Lowell et al., 2013; Levy et al., 2014; Lusas et al., 2017), we infer that the unweathered drift represents the late Holocene ("Historical"; Weidick, 1968) ice extent. Similar lichen-free zones occur on the

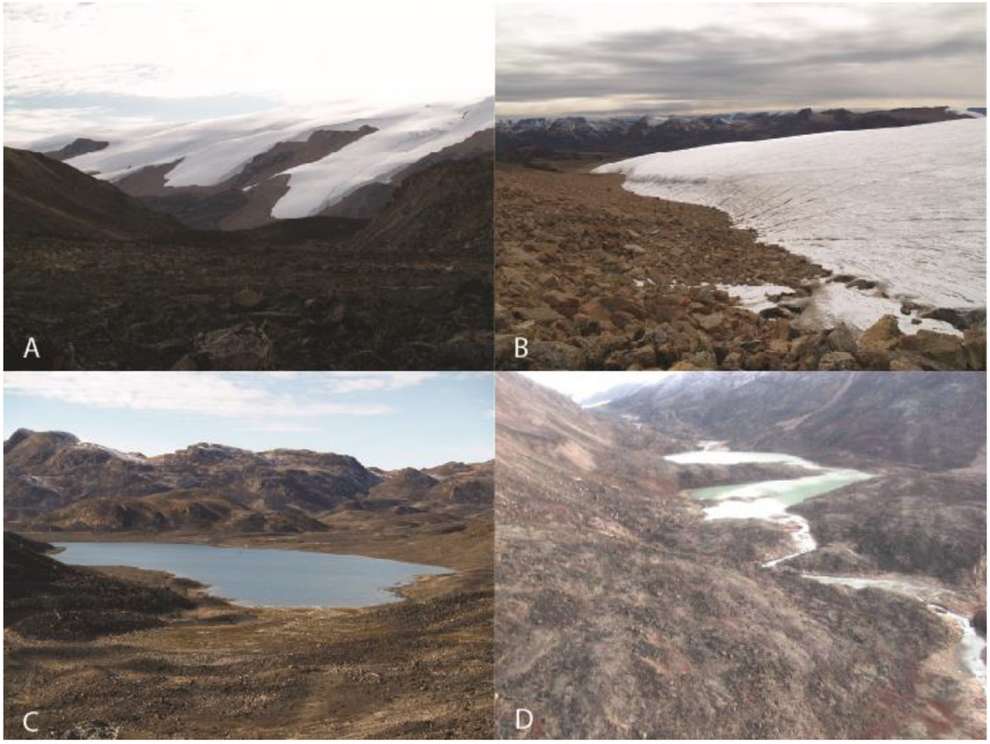

Fig. 4. (A) Photograph of the ice cap south of Catalina Lake. The gray color adjacent to the ice margin marks the inferred late Holocene drift limit. (B) Margin of the Renland ice cap with view to the west. Drift adjacent to the ice margin is unweathered and relatively lichen-free. We infer it dates to the last expansion of the ice cap in the late Holocene. (C) Photograph of Raven Lake with view to the northwest. (D) Aerial view of Bunny (foreground) and Rapids (background) Lakes with view to the east. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

mostly ice-free plateaus above Bunny and Rapids Lakes and also may represent areas of ice or permanent snow cover during the late Holocene.

5. Raven Lake

5.1. Setting and geomorphology

Raven Lake (0.3 km^2, Figs. 2,4C) consists of a relatively large, flat basin with two minor sub-basins reaching depths of ~20 m (Fig. 5). The only inflow observed in our summer field work occurs through minor spring seepage at the east end, and the only outflow (also minor) is filtration through a boulder-filled channel extending from the south shore and leading to Catalina Lake (Fig. 6). Raven Lake does not receive meltwater from Renland Ice Cap at present. However, an abandoned channel, which heads ~200 m from the ice-cap margin and distal to the fresh drift limit, enters the northeast corner of the lake and is graded to present lake level. A second relict inlet, also graded to present lake level, occurs on the north side of Raven Lake and extends to a highland area about 1.1 km distal to the present ice-cap margin. Deltas occur along this channel at about 4 m above present-day lake level and are associated with a distinct raised shoreline at that elevation. A prominent abandoned outlet exits the southwest corner of the lake and heads at the same elevated shoreline. In order for the lake to have been 4 m higher than at present, sediment or ice that is no longer in existence must have dammed two lower-elevation channels – the present-day outlet and the present-day inlet (which would become an outlet if water level were ~>1 m higher than at present). We speculate that the elevated shoreline, deltas, and associated relict outlet channel may have been active during initial stages of deglaciation when thick ice or ice-cored moraine still occupied Catalinadal.

Raven Lake is surrounded by weathered, lichen-covered till. A series of degraded lateral moraines occurs east of the lake in the valley that now hosts the only active inlet. These moraines slope downward to the southeast and relate to an expanded Renland Ice Cap.

5.2. Core data

We collected five cores from the central basin of Raven Lake, reaching as much as 2.7 m below the sediment-water interface (Supplemental Information; Table S2). Our analyses focused on RAV11-1, comprised of two, meter-long segments (RAV11-1A-1, RAV11-1A-2) from the same hole (Fig. 7). The lower unit (Unit A, 160–210 cm depth) consists of pink and gray massive sand with scattered pebbles found below 203 cm depth (Fig. 7). Organic-rich, finely laminated (millimeter-scale) gyttja (Unit B) abruptly overlies the sand and persists to the surface of the core. The laminae group to produce alternating centimeter-thick layers of dark grayish-brown, olive brown, and black sediment, with the lower layers being generally darker in color than those higher in the core.

MS and percent LOI show an inverse relationship (Fig. 7), with the MS values being relatively high and variable near the base of the core but remaining low above 160 cm depth. In contrast, percent LOI jumps from near zero at the base of the core to nearly 30% at 160 cm depth. Percent LOI peaks in the middle of the upper unit and declines gradually thereafter. Cores from other sites are composed of the same two units (Supplemental Information). However, when plotted by depth below the sediment-water interface, the contact is not at the same apparent level everywhere; nor are overlying sediments of the same age (Fig. 8).

Eleven radiocarbon dates of organic samples range from ~970 to

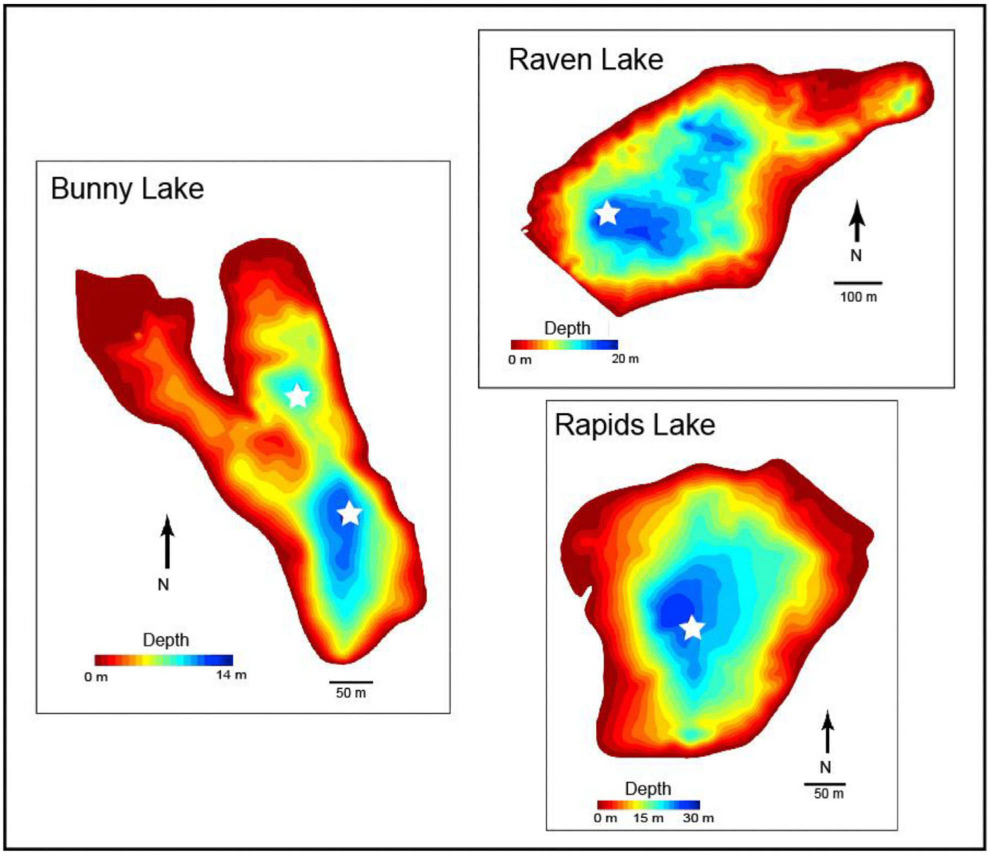

Fig. 5. Bathymetric maps of Raven, Bunny, and Rapids Lakes. Stars mark the location of the master cores.

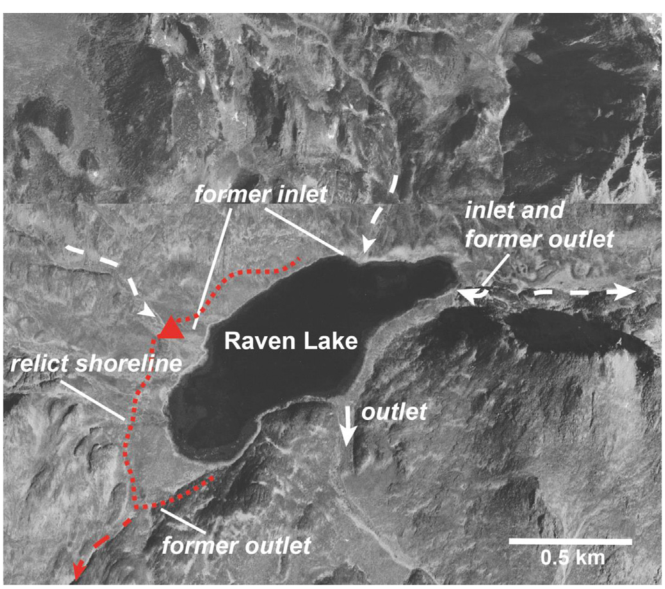

Fig. 6. Geography of Raven Lake, showing current and former outlets and inlets. A higher former shoreline and associated outlet and delta (triangle) are shown in red. Location of image is marked in Fig. 2. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

12,670 yr BP (Table 1). Most remains were too small to identify. All seven dates from the master core were used in the age-depth model (Fig. 9). There were no age reversals. The remaining dates constrain the lower part of the stratigraphy in other cores.

6. Rapids Lake

6.1. Setting and geomorphology

Rapids Lake, a single, steep-sided basin (0.8 km^2, 30 m deep; Fig. 5), is one in a series of lakes connected by a turbulent bedrock-floored river that today drains Catalina Lake (Figs. 2, 4D and 10). The river, which carries glacial rock flour at present, enters the lake through a waterfall to the east and exits the southwest corner, where it continues to Bunny Lake. An abandoned channel extends from the plateau above the lake to the south shore. Although the valley walls show evidence of past rockfall, boulders surrounding the lake have extensive lichen cover, indicating overall stability.

6.2. Core data

We collected three cores from the edge of the central basin in Rapids Lake, as well as an additional core from the break in slope north of the basin. Coring in every instance was halted by rocks. Because all cores showed the same general stratigraphy (Supplemental Information), we chose the longest record, RPD11-1B-1, as the master core, which we hereafter call RPD11-1 (Fig. 11).

The lowest recovered unit (Unit A), at 83–107 cm depth, consists of laminated black clayey-silt with multiple interbedded layers of reddish-gray coarse silt and fine sand (Fig. 11). Dark brown, very finely laminated (0.1 cm) silt (Unit B) overlies the lower unit with a sharp contact and persists until 37 cm depth. These sediments are interrupted by multiple layers of gray clayey silt, the most prominent of which occurs at 50.5 cm depth (0.3 cm thick). Between 37 and 38 cm depth, a transition occurs to overlying gray, laminated,

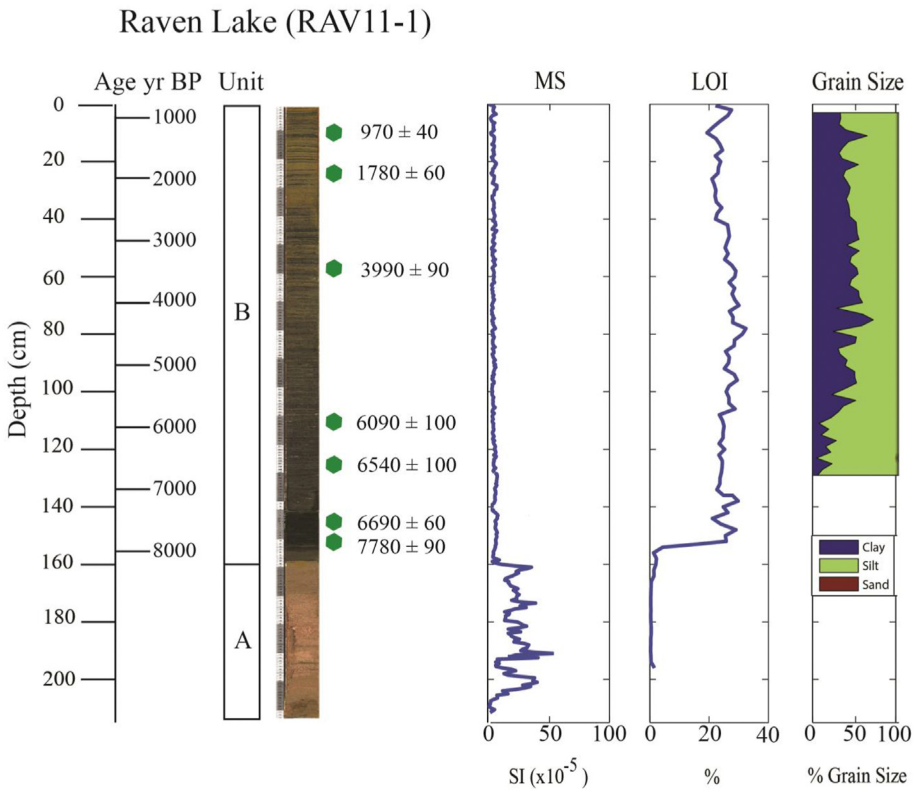

Fig. 7. Data from RAV11-1. Dots adjacent to the core photo mark the location of radiocarbon samples.

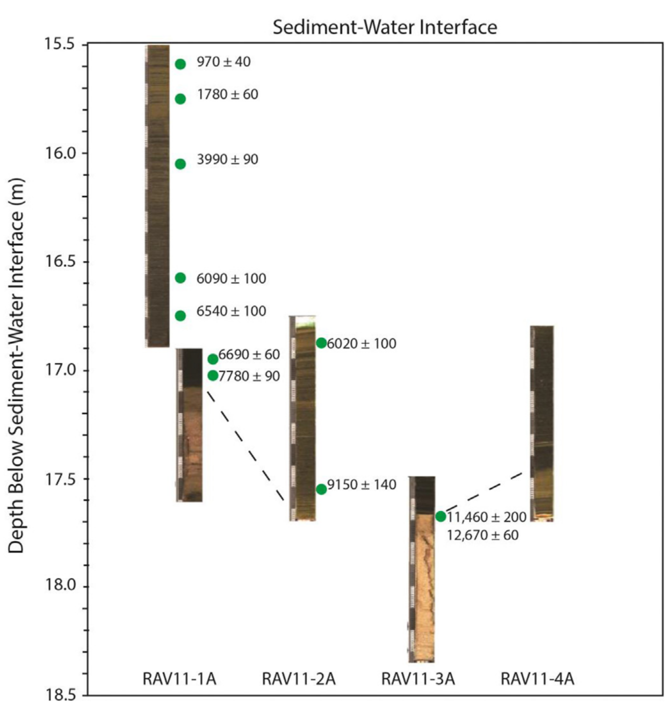

Fig. 8. Raven Lake cores arranged by depth. Dashed lines show correlations among cores at the clay-silt boundary. Green dots mark radiocarbon dates. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

silty-clay. This unit (Unit C) is maintained to the surface and contains four interbedded layers of dark brown, massive, clayey-silt. Two of the layers are oxidized along their upper boundary.

MS and LOI all fluctuate with observed changes in sediment. The gray clayey-silt layers are low in organic material and high in MS. Dark brown silt generally shows the opposite relationship, with the notable exception of the dark brown silty clay layers that interrupt the upper clay unit. These layers have the highest MS observed in the core and very low organic material.

Five radiocarbon dates from RPD11-1 range from ~180 to 8120 yr BP (Table 1). We removed OS-96048 from the age-depth model (Fig. 9), because it is significantly older than two of the underlying dates. Overall, the chronology for RPD11-1 is not robust, which limits the conclusions that can be drawn from this record.

7. Bunny Lake

7.1. Setting and geomorphology

Bunny Lake (0.6 km^2) is ~250 m west of Rapids Lake, along the same drainage (Fig. 10). The two lakes are separated by a large, bouldery deposit with multiple, closely spaced hummocky crests (as much as 10 m of relief), possibly a highly degraded terminal moraine. Surface boulders are covered thickly with lichen. The river flows along the southern shore and discharges through a northwestern outlet. An abandoned channel descends from the south plateau to the lake. Bunny Lake has two major basins separated by a 9-m-deep sill (Fig. 5). The south, ~14 m deep, is affected by the through-flowing river. The north basin (~10 m deep) is more isolated and quiescent.

7.2. Core data

We collected five cores from Bunny Lake, two from the south and three from the north basin. Coring reached a maximum of 1.7 m

Table 1
Radiocarbon dates from lake sediment cores. Master depth refers to the depth on the combined master cores. Percent refers to the percent probability of the calbration. Probabilities <10% are not shown.

Core	Depth (cm)	Master Depth (cm)	Lab ID	Sample Type	^{14}C yr BP	error	Cal yr BP	2s	%	δ13C
Raven Lake										
RAV11-1A-1	9	9	OS-95567	Organic fragments	1080	35	970	40	68	−25
							1040	20	32	
RAV11-1A-1	24	24	OS-96907	Organic fragments	1890	25	1780	60	93	−32.1
RAV11-1A-1	56	56	OS-95590	Macrofossil	3660	30	3990	90	100	−29.5
RAV11-1A-1	108	108	OS-96908	Organic fragments	5300	35	6090	100	97	−30.7
RAV11-1A-1	125.5	125.5	OS-96909	Organic fragments	5740	40	6540	100	99	−31.2
RAV11-1A-2	4	145	OS-95588	Organic fragments	5880	30	6690	60	95	−30.4
RAV11-1A-2	11	152	OS-95589	Organic fragments	6960	35	7780	90	95	−28.6
RAV11-2A-1	9		OS-95566	Organic fragments	5250	45	6020	100	85	−30.4
							6160	20	15	
RAV11-2A-1	76		OS-95569	Organic fragments	8180	50	9150	140	100	−27.7
RAV11-3A-1	17		OS-100776	Organic fragments	10,650	55	12,670	60	91	−29.9
RAV11-3A-1	17		OS-105116	Organic fragments	9990	65	11,460	200	90	−29.7
Rapids Lake										
RPD11-1B-1	5	5	OS-96048	Macrofossil	960	20	830	40	77	−23.1
							910	10	21	
RPD11-1B-1	17.5	17.5	OS-96047	Macrofossil	200	20	180	40	60	−25.2
							280	10	25	
							modern		15	
RPD11-1B-1	21	21	OS-96058	Macrofossil	250	25	300	20	65	−26.5
							160	10	22	
RPD11-1B-1	39	39	OS-96045	Organic fragments	1990	25	1930	60	95	−24.5
RPD11-1B-1	99	99	OS-95933	Macrofossil	7330	55	8120	100	91	−24.1
Bunny Lake South										
BNL11-1A-1	23	13	OS-95595	Organic fragments	425	40	480	50	84	−26.6
							350	20	16	
BNL11-1A-1	37.4	27.4	OS-96874	Macrofossil	1150	30	1050	80	90	−30.8
BNL11-1A-1	61	51	OS-96906	Organic fragments	2910	25	3020	60	75	−31.7
							3120	30	25	
BNL11-1A-1	107	97	OS-95587	Organic fragments	6380	30	7300	40	77	−25.0
							7400	20	17	
BNL11-1B-1	56	97	OS-95586	Organic fragments	6290	40	7240	80	97	−29.7
BNL11-1B-1	77	117	OS-95640	Macrofossil	7170	50	7980	60	88	−24.6
Bunny Lake North										
BNL11-2A-1	13	4.5	OS-100375	Macrofossil	280	20	400	30	52	−24.2
							310	20	48	
BNL11-2A-1	26.8	18.3	OS-100376	Organic fragments	1440	25	1330	30	100	−32.8
BNL11-2A-1	28	19.5	OS-100377	Organic fragments	1460	20	1340	30	100	−31.7
BNL11-2A-1	59.5	51	OS-100378	Organic fragments	3400	40	3640	80	89	−31
BNL11-2A-1	69	60.5	OS-100379	Organic fragments	3280	30	3510	60	98	−32.6
BNL11-2A-1	73	64.5	OS-100380	Organic fragments	3670	30	4000	90	100	−33.5
BNL11-2A-1	74	65.5	OS-100381	Organic fragments	3480	30	3760	70	95	−32.7
BNL11-2A-1	85.5	77	OS-100382	Organic fragments	3430	30	3650	80	82	−32.9
							3810	20	13	
BNL11-2A-1	86	77.5	OS-100516	Organic fragments	4410	30	4960	90	92	−34.4
BNL11-2A-1	95	86.5	OS-100517	Organic fragments	5250	30	5980	50	59	−34.2
							6080	40	29	
							6160	20	12	
BNL11-2A-1	113	104.5	OS-100777	Organic fragments	6520	35	7460	50	64	−28.7
							7360	30	34	
BNL11-2A-2	3	122	OS-100518	Organic fragments	6450	30	7370	60	100	−29
BNL11-2A-2	9.5	128.5	OS-100519	Organic fragments	6710	30	7590	30	66	−28.6
							7530	20	28	
BNL11-2A-2	15	134	OS-100520	Organic fragments	7300	35	8100	80	100	−29.9
BNL11-2A-2	22.5	141.5	OS-100521	Organic fragments	7470	35	8320	40	51	−30.2
							8230	40	49	
BNL11-2A-2	26	145	OS-100778	Organic fragments	7960	40	8840	150	92	−31.8
BNL11-2A-2	29.5	148.5	OS-100522	Organic fragments	7870	35	8680	100	92	−29.8
BNL11-2A-2	33	152	OS-100779	Organic fragments	8500	50	9490	50	100	−30.1

below the sediment-water interface and was halted by gravel. For each basin, we constructed a master core from the longest available records. BNL11-1 (south basin) is formed from a combination of two overlapping core segments (BNL11-1A-1, BNL11-1B-1) from holes a few meters apart, which were correlated using visual stratigraphy and confirmed by radiocarbon dating. BNL11-2 (north basin) is formed from two core segments (BNL11-2A-1, BNL11-2A-2) from the same hole.

The lowest retrieved sediments in BNL11-1 consist of ~30 cm of poorly sorted, sub-rounded sand and gravel, as much as 5 cm diameter (Fig. 12). MS is relatively high and variable. Organic content is negligible. This unit (Unit A) is overlain abruptly at 120 cm depth by nearly a meter of very dark grayish-brown silt (Unit B), and LOI jumps to nearly 30%. Close to the lower contact, the sediment is very finely laminated (millimeter-scale), alternating between the dominant black and less prominent very dark grayish-

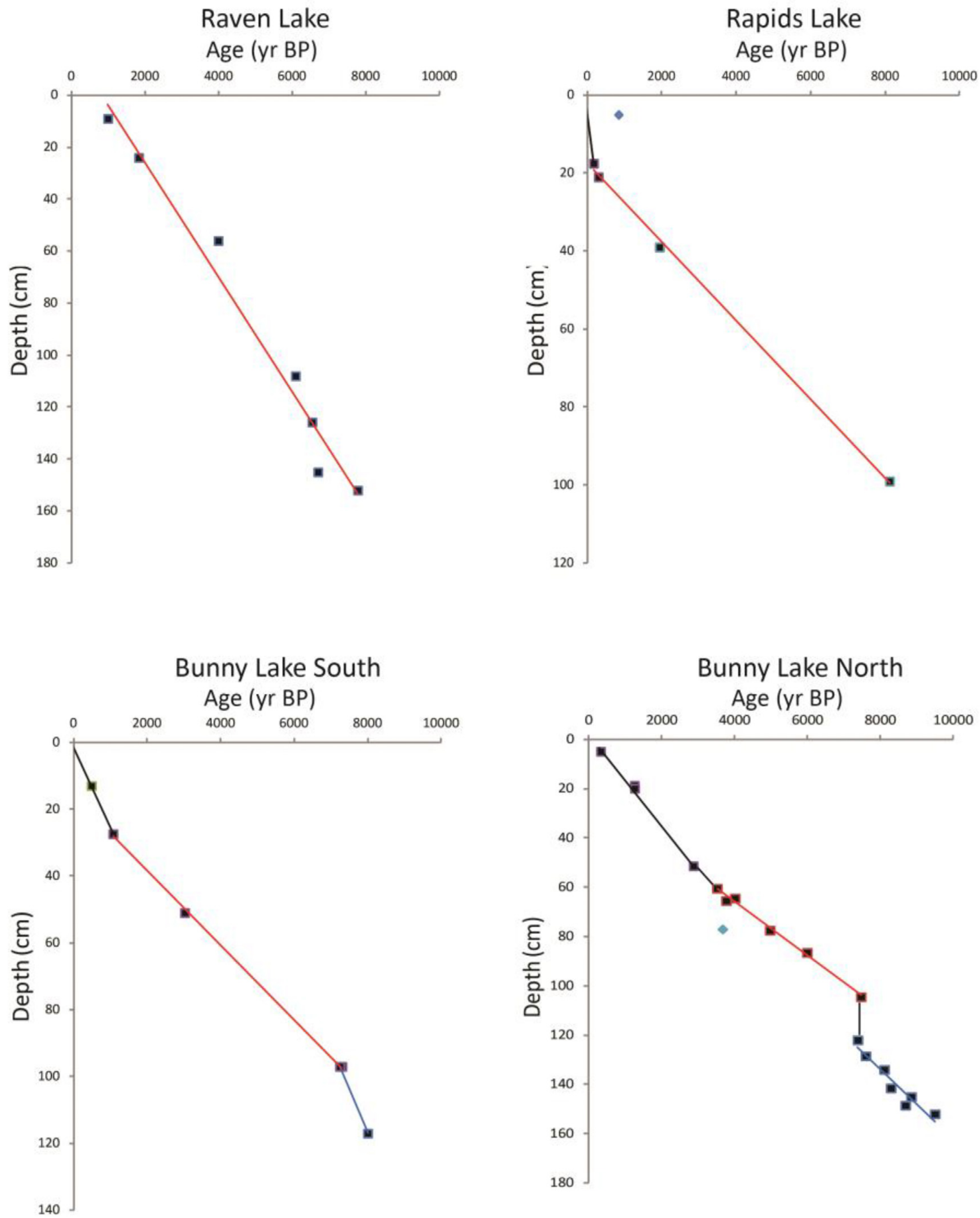

Fig. 9. Age-depth models for the four master cores. We consider the two dates shown as blue diamonds to be outliers and exclude them from the age-depth model calculations. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

brown and gray laminae. Above 99 cm depth, the unit is massive, grading from black to very dark brown. Prominent thin (0.5–2 cm) layers of massive gray clayey-silt occur at 52.5, 86, and 97.5–99 cm depth and are characterized by distinctly higher MS and low organic content.

The very dark brown silt is overlain abruptly at 27 cm depth by gray clayey-silt to silty-clay (Unit C) with high MS and low organic content. This latter unit fines upwards (~25% clay at bottom to 65% at the top) to the surface and is interrupted by a thin reddish-brown silt layer and two oxidized laminae (21 and 3 cm depth), as well as by a thicker, reddish-brown massive silty fine sand layer between 5 and 14 cm depth.

A similar, but higher-resolution record, occurs in the north basin of Bunny Lake ("Bunny Lake North," Fig. 13). Here, the lowest 10 cm consist of dark reddish-gray massive fine sand, which was not retained in the core barrel and thus is not recorded in Fig. 13. The sand is capped by gray, massive clay from 151 to 155 cm depth (Unit A). More than 1.2 m of organic-rich, massive silt overlie the clay with an abrupt contact. This silt (Unit B) grades from black at its base to dark grayish-brown at the top. Similar to BNL11-1, the unit is punctuated by gray, massive, clayey-silt layers (12 in all) that are characterized by distinct MS peaks and low organic content. Two of

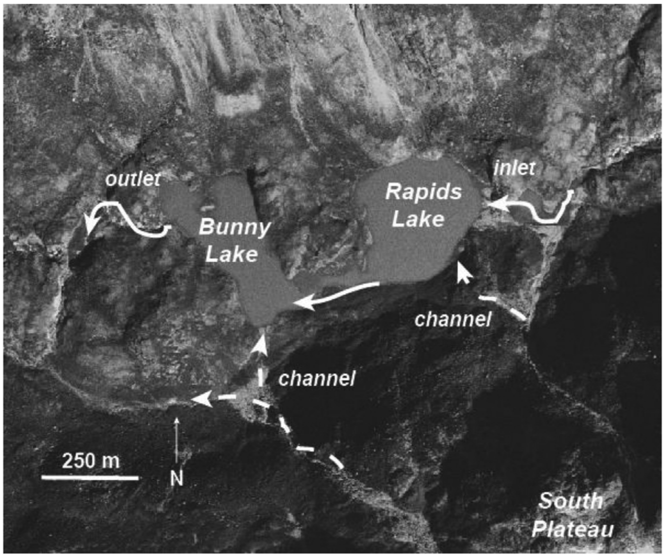

Fig. 10. Geography of the Rapids and Bunny Lake area, showing current and former (dashed) inlets and outlets.

these (51–60.5 and 87.5–88.5 cm depth) also contain dark reddish-gray fine sand, similar to that in the uppermost unit of cores in both basins.

Gray laminated (centimeter-scale) clay (Unit C) overlies the brown silt abruptly at 17 cm depth and persists to the core top. The clay displays high MS and low organic content. Dark reddish-gray massive fine sand interrupts the clay at 3–4 cm depth.

Six radiocarbon dates from the south basin of Bunny Lake range from 480 to 7980 yr BP (Table 1). All are in stratigraphic order and were used to create an age-depth model for BNL11-1 (Fig. 9). Eighteen dates from the north basin range from 400 to 9490 yr BP and show some minor age reversals that are most likely due to uncertainties in radiocarbon dating. One date was excluded from the age-depth model (Fig. 9) because of an age reversal.

8. Recently exposed plant remains

In situ relict plant remains have been exposed by the retreating Renland Ice Cap margin north of Raven Lake (Fig. 2). Fifteen calibrated ages of these materials, primarily moss, but also willow (*Salix*), range from ~400 to 1000 yr BP. The oldest group clusters at ~950 yr BP (Table 2; Fig. 14) and consists of three samples found within 15 m of the 2011 ice margin, all from old landscape surfaces with a thin drape of younger glacial debris. The remaining samples average ~500 yr BP, with most found buried under 0.2–1.0 m of till and exposed in sections near the ice margin. One sample consisted of a rumpled relict surface and two were preserved in cracks in bedrock on the lee side of a hill.

9. Discussion

9.1. Correlation of cores and interpretation of sediments

9.1.1. Raven Lake

The deepest recovered sediments in Raven Lake (our control lake) are coarse sand and gravel, which we interpret as ice-proximal, laid down during initial deglaciation. Following isolation of the lake from glacial meltwater, sedimentation became organic-rich. Two radiocarbon samples from RAV11-3A-1 suggest that this transition occurred shortly before 11,460 ± 200 (OS-105116) to 12,670 ± 60 yr BP (OS-100776) (Fig. 8). These samples were taken from approximately the same depth in the core (although not from the same sample) and their ~1000-yr difference may reflect poor chronologic resolution due to slow sedimentation.

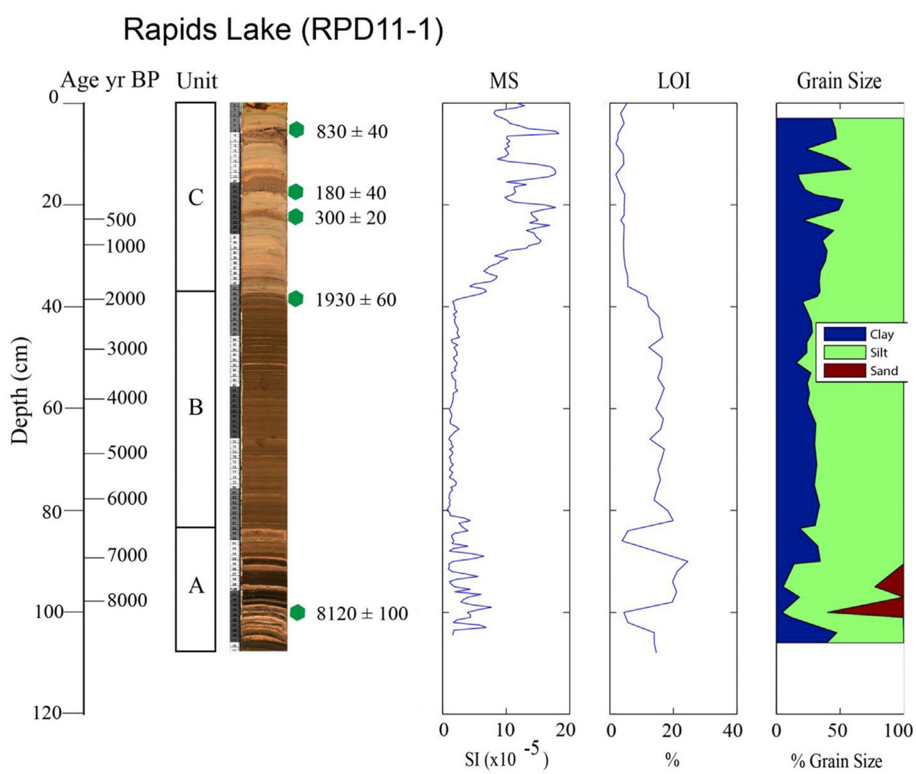

Fig. 11. Data from RPD11-1. Dots adjacent to the core photo mark the location of radiocarbon samples.

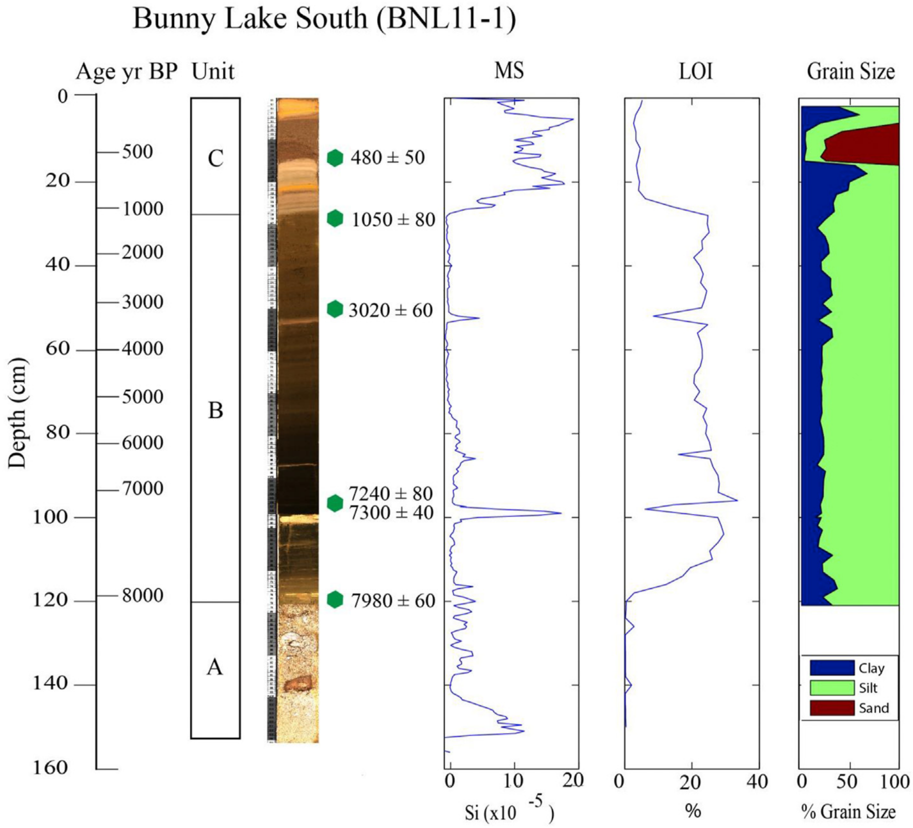

Fig. 12. Data from BNL11-1. Dots adjacent to the core photo mark the location of radiocarbon samples.

In RAV11-2A-1, a sample >15 cm above this contact dates to 9150 ± 140 yr BP (OS-95569), consistent with the dates from RAV11-3A-1. However, in RAV11-1A-2, a date from close to the same contact is only 7780 ± 90 yr BP (OS-95589) in disagreement with dates from other cores (Fig. 8). As mentioned above, the contact between the two units is not horizontal and becomes lower where the older dates occur. We suggest that the contact is unconformable, and material is missing from the shallower cores, possibly due to erosion or non-deposition during a period of low lake level in the early Holocene. Because we cannot accurately correlate RAV11-3A-1, which has the oldest ages, with the master core, we leave dates from this core out of the age model. However, if the dates are correct, by about 12,700 yr BP, the basin must have been free of glacial ice. At minimum, dates from the master core indicate it must have been ice free before 7780 yr BP.

Most sediments in Raven Lake consist of very finely laminated, organic-rich silt, which shows a slight enrichment in coarser particles in the second half of the Holocene. Notably, these sediments lack inorganic clay layers similar to those found in Bunny and Rapids Lakes.

9.1.2. Bunny and Rapids Lakes

Because Rapids and Bunny Lakes occupy the same drainage, we expect their sediments to be similar. For the most part, this is the case, although the sedimentation rate was much slower in Rapids Lake than in Bunny Lake (particularly in the north basin). In general, resolution in the Rapids core is poor, and age reversals in the upper sediments complicate the age model. When records from the two lakes disagree, we favor that from Bunny Lake.

Both lakes show coarse-grained clastic sediments at the base of the cores, interpreted to be proximal glaciolacustrine deposits. In Rapids Lake, the lowest radiocarbon date affords a minimum age of 8120 ± 100 yr BP for ice retreat from the site. The basal age for the south basin of Bunny Lake is similar, but the lowest date from the north basin is 9490 ± 50 years old. Given that the north basin has a more complete record, we suggest that ~9500 yr BP represents the closest minimum-limiting age for deglaciation of both Bunny and Rapids Lakes. As the south basin and Rapids Lake would have been along the path of the outwash stream from the retreating glacier, these earliest sediments may have been scoured (or not deposited) in these basins.

Rock flour is common in the early and recent sediments in both lakes. Immediately above the basal clastic sediments, thin layers of blue-gray inorganic clay are interbedded with organic-rich sediments. This abrupt and recurrent change in sediment type may reflect repeated failure and reformation of the ice dam at the east end of Catalina Lake to allow high frequency switches between glacial and organic sedimentation. Thus, by ~9500 yr BP ice extent was likely already slightly smaller than at present, and it fluctuated around this level for at least 1500 years.

The upper part of the cores is composed of a thicker unit of the same gray, inorganic clayey silt, along with layers of coarser clastic sediments. The onset of this unit is best constrained in Bunny Lake. Here, a precursor event is recorded in the north basin, where a thin gray layer dated firmly at 1340 ± 30 yr BP (AD 610) marks the first visible rock flour since ~3000 yr BP. There is a brief return to organic sedimentation, followed by the onset of the thick clayey silt unit which persists to the top of the core. The onset of the thick clayey silt unit is not dated in the north basin, but in the south basin, it occurred abruptly shortly after 1050 ± 80 yr BP (~AD 900). Based on the thickness of this upper clayey silt, we infer that it represents the longest-lasting and probably the most extensive glacial event of the

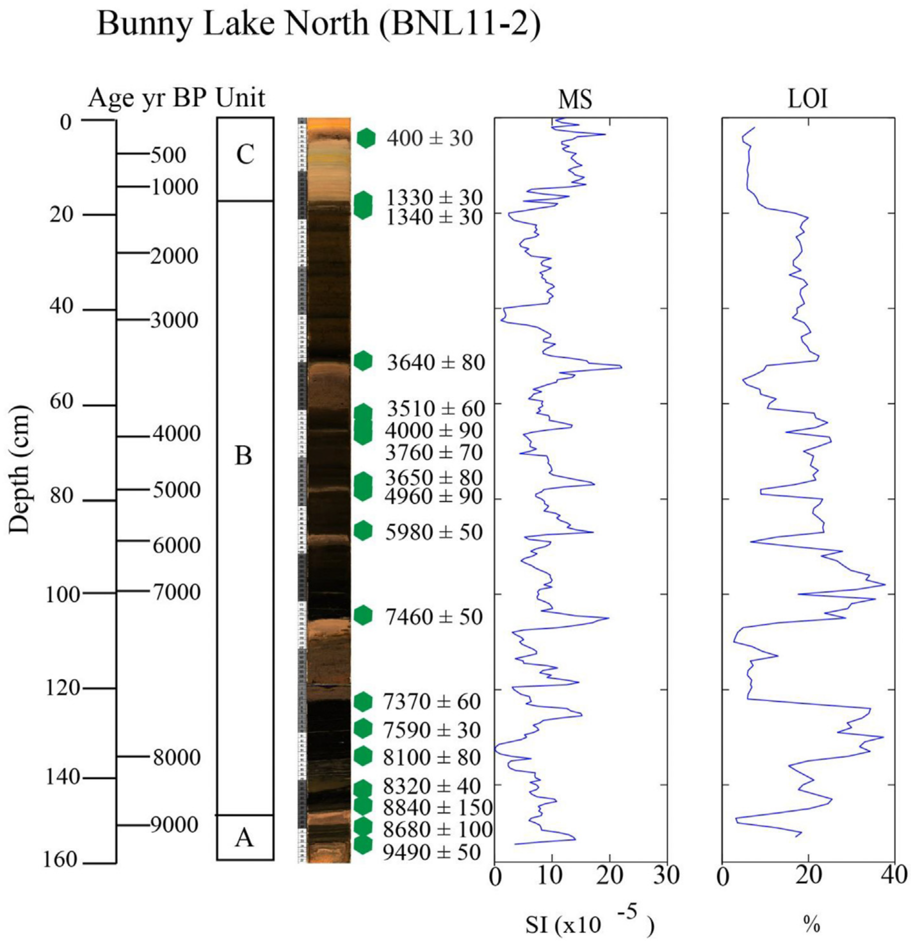

Fig. 13. Data from BNL11-2. Dots adjacent to the core photo mark the location of radiocarbon samples.

Table 2
Radiocarbon ages of organic remains recently exposed by the retreating Renland ice cap. Superscripts in first column are keyed to Fig. 14. Calibration probabilities less than 10% are not shown.

Lab #	Lat.	Long.	Elev. (m)	^{14}C age	Error	Cal. yr	2s	$\delta^{13}C$	Context
[1]OS-103375	71.10710	−27.29640	1645	285	20	400 (60%)	30	−22.3	Moss in bedrock crack on higher summit
						310 (40%)	20		
[2]OS-103378	71.08548	−27.33017	1173	305	30	400 (74%)	60	−23.1	Moss in ice at ice/drift contact
						320 (26%)	20		
[3]OS-94910	71.10485	−27.31255	1419	330	25	390	80	−25.7	Moss/soil under 0.5 m till exposed in cut bank, 3 m from ice
[4]OS-95158	71.10485	−27.31255	1419	370	25	460 (58%)	40	−25.7	Moss/soil under 0.5 m till exposed in cut bank, 3 m from ice
						350 (40%)	30		
[5]OS-103373	71.10708	−27.29762	1642	430	25	490	30	−22.3	Moss in bedrock crack on lower summit
[6]OS-103371	71.10670	−27.30828	1483	435	30	490	40	−26.7	Moss in crack in rocks, about 1 m from ice margin
[7]OS-103377	71.08508	−27.33142	1174	455	25	510	20	−26.0	*Salix* rooted in relictfluvial terrace 2 m from margin
[8]OS-103437	71.10557	−27.31058	1438	470	25	510	20	−24.7	Moss/soil under rock 3 m from ice margin
[9]OS-95042	71.10503	−27.31155	1420	515	25	530	20	−23.6	Moss under 0.4 m of till exposed at base of cut bank 4 m from ice front
[10]OS-103436	71.10557	−27.31058	1438	550	25	540 (65%)	20	−25.4	Moss on rumpled up relict surface in thrust block at ice margin
						610 (35%)	20		
[11]OS-95163	71.10105	−27.31448	1267	555	30	540 (55%)	20	−22.8	Moss/soil layer in section under 20 cm of till, same mat as 10
						610 (45%)	20		
[12]OS-95043	71.10353	−27.31255	1398	675	25	650 (59%)	20	−25.5	Moss/soil layer under 1 m of till exposed in cut through stream terrace, 6 m from ice
						580 (41%)	10		
[13]OS-103376	71.08263	−27.34203	1145	980	20	830 (63%)	40	−26.8	*Salix* rooted in relic surface inside historical drift limit, 15 m from ice margin
						910 (37%)	10		
[14]OS-103372	71.10672	−27.30255	1554	1030	25	940	30	−21.4	Moss right at ice margin
[15]OS-103374	71.10718	−27.29738	1635	1130	30	1020	60	−24.3	Moss right at ice margin

entire record (assuming negligble sediment compaction). Renland glaciers dammed Catalina Lake, and drainage was to the west to Bunny and Rapids Lakes, a situation that has persisted until present day. We also infer that this is the most likely time for ice-cap growth on the south plateau.

The upper clay unit is interrupted in each basin by two or three

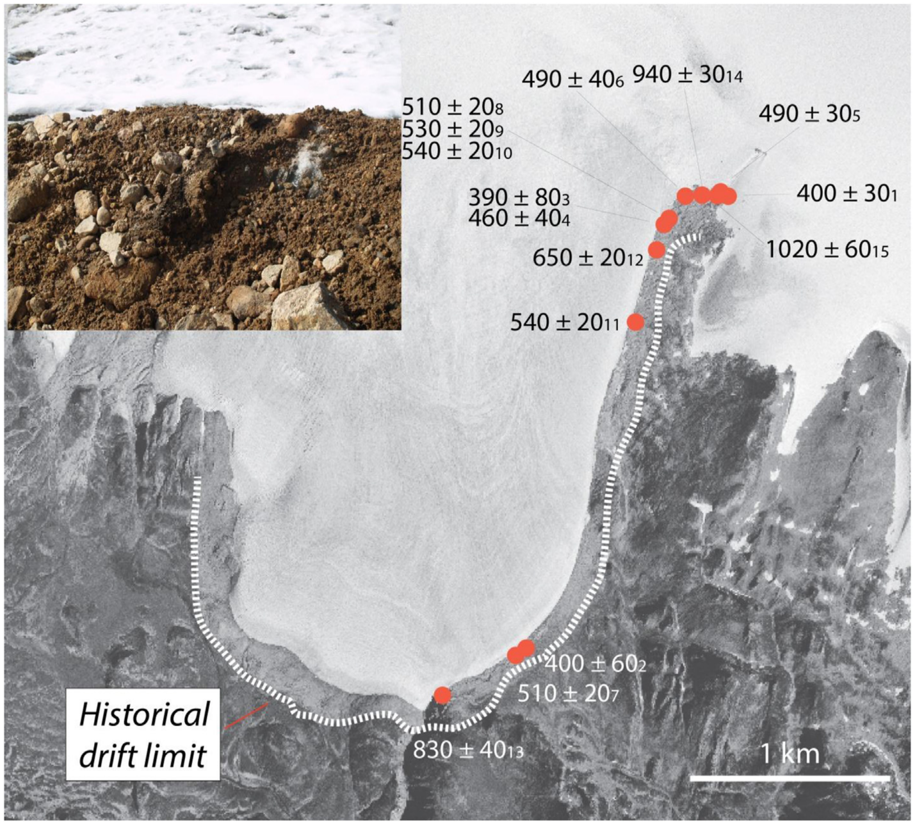

Fig. 14. Location of dated *in situ* plant remains uncovered by recent ice recession. Ages are in calendar years. For samples with multiple calibration peaks, only the most probable peak is shown; all other data are in Table 2. Samples are keyed to Table 2 with the red numbers after each age. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

well-sorted coarser-grained silt and fine sand layers. The origin of these layers, some of which have an orange color, remains uncertain. One option is that they represent mass-wasting events, either from the valley walls or as slush-flow events. Another is that they reflect sediment reworking during a period of low water level associated with a reduction in meltwater. This meltwater reduction could be due either to unusually cold summer conditions or to brief recession of the ice dam which cut meltwater flow to the lakes. If meltwater inflow were curtailed, lake level would drop to the level of the outlet sill. Any continued water-level decrease would be dependent on evaporation (or lake-ice sublimation) exceeding precipitation. Evidence in support of periods of low water level comes from diatom assemblages and water-depth modelling from the south basin of Bunny Lake, which suggest at least two periods of relatively low water level (>6 m below present) dominated by benthic taxa during the last 1000 years (Slemmons et al., 2016). The most recent of these occurred at ~480 yr BP, with a decrease in planktic-benthic ratios, and is associated with the coarse silt/fine sand deposition (Slemmons et al., 2016).

Organic-rich silt sandwiched between the upper and lower clay units comprises most of the cores and indicates little to no input of glacial material for most of the Holocene. Thus, Renland glaciers must have ceased to have dammed Catalina Lake and drainage was to the east. There also is a small decrease in organic content over time, similar to that at Raven Lake, which may be due either to a long-term decline in productivity and/or gradual increase in minerogenic sediment. If minerogenic, it is unlikely to be from rock flour, because the decrease in organic content is not accompanied by an increase in MS (rock flour in these lakes has a high MS signal).

A significant feature of the organic silt unit is that it is interrupted by multiple layers of gray clayey-silt, identical in color, grain size, MS, and LOI to the rock flour being deposited today in these lakes. We favor a glacial origin for these sediments due to this similarity, as well as because of the lack of similar inorganic laminae, such as might be produced by periods of heavy surface runoff, cryoturbation or dust storms, in non-glacial Raven Lake. We cannot exclude the possibility of mass movements producing layers similar to these in the lakes, but given that several occur in more than one basin, we do not favor that scenario (Fig. S1). The two most prominent layers occur at ~3200–3500 and 7200–7600 yr BP. If these layers do represent glacial influx, they would have formed at a time when ice was generally smaller than at present in the mid-Holocene, but expanded just enough during the coldest times to occasionally dam Catalina Lake.

9.2. Interpretation of recently exposed plant remains

Radiocarbon ages of plant remains exposed by recently retreating ice afford information on the timing of marginal fluctuations of the Renland Ice Cap north of Raven Lake. Collectively, the dates indicate times when plants were able to colonize ice-free land that was subsequently covered by ice until recently re-exposed. Moreover, if the plants died because of glacial overriding (as opposed to a glacier overriding long-dead wood), then their ages also afford direct information on the timing of ice expansion.

The oldest ages are ~900–1000 yr BP (n = 3, Fig. 14). Thus, when these plants lived, Renland Ice Cap must have been smaller than it was in AD 2011, allowing plants to grow in spaces that until recently have been covered by ice. While we cannot be certain glacier advance killed these plants, ice expansion at ~900–1000 yr BP would be consistent with the return of rock flour to Bunny Lake shortly after ~1050 yr BP. Other samples (n = 12) date to ~500 ± 100 yr BP, indicating that plants were again at least briefly able to grow in an area close to what would later be the AD 2011 margin. Uncertainties in age calibrations also allow the less likely probability that plants may have grown at other times between ~300 and 700 yr BP. Altogether, the plant samples suggest expansion of ice beyond the current position at some point at or after ~900–1000 yr BP, with at least one period since that time (~500 yr BP) when ice was as small as it is at present, followed by re-expansion. One possibility is that this latter period of restricted ice may correspond with the coarse sediment layer deposited in Bunny Lake at about ~500 yr BP, thought to reflect a period of low water level and reduced meltwater input.

9.3. Summary of Renland lake record and comparison with other Scoresby Sund glacier and lake records

Sediments from the glacially fed lakes afford continuous records of ice-cap behavior in Renland during the Holocene. Such small, terrestrial ice caps are sensitive to changes in snowline elevation, which in turn is influenced primarily by summertime temperature (Oerlemans, 2001). Therefore, we infer that our record largely reflects a history of relative summertime temperature in interior East Greenland over the Holocene.

The Raven Lake record may extend back to the Milne Land Stade, a regional glacial advance in the late-glacial period with both "outer" and "inner" positions. Deglaciation from the outer Milne Land Stade moraines in the Scoresby Sund region occurred prior to ~12,400 yr BP farther east in the Stauning Alper and in Milne Land (Hall et al., 2008b; Kelly et al., 2008; Levy et al., 2016). The degraded lateral moraines down valley of Raven Lake may correlate to this event, although landforms from the Milne Land Stade have not yet been confirmed in Renland. If the oldest age from the base of Raven Lake is correct, then deglaciation of southwest Renland was underway by ~12,670 yr BP, which would be consistent with deglaciation following an outer Milne Land Stade position..

In contrast, the glacial lakes in this study, Bunny and Rapids, have basal ages that extend back only to ~9500 yr BP. Either ice or perhaps large snow drifts persisted in these basins for thousands of years after deglaciation of Raven Lake or the dates obtained are not close-limiting ages for ice retreat, possibly because the basins were dry. We tentatively favor the latter for two reasons. First, based on the differences in basal ages in the cores, we infer that there is an erosional unconformity or hiatus in deposition in Raven Lake, suggesting a possible period of low water in the late-glacial/early Holocene. Second, by 9490 yr BP, organic silt free of visible rock flour already was being deposited in Bunny Lake, suggesting that any glacier margin was distant, so distant that Catalina Lake did not even send meltwater to Bunny Lake. Evidence for restricted ice by about 9500 yr BP also is in accord with data from Istorvet Ice Cap in Liverpool Land, Bregne Ice Cap in Milne Land, and mountain glaciers in the Stauning Alper, all of which indicate ice had retreated to at least within late Holocene positions by ~10,000 yr BP (Hall et al., 2008b; Lowell et al., 2013; Levy et al., 2014; Lusas et al., 2017). It also is consistent with abundant evidence suggesting expansion of thermophilic marine fauna into the Scoresby Sund region in the early Holocene (i.e., Hjort and Funder, 1974; Street, 1977; Hall et al., 2008a) and an abrupt early-Holocene rise in chironomid-inferred temperatures at Last Chance Lake in Milne Land (Axford et al., 2017).

Although apparently smaller than present for much of the Holocene, from at least 9490 until the onset of the late Holocene glaciation about 1050 yr BP, Renland Ice Cap may have undergone periodic fluctuations, as represented by clay layers (Fig. 15). The most prominent event in terms of sediment thickness and spatial extent was at ~7200–7600 yr BP and may have been coeval with the Rødefjord moraines (~7600 yr BP; Funder, 1978), which were deposited by GIS outlet glaciers in the fjords adjacent to Renland. Whether this timing is coincidental or the result of a common climatic forcing of local ice caps and GIS outlet glaciers remains unknown, as the significance of the Rødefjord moraines is uncertain (Funder, 1978).

Clay layers occur less frequently after ~7000 yr BP until the last millennium, indicating that ice only rarely dammed Catalina Lake. The only other significant clay deposition, present in all three basins, occurred at 3200–3500 yr BP. This may represent onset of Neoglaciation with periodic refilling and draining episodes at Catalina Lake. It coincides with a negative temperature anomaly reconstructed from chironomids in nearby Milne Land (Axford et al., 2017). In contrast, inorganic layers at ~5900 and ~4900 yr BP in Bunny Lake North are thin and hard to find in all three basins, suggesting either only brief periods of meltwater input or a non-glacial origin. After a period of organic-rich sedimentation following the ~3200-yr event, a very sharp layer of glacial clay dated in Bunny Lake North to 1340 yr BP (AD 610) represents return to glacial conditions. This is followed by a short interval of organic deposition and then by onset of significant late Holocene glaciation shortly after ~1050 yr BP (AD 900). The deposition of this upper silty clay unit corresponds to the expansion of local ice caps to the limits marked by the gray, unweathered drift (Figs. 4A and 3B).

At face value, onset of late Holocene glacial sedimentation (i.e., the “Little Ice Age”) appears to have occurred several centuries earlier in Renland than at Bone Lake adjacent to the Istorvet Ice Cap in coastal Liverpool Land (Fig. 1; Lowell et al., 2013; Lusas et al., 2017) and nearly a millennium after sustained rock flour input from Bregne Ice Cap in Milne Land (Levy et al., 2014). However, we note that the relationship between the ice caps and the studied lakes differs among these sites. For instance, Bone Lake only receives meltwater when Istorvet Ice Cap is within a few hundred meters of its maximum late Holocene position and thus is suitable for recording only the timing of maximum ice extent. Given that rock flour can be deposited in Bunny Lake under climate and glacier configuration similar to today, our record probably shows glacial events to be of longer duration, with earlier onset and later termination, relative to sites which may record only peak glaciation (e.g., Bone Lake). Thus, our record is likely poised to capture the duration (build up and demise) of glacier fluctuations, but less accurate at identifying the time of maximum ice extent. As only the very warmest times will result in sustained retreat of the ice dam and drainage of Catalina Lake to the east, our data are perhaps more usefully interpreted as documenting peak warmth during times of organic deposition. The high elevation of the Renland Ice Cap makes it ideal to record glacial activity during former warm periods because, unlike many other indpendent ice caps in this region, it was able to maintain an accumulation zone during the Holocene thermal maximum. These observations lead us to stress the importance of collecting records from numerous lakes in different geographic and geomorphologic settings to develop a consensus on the timing of glacial and climate events.

9.4. Comparison with other east Greenland climate proxy data

The original reason for selecting our field area was to compare ice-marginal fluctuations with the adjacent Renland ice core climate data (Vinther et al., 2008). However, while both datasets are consistent with a warm early Holocene, we see little other direct correspondence between our record of ice-marginal fluctuations and the stable isotopic data from the top of the ice cap (Fig. 15). Oxygen-isotope data from the Renland ice core trend nearly

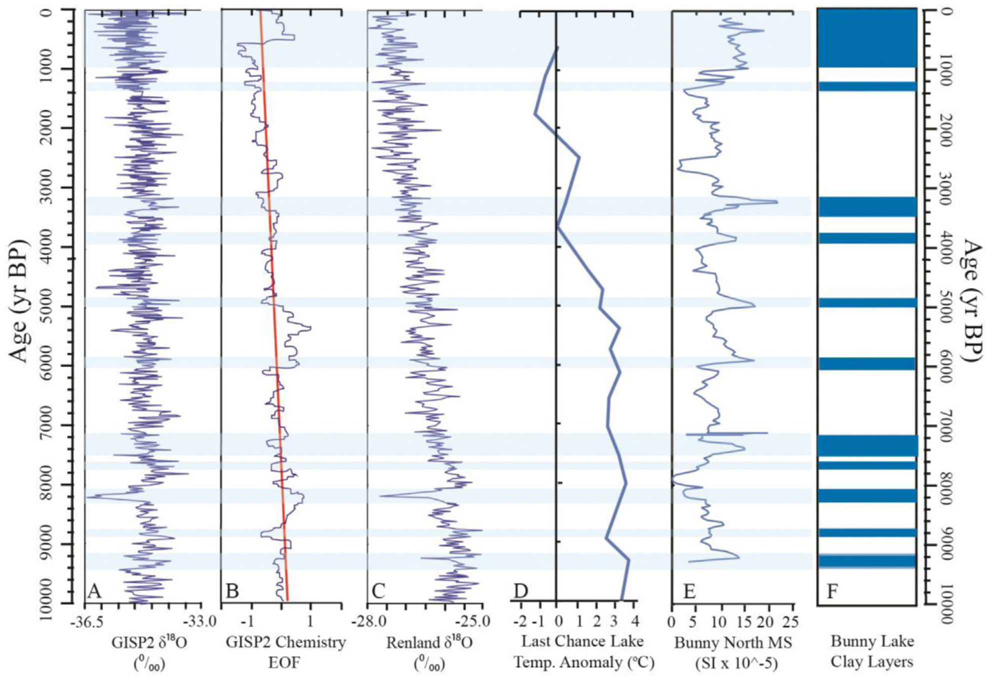

Fig. 15. Comparison of the Renland lake record with regional climate proxies. A. GISP2 $\delta^{18}O$ profile (Grootes and Stuiver, 1997); B. GISP2 chemistry data as represented by an empirical orthogonal function (O'Brien et al., 1995); C. Renland ice core $\delta^{18}O$ profile (Vinther et al., 2008); D. Temperature anomaly relative to preindustrial values inferred from chironomids in Last Chance Lake, Milne Land (Axford et al., 2017); E. Magnetic susceptibility record - a proxy for glacial sediment input - from the north basin of Bunny Lake; F. Chronology of the visible inorganic clayey-silt layers in Bunny Lake indicated by blue lines. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

monotonically toward more negative values, starting at ~9500 yr BP. Small, centennial-scale fluctuations are superimposed on the overall record, but there is little expression of millennial-scale changes, other than the “8200 yr event”. Strong evidence of other millennial fluctuations, such as the Little Ice Age, is missing. One possibility is that the ice-cap mass balance as recorded by marginal fluctuations and meltwater is reacting largely to summer temperature variations, whereas the $\delta^{18}O$ in Greenland ice cores is thought to be strongly affected by winter conditions (Denton et al., 2005; Buizert et al., 2014).

Dust proxy records from the new Renland ice core (RECAP core) are in broad agreement with our data. Although not a direct measure of ice extent in Renland, Simonsen et al. (2019) interpreted coarse particles in the ice core as indications of ice-sheet retreat in central East Greenland at 12,100–9000 yr BP, consistent with our evidence for rapid retreat of Renland Ice Cap during that same time period.

Insect assemblages at Last Chance Lake in Milne Land indicate warmer summer temperatures than today throughout the early and most of the middle Holocene (Axford et al., 2017), in good agreement with our records. The temporary migration of dense dwarf shrub heath northward into the Scoresby Sund region ~8800-5500 yr BP also suggests warmer temperatures than at present (Funder, 1978). Moreover, the period ~7800-4100 yr BP also corresponds to a time of minimum Holocene ice extent in southeast Greenland (Balascio et al., 2015).

We see weak correspondence of our Renland glacial record with the GISP2 atmospheric chemistry data, taken to reflect either strength of the polar vortex or fluctuations in meridional atmospheric circulation. Periods of strong polar circulation, indicated by increases in an empirical orthogonal function representing variance in the chemical assemblage, are thought to characterize cool phases of millennial-scale oscillations (O'Brien et al., 1995). Some periods of strong polar circulation are coeval with times of glacial sediment input into the Renland lakes (i.e., at ~8200 yr BP), but in general the two records show little resemblance to each other. During the best-dated glacial expansion - that which occurred during the last millennium - ice advanced not only in Renland but elsewhere in east (e.g., Lowell et al., 2013; Lusas et al., 2017), northeast (Adamson et al., 2019), and southeast (Van der Bilt et al., 2018, 2019) Greenland several hundred years earlier than the corresponding change in ice-core chemistry, thought to have occurred at AD 1400 (Mayewski et al., 2004). Ice-core chemistry changes also appear to have begun after sea-ice expansion in the North Atlantic by ~ AD 1300 (Oglivie, 1992; Oglivie and Jónsson, 2001; Massé et al., 2008). This relative timing among proxies may suggest that atmospheric circulation changes lagged onset of cooling, perhaps only initiating when a temperature threshold was crossed.

Our data afford support for suggestions from earlier work in the Scoresby Sund region (Lowell et al., 2013; Lusas et al., 2017), as well as from parts of northeast and southeast Greenland (Balascio et al., 2015; Adamson et al., 2019; Van der Bilt et al., 2018, 2019), of a precursor event or events just prior to the onset of the "classical" Little Ice Age. For example, a "Dark Ages" advance of the Renland Ice Cap at 1340 yr BP is coeval with expansion of Kulusuk glaciers in southeast Greenland at ~1300 yr BP. This is likely the same event noted in Madsen Lake on the Wollaston foreland in northeast Greenland at 1350-1190 yr BP (Adamson et al., 2019). Glacier advance during the main phase of the Little Ice Age in the Scoresby Sund region began at ~1050 yr BP at Renland (this study) and neared its maximum in coastal Liverpool Land at Istorvet ice cap at ~800 yr BP (Lusas et al., 2017). This advance corresponds in time to a similar expansion in Madsen Lake at 940-825 yr BP (Adamson et al., 2019), followed there by minor recession and then readvance at ~700 yr BP. Although the Renland lake records do not show evidence of such a recession and readvance, the oldest group of plant remains adjacent to the Renland ice cap date to ~900 yr BP and may record this minor recession. A subsequent period of ice-marginal retreat at Renland recorded by plant remains dating to ~500 yr BP does not seem to appear in either the Madsen Lake or Kulusuk records. Altogether, however, there appears to be strong evidence in many East Greenland records for glacial advance at ~1300 yr BP, followed by at least one additional advance prior to the "classic" Little Ice Age onset at ~700 yr BP. These data contribute to a growing body of evidence in Greenland (i.e., Lowell et al., 2013; Young et al., 2015) and elsewhere (i.e., Holzhauser et al., 2005) of centennial-scale, pre-Little Ice Age glacial advances within the past ~1300 years.

10. Conclusions

Analysis of two glacially fed lakes and one non-glacial lake affords information about past ice extent in southwest Renland. Deglaciation commenced as early as ~12,670 yr BP and by ~9500 yr BP, Renland Ice Cap had retreated behind its present-day extent. The presence of organic silt within the glacially fed lakes indicates that the ice cap was smaller than present during most of the Holocene, consistent with the previously documented history of independent ice caps in the region. However, periodic inputs of inorganic sediments to Bunny and Rapids Lakes may suggest repeated fluctuations of Renland Ice Cap on millennial timescales, particularly at ~7200–7600 and 3200–3400 yr BP. Expansion beginning shortly after 1050 yr BP was the most prominent of the Holocene and may have been accompanied by growth of local ice caps on the plateaus above Bunny Lake. The general pattern of ice expansion and contraction in Renland is similar to that at other ice caps in the region, but also has important differences. These differences include the preservation of a possible mid-Holocene record of ice fluctuations at times when lower-elevation ice caps in the Scoresby Sund region were either absent or too small to discharge meltwater into the studied lakes. This finding reinforces the concept that examination of multiple geographic and geomorphologic settings is necessary for a full understanding of ice variations in a region.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Acknowledgements

This work was funded by the Arctic Natural Sciences Program of the United States National Science Foundation (Kelly: ARC-0909270; Lowell: ARC-0909285; Hall: ARC-0908081). We thank CM2Hill, the Danish Polar Center, and Air Greenland for logistical support, and the staff at the Limnological Research Center at the University of Minnesota for assistance with core transport and initial processing. We especially wish to thank two anonymous reviewers for thoughtful, detailed comments.

Appendix A. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.quascirev.2021.106883.

Author statement

Aaron K. Medford, carried out field work and collected samples, performed the laboratory work. Brenda L. Hall, conceived and obtained funding for the project, carried out field work and collected samples, performed the laboratory work. Thomas V. Lowell, conceived and obtained funding for the project, carried out field work and collected samples. Meredith A. Kelly, conceived and obtained funding for the project, carried out field work and collected samples. Laura B. Levy, carried out field work and collected samples. Paul S. Wilcox, carried out field work and collected samples. Yarrow Axford, carried out field work and collected samples, All authors contributed to the interpretations and the preparation of the manuscript.

References

Adamson, K., Lane, T., Carney, M., Bishop, T., Delaney, C., 2019. High-resolution proglacial lake records of pre-Little Ice Age glacier advance, northeast Greenland. Boreas 48, 535–550.
Alley, R.B., 2003. Comment on "When earth's Freezer door is left ajar. EOS 84, 315.
Axford, Y., Levy, L., Kelly, M., Francis, D., Hall, B., Langdon, P., Lowell, T.V., 2017. Timing and magnitude of early to middle Holocene warming in East Greenland inferred from chironomids. Boreas 46, 678–687.
Axford, Y., de Vernal, A., Osterberg, E., 2020. Past warmth and its impacts in Greenland during the Holocene thermal maximum. Annual Reviews of Earth and Planetary Sciences (in press).
Bakke, J., Lie, O., Nesje, A., Dahl, S.O., Paasche, O., 2005. Utilizing physical sediment variability in glacier-fed lakes for continuous glacier reconstructions during the Holocene, Northern Folgefonna, Western Norway. Holocene 15 (2), 161–176.
Bengtsson, L., Enell, M., 1986. Chemical analysis. In: Berglund, B.E. (Ed.), Handbook of Holocene Palaeoecology and Palaeohydrology. John Wiley, Chichester, pp. 423–454.
Bond, G., Kromer, B., Beer, J., Muscheler, R., Evans, M.N., Showers, W., Hoffman, S., Lotti-Bond, R., Hajdas, I., Bonani, G., 2001. Persistent solar influence on North Atlantic climate during the Holocene. Science 294, 2130–2136.
Briner, J.P., Steart, H.A.M., Young, N.E., Philips, W., Losee, S., 2010. Using proglacial-threshold lakes to constrain fluctuations of the Jackobshavn Isbrea ice margin, Western Greenland, during the Holocene. Quat. Sci. Rev. 29, 3861–3874.
Buizert, C., Gkinis, V., Severinghaus, J.P., He, F., Lecavalier, B.S., Kindler, P., Leuenberger, M., Carlson, A.E., Vinther, B., Masson-Delmotte, V., White, J.W.C., Liu, Z., Otto-Bliesner, B., Brook, E.J., 2014. Greenland temperature response to climate forcing during the last deglaciation. Science 345, 1177–1180.
Dahl, S.O., Bakke, J., Lie, O., Nesje, A., 2003. Reconstruction of former glacier equilibrium-line altitudes based on proglacial sites: an evaluation of approaches and selection of sites. Quat. Sci. Rev. 22, 275–287.
Denton, G.H., Alley, R.B., Comer, G.C., Broecker, W.S., 2005. The role of seasonality in abrupt climate change. Quat. Sci. Rev. 24, 1159–1182.
Denton, G.H., Broecker, W.S., 2008. Wobbly ocean conveyor circulation during the Holocene. Quat. Sci. Rev. 27, 1939–1950.
Denton, G.H., Karlén, W., 1973. Holocene climatic variations-Their pattern and possible cause. Quat. Res. 3, 155–205.
Funder, S., 1978. Holocene stratigraphy and vegetation history in the Scoresby Sund area, east Greenland. Grønlands Undersøgelse Bulletin 129, 1–66.
Grinsted, A., Hvidberg, C.S., Campos, N., Dahl-Jensen, D., 2017. Periodic outburst floods from a lake in East Greenland. Sci. Rep. 7 https://doi.org/10.1038/s41598-017-07960-9.
Grootes, P.M., Stuiver, M., 1997. Oxygen 18/16 variability in Greenland snow and ice with 10^{-3}- to 10^{5}-year time resolution. J. Geophys. Res. 102 (C12), 26445–26470.
Hall, B.L., Baroni, C., Denton, G.H., 2008a. The most extensive Holocene advance in the stauning alper, east Greenland, cccurred in the little ice age. Polar Res. 27, 128–134.
Hall, B.L., Baroni, C., Denton, G.H., Kelly, M.A., Lowell, T.L., 2008b. Relative sea-level change, Kjove land, Scoresby Sund, east Greenland: implications for seasonality in younger dryas time. Quat. Sci. Rev. 27, 2283–2291.
Hjort, C., Funder, S., 1974. The subfossil occurrence of *Mytilus edulis L.* in central East Greenland. Boreas 3, 23–33.
Holzhauser, H., Magny, M., Zumbühl, H.J., 2005. Glacier and lake-level variations in west-central Europe over the last 3500 years. Holocene 15, 789–801.
Johnsen, S.J., Clausen, H.B., Dansgaard, W., Gundestrup, N.S., Hansson, M., Jonsson, P., Steffensen, J.P., Sveinbjornsdottir, A.E., 1992. A "deep" ice core from East Greenland. Meddelelser om Grønland Geoscience 29, 3–22.
Jorgenson, M.T., Shur, Y.L., Pullman, E.R., 2006. Abrupt increase in permafrost degradation in Arctic Alaska. Geophys. Res. Lett. 33 https://doi.org/10.1029/2005GL024960.
Kaufman, D.S., Ager, T.A., Anderson, N.J., Anderson, P.M., Andrews, J.T., Bartlein, P.J., Brubaker, L.B., Coats, L.L., Cwynar, L.C., Duvall, M.L., Dyke, A.S., Edwards, M.E., Eisner, W.R., Gajewski, K., Geirsdóttir, A., Hu, F.S., Jennings, A.E., Kaplan, M.R., Kerwin, M.W., Lozhkin, A.V., MacDonald, G.M., Miller, G.H., Mock, C.J., Oswald, W.W., Otto-Bliesner, B.L., Porinchu, D.F., Rühland, Smol, J.P., Steig, E.J., Wolfe, B.B., 2004. Holocene thermal maximum in the western arctic (0-180°W). Quat. Sci. Rev. 23, 529–560.
Kelly, M.A., Lowell, T.V., Hall, B.L., Schaefer, J.M., Finkel, R.C., Goehring, B.M., Alley, R.B., Denton, G.H., 2008. A 10Be chronology of late glacial and Holocene mountain glaciation in the Scoresby Sund region, East Greenland: implications

for seasonality during late glacial time. Quat. Sci. Rev. 27, 2273–2282.
Kwok, R., Cunningham, G.F., Wensnahan, M., Rigor, I., Zwally, H.J., Yi, D., 2009. Thinning and volume loss of the Arctic Ocean sea ice cover: 2003-2008. J. Geophys. Res. 114, C07005. https://doi.org/10.1029/2009JC005312.
Levy, L.B., Kelly, M.A., Lowell, T.V., Hall, B.L., Hempel, L.A., Honsaker, W.M., Lusas, A.R., Howley, J.A., Axford, Y.L., 2014. Holocene fluctuations of Bregne ice cap, Scoresby Sund, east Greenland: a proxy for climate along the Greenland ice sheet margin. Quat. Sci. Rev. 92, 357–368.
Levy, L.B., Kelly, M.A., Lowell, T.V., Hall, B.L., Howley, J.A., Smith, C.A., 2016. Coeval fluctuations of the Greenland Ice Sheet and a local glacier, central East Greenland, during late glacial and early Holocene time. Geophys. Res. Lett. https://doi.org/10.1002/2015GL067108.
Lowell, T.V., Hall, B.L., Kelly, M.A., Bennike, O., Lusas, A.R., Honsaker, W., Smith, C.A., Levy, L.B., Travis, S., Denton, G.H., 2013. Late Holocene expansion of the Istorvet ice cap, Liverpool land, east Greenland. Quat. Sci. Rev. 63, 128–140.
Lusas, A.R., Hall, B.L., Lowell, T.V., Kelly, M.A., Bennike, O., Levy, L.B., Honsaker, W., 2017. Holocene climate and environmental history of East Greenland, inferred from lake sediments. J. Paleolimnol. 57, 321–341.
Maslanik, J.A., Fowler, C., Stroeve, J., Drobot, S., Zwally, J., Yi, D., Emery, W., 2007. A younger, thinner Arctic ice cover: increased potential for rapid, extensive sea-ice loss. Geophys. Res. Lett. 34 https://doi.org/10.1029/2007GL032043.
Massé, G., Rowland, S.J., Sicre, M.-A., Jacob, J., Jansen, E., Belt, S.T., 2008. Abrupt climate changes for Iceland during the last millennium: evidence from high-resolution sea-ice reconstructions. Earth Planet Sci. Lett. 269, 565–569.
Mayewski, P.A., Rohling, E.E., Stager, J.C., Karlen, W., Maasch, K.A., Meeker, L.D., Meyerson, E.A., Gasse, F., van Kreveld, S., Holmgren, K., Lee-Thorp, J., Rosqvist, G., Rack, F., Staubwasser, M., Schneider, R.R., Steig, E.J., 2004. Holocene climate variability. Quat. Res. 62, 243–255.
O'Brien, S.R., Mayewski, P.A., Meeker, L.D., Meese, D.A., Twickler, M.S., Whitlow, S.I., 1995. Complexity of Holocene as reconstructed from a Greenland ice core. Science 270 (22 Dec), 1962–1964.
Oerlemans, J., 2001. Glaciers and Climatic Change. A.A. Balkema Publishers, Lisse, p. 148p.
Oglivie, A.E.J., 1992. Documentary evidence for changes in the climate of Iceland, AD 1500-1800. In: Bradley, R.S., Jones, P.D. (Eds.), Climate since AD 1500. Routledge, London, pp. 94–117.
Oglivie, A.E.J., Jónsson, T., 2001. Little ice age research: a perspective from Iceland. Climate Change 48, 9–52.
Reimer, P.G., Austin, W.E.N., Bard, E., Bayliss, A., Blackwell, P.G., Bronk Ramsey, C., Butzin, M., Cheng, H., Edwards, R.L., Friedrich, M., Grootes, P.M., Guilderson, T.P., Hajdas, I., Heaton, T.J., Hogg, A.G., Hughen, K.A., Kromer, B., Manning, S.W., Muscheler, R., Palmer, J.G., Pearson, C., van der Plicht, J., Reimer, R.W., Richards, D.A., Scott, E.M., Southon, J.R., Turney, C.S.M., Wacker, L., Adolphi, F., Büntgen, U., Capano, M., Fahrni, S.M., Fogtmann-Schulz, A., Friedrich, R., Köhler, P., Kudsk, S., Miyake, F., Olsen, J., Reinig, F., Sakamoto, M., Sookdeo, A., Talamo, S., 2020. The IntCal20 northern hemisphere radiocarbon age calibration curve (0–55 cal kBP). Radiocarbon 62, 725–767.
Serreze, M.C., Barry, R.G., 2011. Processes and impacts of arctic amplification: a research synthesis. Global Planet. Change 77, 85–96.
Simonsen, M.F., Baccolo, G., Blunier, T., Borunda, A., Delmonte, B., Frei, R., Goldstein, S., Grinsted, A., Kjaer, H.A., Sowers, T., Svensson, A., Vinther, B., Vladimirova, D., Winckler, G., Winstrup, M., Vallelonga, P., 2019. East Greenland ice core dust record reveals timing of Greenland ice sheet advance and retreat. Nat. Commun. https://doi.org/10.1038/s41467-019-12546-2.
Slemmons, K.E.H., Medford, A., Hall, B.L., Stone, J.R., McGowan, S., Lowell, T., Kelly, M., Saros, J.E., 2016. Changes in glacial meltwater alter algal communities in lakes of Scoresby Sund, Renland, East Greenlan throughout the Holocene: abrupt reorganizations began 1000 years before present. Holocene. https://doi.org/10.1177/0959683616678468.
Street, F.A., 1977. Deglaciation and marine paleoclimates, schuchert dal, Scoresby Sund, east Greenland. Arct. Alp. Res. 9, 421–426.
Stuiver, M., Reimer, P.J., Reimer, R.W., 2021. Calib. accessed 2021-2-23. http://calib.org.
Stuiver, M., Reimer, P.J., 1993. Extended 14C database and revised CALIB radiocarbon calibration program. Radiocarbon 35, 215–230.
Van der Bilt, W.G.M., Rea, B., Spagnolo, M., Roerdink, D.L., Jørgensen, S.L., Bakke, J., 2018. Novel sedimentological fingerprints link shifting depositional processes to Holocene climate transitions in East Greenland. *Global and Planetary Change* 164 52–64.
Van der Bilt, W.G.M., Born, A., Haaga, K.A., 2019. Was Common Era glacier expansion in the Arctic Atlantic region triggered by unforced atmospheric cooling? Quat. Sci. Rev. 222, 105860.
Velicogna, I., 2009. Increasing rates of ice mass loss from the Greenland and Antarctic ice sheets revealed by GRACE. Geophys. Res. Lett. 36 https://doi.org/10.1029/2009GL040222.
Vinther, B.M., Clausen, H.B., Fisher, D.A., Koerner, R.M., Johnsen, S.J., Anderson, K.K., Dahl-Jensen, D., Rasmussen, S.O., Steffensen, J.P., Svensson, A.M., 2008. Synchronizing ice cores from the Renland and Agassiz ice caps to the Greenland ice core chronology. J. Geophys. Res. 113 https://doi.org/10.1029/2007JD009143.
Wagner, B., Melles, M., 2002. Holocene environmental history of Western Ymer Ø, East Greenland, inferred from lake sediments. Quat. Int. 89, 165–176.
Weidick, A., 1968. Observations on some Holocene glacier fluctuations in West Greenland. Meddelelser om Grønlands Geologiske Undersøgelse 22.
Young, N., Schweinsberg, A., Briner, J., Schaefer, J., 2015. Glacier maxima in baffin bay during the medievel warm period coeval with norse settlement. Science Advances 1, e1500806.
Zemp, M., Frey, H., Gärtner-Roer, I., Nussbaumer, S.U., Hoelzle, M., Paul, F., Haeberli, W., Denzinger, F., Ahlstrøm, A.P., Anderson, B., Bajracharya, S., Baroni, C., Braun, L.N., Cáceres, B.E., Casassa, G., Cobos, G., Dávila, L.R., Delgado Granados, H., Demuth, M.N., Espizua, L., Fischer, A., Fujita, K., Gadek, B., Ghazanfar, A., Hagen, J.O., Holmlund, P., Karimi, N., Li, Z., Pelto, M., Pitte, P., Popovnin, V.V., Portocarrero, C.A., Prinz, R., Sangewar, C.V., Severskiy, I., Sigurđsson, O., Soruco, A., Usubaliev, R., Vincent, C., 2015. Historically unprecedented global glacier decline in the early 21st century. J. Glaciol. 61, 745–761.