https://doi.org/10.5194/egusphere-2023-737
Preprint. Discussion started: 2 May 2023

Moss kill-dates and modeled summer temperature track episodic snowline lowering and ice-cap expansion in Arctic Canada through the Common Era

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ABSTRACT

Most extant ice caps mantling low-relief Arctic Canada landscapes remained cold-based throughout the late Holocene, preserving *in situ* bryophytes killed as ice expanded across vegetated landscapes. As Arctic summers warmed after 1900 CE, ice caps receded, exposing entombed vegetation. The calibrated radiocarbon ages of dead moss collected near ice-cap margins (kill-dates) define when ice advanced across the site, killing the moss, and remained over the site until the year of their collection. In an earlier study we reported 94 Last Millennium radiocarbon dates on *in situ* dead moss collected at ice-cap margins across Baffin Island, Arctic Canada. Tight clustering of those ages indicated an abrupt onset of the Little Ice Age ~1240 CE, and further expansion ~1480 CE, coincident with episodes of major explosive volcanism. Here we test the confidence in kill dates as reliable predictors of expanding ice caps by re-sampling two previously densely-sampled ice complexes 14 years later, after ~250 m of ice recession. The probability density functions (PDF) of the more recent series of ages matches PDFs of the earlier series, but with a larger fraction of early CE ages. Post 2005 CE ice recession has exposed relict ice caps that grew during earlier Common Era advances, and were preserved beneath later ice-cap growth. We compare 107 kill dates from the two ice complexes with 79 kill dates from 62 other ice caps within 250 km of the two densely sampled ice complexes. The PDF of kill dates from the 62 other ice caps cluster in the same time windows as those from the two ice complexes alone, with the PDF of all 186 kill dates documenting episodes of widespread ice expansion restricted almost exclusively to 250-450 CE, 850-1000 CE and a dense early Little Ice Age cluster with peaks at ~1240 and ~1480 CE.

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Ice continued to expand after 1480 CE, reaching maximum dimensions ~1880 CE, still visible as zones of sparse vegetation cover in remotely sensed imagery. Intervals of widespread ice-cap expansion coincide with persistent decreases in mean summer surface air temperature for the region in a Community Earth System Modeling (CESM) fully coupled Common Era simulation, suggesting primary forcing of the observed snowline lowerings were both modest declines in summer insolation, and cooling resulting from explosive volcanism, most likely intensified by positive feedbacks from sea-ice expansion and reduced northward heat transport by the oceans. The clusters of ice-cap expansion defined by moss kill-dates are mirrored in an annually resolved Common Era record of ice-cap dimensions in Iceland, suggesting this is a circum-North-Atlantic-Arctic climate signal for the Common Era. During the coldest century of the Common Era, 1780-1880 CE, ice caps mantled >11,000 km^2 of north-central Baffin Island, whereas <100 km^2 is glaciated at present. The peak Little Ice Age state approached conditions expected during the inception phase of an ice age, and was only reversed after 1880 CE by anthropogenic alterations of the planetary energy balance.

1. INTRODUCTION

The Canadian Arctic Archipelago (CAA) supports nearly 150,000 km^2 of glaciers and independent ice caps, the greatest concentration of land ice outside Greenland and Antarctica (Medrzycka et al., 2022). As Earth warms, regardless of forcing mechanism, strong positive feedbacks in the Arctic result in greater temperature increases there relative to the planetary average on all timescales (Miller et al., 2010). Consistent with that, the rate of contemporary warming for the Arctic is greater than for most other geographic regions (Rantanen et al., 2022), particularly across the North Atlantic sector (Zhong et al., 2018). All CAA glaciers are now losing mass as summers continue to warm (Gardner et al., 2012; Lenaerts et al, 2013; Noël et al., 2018). In a review of controls on glacier mass balance in the CAA, Koerner (2005) demonstrated that summer temperature explains ~95% of annual mass balance variations over the previous 40 years of record, with the winter mass balance having negligible explanatory power. Consequently, changes in glacier dimensions dominantly reflect changes in summer temperature, and changes in glacier dimensions over time provide one of the most reliable records for the evolution of past summer temperatures. Unlike biological proxies that require evaluating an organism's adaptions to mitigate climate stressors, glaciers have no strategy for survival. If it snows more in winter than melts in summer, non-calving glaciers expand; conversely if more mass is lost by summer melt than is gained through winter snowfall, they recede.

Baffin Island, the largest island in the Canadian Arctic, is an elongate tilted Precambrian block, stretching ~1600 km paralleling the west coast of Greenland. The island was inundated by the Laurentide Ice Sheet (LIS) at the last glacial maximum. Although deglaciation proceeded rapidly under high summer insolation in the early Holocene, our field area (Fig. 1), remained beneath the residual ice sheet 7000 years ago, and probably was not free of Laurentide ice until after 5 ka (Dyke, 2004). The Barnes Ice Cap, at the southern edge

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of our field area, is a remnant of the LIS, and the coastal highlands east of our field area never fully deglaciated in the Holocene (Pendleton et al., 2019a).

Our primary focus is the 30,000 km^2 north-central uplands inscribed in Fig. 1, which rise gradually from sea level at the west coast to ~900 m asl inland, after which they rise rapidly to the high-elevation, deeply dissected fiord country along the east coast, where the interfluves are mantled by large ice caps and cirque glaciers. In contrast, isolated small, thin ice caps currently cover ~100 km^2 of the highest summits across the low-relief uplands. Vegetation is sparse, dominated by lichens and bryophytes, with less common vascular plants. The only woody plants are prostrate species of willow (*Salix*). Our sampled ice caps (Fig. 1) occur across a larger region of about 55,000 km^2.

2. NEOGLACIATION

Although deglaciation of the Laurentide Ice Sheet from Baffin Island continued through most of the Holocene, the orbitally driven decline of Northern Hemisphere summer insolation after ~11 ka resulted in a lowering of snowline and high-elevation ice-cap expansion as early as 9 ka (Lecavalier et al., 2017; Pendleton et al., 2017). Local ice cap and cirque glaciers were advancing by 5 ka (Moore et al., 2001), considered the onset of Neoglaciation, with ice expansion widespread by ~3 ka (Miller, 1973). In response to the continuing decline in Northern Hemisphere summer insolation, Baffin Island ice caps reached their maximum dimensions during the Little Ice Age (1240-1900 CE), the coldest centuries of the Holocene (Miller et al., 2012). Initial dating of Neoglacial ice margins was by lichenometry (Miller, 1973; Locke and Locke, 1977; Davis, 1985) and subsequently based on cosmogenic isotopes in moraine boulders (Anderson, et al., 2008; Briner et al., 2009; Crump et al., 2017). Continuous records of Neoglacial activity on Baffin Island are derived from lake sediment cores (Moore et al., 2001; Miller et al., 2005; Briner et al., 2006; Briner et al., 2016), all confirming maximum ice dimensions during the Little Ice Age. Ives (1957; 1962) was the first to recognize that spectral differences in aerial imagery of northern Baffin Island uplands indicated that sparsely vegetated regions reflected a Little Ice Age (LIA) lowering of snowline that killed all vegetation across, and that early 20th Century warming had reduced an extensive LIA ice complex to a few small remaining ice caps that mantle the highest summits (Williams, 1978). Wolken et al. (2005) evaluated the evidence for and against permanent ice cover explaining the sparsely vegetated zones, concluding that permanent ice cover was indeed the most likely cause. Although re-vegetation is rapidly occurring, the spectral signal remains preserved in recent satellite imagery (Fig. 2), although less distinct than in imagery from the 1950s. Andrews et al. (1975; 1976) used the light-toned areas on air photos and satellite imagery to produce the first map of perennial ice-cover over north-central Baffin Island at the peak of the Little Ice Age. However, the dating of the LIA snowline minimum remains uncertain.

Falconer (1966) was the first to recognize that the receding upland ice caps were frozen to their beds. Rather than active erosive agents, they are exceptional preservation agents. Bedrock striae appearing as contemporary ice recedes, show ice flow to the NE

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around all margins of current ice caps, reflecting only regional Laurentide Ice Sheet flow at the last glacial maximum (Dyke, 2004). Striae, relict patterned ground (Falconer, 1966), delicately perched erratic boulders appearing as contemporary ice surfaces lower (Fig. 3), and the lack of englacial debris confirm cold-based ice. As ice caps recede, they also reveal entombed vegetation in position of growth. The radiocarbon ages of dead moss are "kill dates", that define when ice expanded across the site, killing vegetation, and then remained over the site, preserving the dead plants until ice recession. Once exposed, dead vegetation is efficiently removed by meltwater or wind-blown snow within a few years, although in rare settings devoid of meltwater, they are known to persist for several decades (e.g., Miller et al., 2017; Pendleton et al., 2017). Falconer (1966) reported the first radiocarbon date on *in situ* dead moss collected at the margin of Tiger Ice Cap (Fig. 2), 330±75 yr BP (I-1204), confirming ice expansion occurred during the Little Ice Age, but the precision is too low to include here. Miller visited the same ice cap in 1981, collecting moss at the ice edge that returned a ^{14}C age of 460±25 (GSC-5025), early in the Little Ice Age.

Here we report 186 calibrated radiocarbon dates on dead moss collected at the margins of receding ice caps that mantle summits across the uplands of northern Baffin Island, including all 107 ice-marginal dates within the area inscribed in Fig. 1, and an additional 79 CE dates from 62 other local glaciers and ice caps in the adjacent high-elevation mountainous regions (Fig. 1). Moss "kill-dates" define when an adjacent ice cap advanced across the site during an episode of cooler summers, and remained over the site for decades, killing the moss. We report kill-dates from two densely sampled ice-cap complexes, each sampled ~15 years apart, and compare those ages with Common Era dates from 62 other ice caps within 250 km of the densely sampled ice complexes. Three age clusters defined by the probability density functions (PDF) of the composited 186 kill-dates are compared with mean summer temperatures simulated in a fully coupled climate model for the Common Era (CESM past2k; Zhong et al., 2018). The kill-date clusters correspond to century-scale anomalously cold summers in the model, whereas gaps between kill-date clusters are relatively warm summers in the model. The close correspondence between modeled summer temperatures and changes in ice-cap dimensions, allows us to constrain the age of previously undated, low-elevation, sparsely vegetated zones that define an interval of extensive ice-cap cover at elevations up to 400 m below extant ice caps.

We conclude that ice caps throughout the field area expanded to increasingly lower elevations between 250 and 450 CE, 850 and 1000 CE, and 1240 to 1480 CE, eventually reaching maximum coverage at lowest elevations between 1780 and 1880 CE, effectively resulting in "instantaneous glacierization", a condition predicted to be the prelude to an ice-age cycle. Anthropogenic influences on the planetary energy balance after 1850 CE likely thwarted any possibility of a new ice age, despite optimal orbital configuration for glacial inception.

3. METHODS

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We built on the pioneering work of Ives, Andrews, and Falconer beginning in 2005, collecting *in situ* dead moss, usually within a meter of the ice margin at the time of collection. We concentrated our field studies on the low-relief upland of north-central Baffin Island between the Barnes Ice Cap and Lancaster Sound (Fig. 1), with permission from the Qikiqtani Inuit licensed through the Nunavut Research Institute. We densely sampled two ice complexes mantling the highest uplands: the Serpens Ice Complex (SRP) consisting of three primary ice caps, and the Orion Ice Complex (ORN; both unofficial names), consisting of six ice caps in 2005 (Fig. 2). Additional collections were made in 2009, 2010, 2018, and 2019.

3.1 Radiocarbon dated moss.

Our primary data set is the radiocarbon ages of *in situ* ice-edge dead moss. Ice-margin retreat rates were measured at ~10 m a^{-1}, and ≥1 m a^{-1} vertical lowering (2006-2008 CE), confirming vegetation collected a few meters from the ice margin was exposed during the year of collection. Regions where similar studies use moss kill dates to reconstruct the timing of Neoglacial ice cap expansion include Franz Josef Land (Lubinski et al., 1999), Svalbard (Miller et al. 2017), Greenland (Lowell et al., 2013; Medford et al., 2021; Schweinsberg et al., 2017; 2018; Søndergaard et al., 2019), Baffin Island (Falconer, 1966; Anderson et al., 2008; Miller et al., 2013; Margreth et al., 2014; Pendleton, 2019a&b), Alaska (Calkin and Ellis, 1981), Iceland (Harning et al., 2016), and Antarctica (Yu et al., 2016; Groff et al., 2023).

We target members of the widespread moss family Polytrichaceae, primarily the genus *Polytrichum*, which has a thickened central stem of sufficient mass that a single stem is adequate for a precise AMS ^{14}C date. We previously showed that the ^{14}C concentration in living *Polytrichum* is in equilibrium with that of the contemporary atmosphere, that ^{14}C dates on different *Polytrichum* stems in a single clump are statistically indistinguishable, confirming that they grow relatively quickly, and that ^{14}C dates on *Polytrichum* collected at four sites along 200 m of a modern ice cap margin are statistically indistinguishable (Supplemental Fig. 1). These datasets support the interpretation that the timing of ice-cap expansion can be reliably reconstructed from the ^{14}C kill dates on *in situ* dead *Polytrichum* and other mosses revealed as ice caps recede under contemporary warming. We note, however, that very different kill dates can occur over ice-margin distances of a few tens of meters, when the receding ice margin intercepts a buried, older ice cap, preserved beneath a more recent ice expansion.

Although *Polytrichum* is the most common moss appearing as ice margins recede, if *Polytrichum* is not present, we rely on *Andreaea*, a genus of small rock mosses that commonly grow on siliceous rocks; they represent 9% of the dated samples. Because they are small, multiple strands are required for dating. Another 30% of the dated samples were collected before we gave taxonomic names to moss collections; they are almost all likely to be Polytrichaceae. Radiocarbon dates on co-located lichens are always older and are not included here. In the lab, plant samples are sonicated in de-ionized water, then freeze-dried. From the freeze-dried samples sufficient material (~2 mg) was separated and submitted for dating. Combustion and graphitization of cleaned CO_2 is accomplished in INSTAAR's

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Laboratory for AMS Radiocarbon Preparation and Research (NSRL) and measured at the W. M. Keck Carbon Cycle Accelerator Mass Spectrometry Laboratory at the University of California, Irvine; a few samples were measured at the National Ocean Sciences Accelerator Mass Spectrometry at Woods Hole Oceanographic Institution. Radiocarbon dates were calibrated using OxCal 4.2.4 and IntCal20 (Bronk Ramsey, 2009; Reimer et al., 2020) and expressed in years CE. Some of our dates were reported previously using IntCal13, resulting in small changes in the calibrated age distributions in this paper. The OxCal program reports the probability as a percentage for each year provided by the calibration process. For collections of interest, we sum the yearly probabilities, then normalize and plot as a probability density function (PDF). Samples from multiple years and from different regions can be aggregated and effectively compared this way. A complete tabulation of sample details and calibrated ages for all 186 samples used in this report is available at the NSF Arctic Data Center (https://arcticdata.io).

3.2 Climate Modeling

We make use of the CESM1-CAM5 (Hurrell et al., 2013) "past2k" transient climate simulation of the Common Era (1-2005 CE) (Zhong et al., 2018 GRL). This simulation has 2° resolution in the atmosphere and land and ~1° resolution in the ocean and sea ice, the same configuration as used for the CESM Last Millennium Ensemble (LME; Otto-Bliesner et al., 2016). We extract the simulated summer (JJA) 2-m air temperature averaged over North Atlantic Arctic land areas (>60°N and between 90°W and 30°E) through the Common Era (1-2005 CE).

The past2k forcing data (Fig. 4) were mostly from the Paleoclimate Model Intercomparison Phase 4 project (PMIP4; Jungclaus et al., 2017) for the Coupled Model Intercomparison Project Phase 6. They include changes in solar irradiance and insolation, volcanic aerosols (Toohey et al., 2016), and greenhouse gas levels (MacFarling et al., 2006). The land-cover forcing for the past2k simulation from 850 CE to 2005 CE is taken from the LME, which implements changes in grassland and cropland as compiled by PMIP3. The land cover forcing from 1 to 849 CE is a superposition of HYDE3.1 cropland changes (Goldewijk et al., 2011) onto the land cover forcing from the LME for 850 CE (see Zhong et al., 2018 for full details).

For us to assess the importance of anthropogenic activities in terminating the Little Ice Age, we also use existing CESM1-CAM5 simulations at ~1° resolution in all model components for 1850-2005. Specifically, we make use of the simulated summer temperatures from four simulations with all forcings (Hurrell et al. 2013), the standard "Historical" forcing used for the CESM1 CMIP5 simulations. In addition, we also use the summer temperatures from three simulations with only "Natural" forcings. The "Natural" forcing experiment is forced only by solar variability and volcanic aerosol emissions, following the CMIP5 protocol (Taylor et al., 2012), and is part of a published dataset of single forcing experiments with the CESM1-CAM for 1850-2005 that were previously used in other studies (Meehl et al., 2020; Taylor et al., 2012; Xu et al. 2022). By using all available ensemble members from these two

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experiments, we can assess the impact of the anthropogenic forcing beyond the influence of internal variability.

4. RESULTS

4.1 Repeat sampling around the ORN and SRP ice complexes

In 2005 we collected dead moss around the margins of the Orion (ORN: 18 dated samples) and Serpens (SRP: 22 dated samples, 4 collected in 2009) ice complexes (Fig. 2). The composite PDF of both datasets showed clusters of similar kill dates (Miller et al., 2012). To test whether those clusters were artifacts of the state of the ice caps during the year of collection, we re-sampled ORN in 2018, and SRP in 2019, after ~250 m of ice recession at both, and with a greater sampling density.

4.1.1 Orion Ice Complex (Fig. 2) The 2005 moss kill-dates (Fig. 5A) cluster between 850 and 1000 CE and 1240 to 1480 CE, with a scattering of dates between the two clusters. The PDF of 33 kill dates collected in 2018 (Fig. 5B) is broadly similar, with dominant clusters between 850 and 1000 CE, and between 1240 and 1500 CE, but also an older cluster between 250 and 450 CE. In the 2005-2018 composite PDF (Fig. 5C) the older cluster reflects moss killed during an early First Millennium expansion of ORN, that likely melted partially after 450 CE, but never completely disappeared. The residual early First Millennium ice cap was subsumed by a late First Millennium expansion beginning ca. 850 CE, and preserved beneath younger ice, only to re-emerge and re-start recession under contemporary warming, exposing plants killed during the early First Millennium ice expansion. A primary conclusion from these data is that summer temperatures over the past century are warmer, on average, than any century since 450 CE. The 2018 glacier where ice-marginal moss produced early First Millennium ages has surface characteristics visible in high-resolution imagery captured for us by Digital Globe that differ in character from the glacier surface farther along the ice margin in either direction where the moss kill dates are much younger, supporting our interpretation that in 2018 a remnant early first millennium ice cap that never melted completely was emerging from beneath a younger ice cap that grew when snowline subsequently fell across the region. The near identical alignment of the two younger 2018 clusters with those from 2005 confirms the utility of moss kill dates as reliable ages of Common Era ice-cap expansion, regardless of collection date.

4.1.2 Serpens Ice Complex The PDFs of the 2005/2009 (22 dates) and 2019 (34 dates) collections from SRP (Fig. 6 A&B) produce clusters in nearly identical time intervals, and that match clusters in the ORN composite PDF (Fig. 5C): 250 to 450 CE, 850 to 1000 CE and 1240 to 1480 CE.

4.1.3 The SRP+ORN PDF (Fig. 4-3) The PDFs for all dates from SRP (Fig. 7A) and from ORN (Fig. 7B) are combined as the SRP+ORN composite (Fig. 7C), consisting of 107 calibrated

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^{14}C dates. That PDF produces three well-defined clusters, reflecting an early first millennium expansion between 250 and 450 CE found in 5 samples from SRP and 3 from ORN. There is virtually no evidence of ice expansion between 450 and 850 CE, after which both ice complexes show expanding margins from 850 to 1000 CE (11 from ORN and 12 from SRP). ORN yielded 6 samples scattered between 1050 and 1160 CE, but with no central tendency, whereas SRP has none. However, both SRP and ORN have large numbers of dates beginning ~1240 CE and ~780 CE is related to calibration. Radiocarbon ages between 1080 and 1180 yr BP have continuing until ~1480 CE (25 from ORN and 37 from SRP), indicating ice margins expanded throughout that interval.

In both SRP and ORN composite PDFs (Fig. 7) a sharply defined peak centered on ~780 CE is related to calibration. Radiocarbon ages between 1080 and 1180 yr BP have median calibrated ages between 850 and 960 CE, but they all also have a small (<~5%) probability of an age clustered tightly around 780 CE. Because 21% of the radiocarbon ages the 109 radiocarbon dates from the two complexes are between 1080 and 1180 yr BP, their calibration creates a low probability but strongly expressed peak clustered tightly around 780 CE. Because of the low probability that this is the actual calibrated age, and lacking other samples in the same time range, we consider it highly unlikely that 780 CE is an interval of expanding ice caps.

We test the potential for variations in atmospheric $^{14}CO_2$ concentrations to have influenced the structure of the SRP+ORN PDF by creating 160 random radiocarbon ages between 1 and 1500 CE with similar uncertainties (±25 years) as our moss dates. That calibrated PDF is compared with the SRP+ORN composite PDF in Fig. 8. There are clear natural biases in the random series, with artificial peaks centered on 660 and 1020 CE, time periods where there is no power in the SRP+ORN PDF, but also peaks at about 1280 and 1480 CE, which likely contribute to the height of those peaks in the SRP+ORN PDF, but do not alter the conclusion that widespread ice-cap expansion occurred between 1240 and 1480 CE.

4.1.4 Testing the SRP+ORN Composite (Fig. 9) The combined SRP+ORN PDF defines sustained Common Era ice-cap expansion limited to three primary intervals: 250 to 450 CE, 850 to 1000 CE and after 1240 CE. The SRP Complex consists of 3 independent ice caps, whereas ORN consisted of 6 independent ice caps sampled in 2005, of which portions of only 3 remain. To test whether the densely sampled ice complexes reflect regional snowline lowering across northern Baffin Island we collected 79 additional dead moss from the margins of 62 other ice caps. Those additional ice caps include other small ice caps situated on low relief landscapes similar to ORN and SRP, but also include low-relief summits within much larger ice complexes where ice-expansion is primarily by vertical thickening rather than ice flow. None of the samples are from outlet glaciers; all are within 250 km of SRP and ORN. We avoid cirque glaciers because response times to snowline lowering are more complex than for small upland ice caps or the vertical thickening of large ice caps, both of which change in direct response to summer temperature and are largely unrelated to ice flow.

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The resultant PDF of those 79 calibrated ages is compared with the PDF of the 107 dates from SRP+ORN in Fig. 9. The isolated ice cap dates (Fig. 9B) cluster primarily in the same three groupings apparent in the SRP+ORN PDF (Fig. 9A), although the early first millennium 250-450 CE cluster is more strongly represented, and the onset extends to near the start of the First Millennium. The three samples that contribute to that tail are from elevations above 900 m asl, higher than ORN and SRP, suggesting that summer cooling began earlier than recorded by SRP+ORN, and that snowline did not drop below their elevations until after 250 CE, or that subsequent summer warmth fully melted all pre-CE advances. The 850-1000 CE cluster is apparent in both data sets, and with similar overall structure, suggesting regional climate forcing. Both data sets lack significant ice expansion between 1000 and 1240 CE, and both exhibit evidence of widespread expansion beginning abruptly ~1240 CE and continuing to ~1480 CE, with at least some of the spikiness of the PDF likely an artifact of the variations in ^{14}C concentration of atmospheric CO_2 across that time range (Fig. 8).

The large cluster of kill-dates 1240-1480 CE suggest snowline declined throughout that interval. However, at multiple sites moss collected in 2005 have kill-dates ~1250 CE, whereas in 2019, after ~200 m of ice recession, ice-margin moss have younger kill dates (~1450 CE), requiring an interval of modest ice recession before a readvance late in that time interval, with preservation of the older moss beyond the receded ice margin. These results suggest this 240-year interval was not characterized by monotonic summer cooling; rather summer cooling was interrupted by at least one episode of warm summers leading to ice recession, without all moss killed by the initial advance eroded from the landscape. Because dead moss are relatively easily eroded, the duration of ice-recession interval must have been brief, no more than a few decades, and followed by rapid re-expansion of the ice cap(s).

The composite record consisting of 186 calibrated radiocarbon dates from 70 different ice caps (Fig. 9C) constrains ice-cap expansion during the Common Era across northern Baffin Island. The well-defined clusters suggest snowline began to decline early in the First Millennium CE, resulting in widespread ice-cap expansion between 250 and 450 CE (21 dates). There is no evidence from kill-dates for additional ice-cap growth until after 800 CE, with widespread ice-cap expansion between 850 and 1000 CE (43 dates). There is little evidence for snowline lowering during Medieval times, between 1000 and 1240 CE, but summer temperatures were also not high enough then to raise the snowline above all ice caps that formed earlier in the first millennium, or long enough, to cause those First Millennium ice caps to disappear. This suggests that Medieval times were certainly not as warm, on average, as the earliest centuries of the Common Era, but also that Medieval times experienced no sustained decades of significant summer cold. 1240 CE marks the start of the Little Ice Age, with clear evidence of intermittent ice-cap expansion in all datasets, continuing for the next 240 years (97 dates).

None of the kill-dates from plateau ice caps pre-date the Common Era. Consequently, if earlier late Holocene episodes of summer cold resulted in long-lived ice caps across Baffin Island uplands, subsequent summers were warm enough and/or long enough to completely melt those ice caps before 1 CE. Our moss dates are relatively silent about the state of the ice

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caps after ~1500 CE. We turn to climate models to derive information on ice cap dimensions after ~1500 CE

4.2 Modeling the evolution of Common Era summer temperature

From the past2k model output we extract the evolution of summer (JJA) temperatures over land areas north of 60°N that lie between 30°W and 90°E, capturing a broad representation of North Atlantic Arctic lands, and including the climate impacts from sea-ice feedbacks. Although climate is often approximated using a 30-year average, cold-based ice caps mantling low-relief landscapes on Baffin Island have a longer, multi-decadal response time, so we use a 50-year running mean for our data-model comparison. The 50-year running mean of summer temperature through the Common Era exhibits a declining first-order trend up to 1890 CE (Fig. 10), consistent with the strong influence from declining NH summer insolation from orbital terms. Because Baffin Island is only a portion of the land included in our extraction of mean JJA summer temperatures (Fig. 10 Inset), the absolute values of modeled temperatures are unlikely to match those of our field area, and we evaluate the relative changes in summer temperature without regard to absolute values.

To assess how well the past2k summer temperatures represent the summer temperature evolution in the study region, we compare the past2k summer temperatures with a 50-year record of summer temperature from the only Baffin Island interior weather station, Dewar Lakes, ~350 km SE of our field area, at 500 m asl. The mean summer (JJA) temperature at Dewar Lakes from 1958-2022 exhibits high inter-annual variability (up to 5°C, and a range of almost 9°C) but the 10-year running mean (Fig. 11) shows that relatively warm summers in the 1950s were followed by colder summers in the 1960s and early 1970s, after which variably warming summers dominate the record. The 5-year running mean summer temperature record in our CESM past2k simulation for the period of overlap (1958-2005; Fig. 11) reasonably mimics the Dewar Lakes record, when accounting for the expected much lower variability in the model results due to the averaging over a much larger area.

Both the Dewar Lakes recorded summer temperatures and CESM simulated summer temperatures indicate cold summers in the 1960s and early 1970s. We compare those records with ice-cap retreat rates for three ice caps (WT, SRP, ORN, Fig. 2) from remote images spanning 1957-2020 CE. All three ice caps show reduced rate of aerial decrease between 1960 and the early 1970s (Fig. 12), lending additional confidence to the simulated summer temperatures in the past2k results. Summer warming in both Dewar Lakes and past2k records since 1980 is matched by increased ice-cap recession rates. The close correspondence between past2k modeled summer temperatures and the Dewar Lakes recorded temperatures, and rates of ice-cap recession for WT, SRP, and ORN, provide a measure of confidence in the CESM modeled summer temperatures through the Common Era.

4.3 Comparing modeled summer temperatures with ice-cap kill dates.

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Because summer temperature is the dominant control on Baffin Island glacier mass balance (Koerner, 2005) we expect below average modeled summer temperatures to correspond with episodes of ice cap expansion. The modeled 50-year mean summer temperatures (Fig. 10) contains significant centennial-scale structure. Primary forcings leading to the temperature decline in the model (Fig. 4) include a reduction in Northern Hemisphere summer insolation through the Common Era, small reductions in TSI between 1300 and 1700 CE, and episodes of sulfur-rich explosive volcanism with the associated positive feedbacks from expanded sea ice, additional snow cover over land, and changes in ocean circulation. Anthropogenic impacts become important after ~1850 CE.

A piecewise regression of JJA temperatures (1-1900 CE; Supplemental Figure 2) suggests an initial steep decline from 1 to 900 CE, warming from 900 to 1050 CE, followed by a slower decline between 1050 and 1900 CE. While the piecewise regression captures some of the Common Era structure, it lacks the detail necessary to compare with the kill-date composite PDF (Fig. 9C).

To more directly compare past2k 50-year mean summer temperatures with the information provided by our moss kill dates, we instead calculate multi-century linear regressions of modeled summer temperature across time periods of ice-cap and across intervals lacking evidence of expanding ice caps expansion based on the kill dates (Fig. 10), to better evaluate whether the climatic implications of kill-dates align with modeled summer temperatures. This relaxes the constraints for piecewise regressions that requires individual regressions be continuous. The linear regression through mean summer temperatures between 250 and 450 CE defines a steady decline in summer temperature that accounts for ~25% of the modeled temperature decline between 1 and ~1880 CE. The modeled summer temperature decline is consistent with the earliest kill-date cluster centered on 250-450 CE (Fig. 9C). The modest number of kill-dates between 1 and 250 CE (Fig. 9C) are consistent with the onset of cooling summers 1-250 CE in the model, prior to the main cluster of kill dates beginning at 250 CE.

Declining summer temperatures 250-450 CE were followed in the model by 50 years of warmth before a 50-year strongly negative temperature excursion resulting from 536 CE explosive volcanism (Büntgen et al., 2016) and associated positive feedbacks. A record of ice-cap expansion at that time is not present in our kill dates. We presume snowline dropped and ice caps grew at this time, but the short duration and rapid return to a century of relative warmth after the eruption likely resulted in enough melt that evidence of any 536 CE ice-cap expansion is lost.

High multi-decadal temperature variability without a trend between 450 and 800 CE is predicted to result in changes in snowline, but because cold summers are brief (2 to 3 decades) and followed by warm summers, development of persistent ice caps in unlikely, hence there are no kill dates. However, a strong 100-year cooling trend between 840 and 940 CE produced the coldest sustained temperatures of the First Millennium, and aligns with the second main kill-date cluster between 850 and 1000 CE.

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Modeled Medieval temperatures in our past2k simulation suggest high-magnitude multidecadal variability between 950 and 1200 CE, but without a trend, and at a mean temperature similar to the coldest decades around 400 CE. This is consistent with Medieval times not being warm enough to fully melt ice caps that formed early in the First Millennium. However, the mean Medieval summer temperature is warmer than summers during the 850-1000 CE ice expansion, yet the forty-two sites that have kill-dates between 850 and 1000 CE require ice-cap survival there through Medieval times. The model-data differences through this interval remain unresolved.

From 1240 to 1880 CE summer temperatures decline overall in the model, consistent with our kill dates and with the low snowline required by the elevations of sparsely vegetated lowlands (Fig. 2). Finding ~1240 CE kill dates outboard of 1480 CE dates (section 4.1.4), requires a multi-decadal oscillation of ice-cap dimensions (e.g., a ~1240 CE advance, followed by ice recession without all dead moss eroded away, followed by a subsequent advance ~1480 that allowed ice caps to re-cover moss killed in 1240 CE), and to also kill newly colonized moss both inboard and outboard of the 1240 CE sites. This is consistent with modeled summer temperatures exhibiting significant multi-decadal variability, but never warm enough or long enough to melt all ice that expanded ~1240 CE, early in the Little Ice Age. The lack of kill dates younger than 1500 CE suggests it is unlikely there was any century after 1500 CE with average summer temperatures higher than the 1240-1480 CE mean until 20th Century warming, which is consistent with modeled summer temperatures (Fig. 10).

Modeled summer temperature decreases after Medieval Times are a response to major explosive volcanism ~1240 and ~1440 CE (40 and 50 years, respectively; Sigl et al., 2015). Ten percent of all kill dates (18 of 186 dates; Fig. 9C) have median calibrated ages between 1440 and 1480 suggesting widespread ice expansion then, although the peak at 1440 CE in our evaluation of calibration bias (Fig. 8) suggests that the concentration of kill dates is at least in part an artifact related to variations in atmospheric $^{14}CO_2$ concentrations. Nevertheless, 97 sites with kill dates between 1240 and 1500 CE indicate expansion of ice caps across those two centuries, consistent with large sulfur-rich volcanic eruptions that produce a first-order declining trend in modeled summer temperatures.

There is broad agreement between modeled summer temperatures and the temporal distribution of kill-dates, with the three primary clusters of kill dates centered on episodes of centennial-scale declines in past2k modeled summer temperatures. The cluster of kill dates between 1240 and 1480 CE align with the beginning of a 600-year irregular temperature decline in modeled summer temperature, culminating in the coldest century of the Common Era between 1780 and 1880 CE. This cold century is our estimate for the time that permanent ice caps covered the extensive sparsely vegetated regions first noted 70 years ago (Ives, 1957), a duration that meets that time required to kill all vegetation. Rapid warming in the past2k model after ~1880 CE is consistent with rapid cryosphere recession that reduced the ice-covered area across the northcentral Baffin Island uplands from ~11,000 km^2 in ~1880 CE to ~400 km^2 by 1960 CE, and <100 km^2 in 2022 CE.

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5. DISCUSSION

5-1 Age of maximum Common Era ice-cap expansion

The swaths of sparsely vegetated regions across north-central Baffin Island at elevations well below extant ice caps that define the lowest snowline of the Common Era, have been considered to represent peak cold of the Little Ice Age (Ives, 1962; Andrews et al., 1975; 1976, Williams, 1978; Locke & Locke, 1977), but independent dating of that interval has been elusive. To kill all vegetation requires multi-decadal burial beneath ice. Because the regions that define peak LIA snowline lowering by limited vegetation cover are discontinuous (Fig. 2), more than 100 independent, cold-based ice caps, each a few tens of meters thick, and ranging from ~50 km^2 to a few hundred km^2 in area must have mantled the currently sparsely vegetated uplands during a century or more of coldest LIA summers. The spectral character along the outer perimeters of the sparsely vegetated regions (Fig. 2) is mottled, suggesting a dominant "permanent" ice cap over higher hills, and an outer perimeter where permanent snow lay in landscape hollows long enough to kill vegetation, but not long enough to coalesce into the main ice cap.

Although sparsely vegetated landscapes define the maximum dimensions of ice caps in the Common Era, vegetation killed by that expansion was vulnerable to efficient removal by wind and water erosion as ice caps receded. In 2005 we visited a "sparsely vegetated" region apparent in remote imagery and found that re-vegetation was already advanced. We sampled a dead willow (*Salix*) stem at 629 m asl, lower than any extant ice cap. The ^{14}C age of the stem is "modern" (CURL-7929: 1.16 FM), suggesting that vegetation killed by the lowest LIA snowline is rare or absent, and little chance that radiocarbon can resolve when snowline reached its maximum LIA dimensions.

Recession of ice caps from regions defined by vegetation contrasts was nearly complete by the middle of the 20th Century when remotely sensed imagery first became available. Because the spectral contrast that defines the sparsely vegetated landscape in imagery from the 1950s is much stronger than in contemporary imagery due to on-going re-vegetation, and because the sparsely vegetated landscapes lack gradients in their spectral contrasts, ice recession from maximum LIA dimensions must have been relatively rapid and continuous.

The youngest kill-dates are ~1500 CE, so the timing of maximum snowline lowering must post-date 1500 CE. Having demonstrated that clusters of kill dates align with centuries of simulated cold summers in the past2k simulation, and the past2k simulation aligns with the Dewar Lakes summer temperature record, we use the past2k simulation to estimate the timing of peak snowline lowering defined by sparsely vegetated regions in remote imagery. In the past2k simulation from 1500 to 2000 CE (Fig. 14) an irregular decline in summer temperature reaches lowest values between 1780 and 1880 CE, the coldest century of the Common Era. Subsequently, temperatures rose steadily to the warm decades of the 1930s - 1950s. We conclude that the ice caps that killed vegetation apparent in remote imagery, persisted for at least the 100 years of cold summers between 1780 and 1880 CE.

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532
533 5.2 Comparing Baffin Island kill-date records with a continuous proxy record of ice-cap
534 expansion in Iceland though the Common Era
535 The composite PDF of 186 kill-dates from 70 ice caps (Fig. 9C) provides a direct record of past
536 ice-cap expansion, with changes in mean summer temperature the primary control on ice-
537 cap dimensions. A major advantage of our dataset is that moss kill-dates require no training
538 sets for their interpretation. Although they are discontinuous records, the discontinuities are
539 often important. Consequently, they differ in significant ways from continuous records of
540 glacier-dimension proxies. To evaluate whether the Baffin kill-date record reflects the
541 evolution of summer temperature for the North Atlantic Arctic, as suggested by the strong
542 correlation with climate model results for those land, we compare the discontinuous moss
543 kill-date record that defines time of glacier expansion on Baffin Island with a continuous,
544 annually resolved Common Era record of glacier dimensions from Iceland, also a part of the
545 North Atlantic Arctic lands included in our model.
546 The Iceland record is derived from annual varve thicknesses in a sediment core from
547 glacial lake Hvítárvatn that drains Langjökull, the second largest ice cap in Iceland. Two
548 primary Langjökull outlet glaciers drain into the lake, and during the Little Ice Age, terminated
549 in the lake. Varve thicknesses are a direct measure of glacier dimensions; as the glacier grows,
550 more erosion occurs, and consequently, thicker annual sediment packets in Hvítárvatn.
551 Because Hvítárvatn freezes in winter, and Langjökull occupies easily eroded basaltic terrain,
552 Hvítárvatn sediment is characterized by high sedimentation rates and consequently
553 measurable annual varve couplets for the past 3000 years or more (Larsen et al., 2011).
554 The Hvítárvatn varve-thickness record over the Common Era is plotted against the
555 moss kill-date record in Fig. 14. Varve thicknesses are low, but easily measured, at the start
556 of the Common Era. There is a slight increase in varve thickness 250-450 CE, coincident with
557 the first cluster of kill-dates. A more prominent spike in varve thickness occurs between 850
558 and 950 CE is coincident with the second Baffin Island kill-date clusters between 850 and 1000
559 CE. The final kill-date cluster (1240-1480 CE) is represented in the Hvítárvatn varve record by
560 increasing varve thicknesses that double between 1240 and 1480 CE. Our estimated age for
561 the lowest Baffin Island snowline (1780-1880 CE) is matched by Langjökull varves remaining
562 relatively thick after 1480 CE until thickening rapidly between 1780 and 1880 CE, when the
563 outlet glaciers terminated in the lake (Larsen et al., 2011). Although historical records confirm
564 Langjökull has been receding since ~1900 CE, Hvítárvatn varves remain thick because
565 Langjökull is much thicker (~600 m) than Baffin Island upland ice caps, so has a longer time
566 constant when responding to large increases in summer temperature.
567 The commonalities between the Hvítárvatn varve thicknesses record as a proxy for
568 Langjökull dimensions and moss kill-dates representing episodes of Baffin Island ice cap
569 expansion, suggests both datasets provide a similar record of regional North Atlantic Arctic
570 summer climate through the Common Era. Most other Arctic Common Era summer
571 temperature reconstructions are pan-Arctic (e.g. McKay and Kaufman, 2014; Werner et al.,
572 2018), and show summer temperature variability through the First Millennium, but without

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a trend, which differs from our records that show overall summer cooling through the same time interval. The difference is likely because those records are pan-Arctic, whereas our results are limited to the North Atlantic Arctic where the sea ice feedback is stronger, and climate modeling shows pan-arctic spatial gradients in response to natural forcing over the Common Era (Zhong et al., 2018).

5.3 Evolution of Baffin Island snowline through the Common Era

The PDF of all moss kill-dates (Fig. 9C) defines three episodes of ice-cap expansion. We derive an estimate of the corresponding snowline during each expansion from the mean elevation of the lowest 20% of all dated sites for each interval. We define snowline as the lowest elevation that snow or ice persisted long enough to kill all vegetation (multiple decades), sufficient time for snow to turn into ice. During times of lowering snowline, ice margins expand at many elevations. But because the ice caps are cold based and ice velocities are negligible, the lowest ice margins are the closest quantitative estimate of snowline available. We recognize that ice cap margins defined by their kill dates may be smaller than their maximum dimensions due to ice recession before re-burial under a younger ice cap, but those elevations remain the best estimate for the minimum snowline lowering. We estimate snowline at the peak Little Ice Age from the lowest persistent elevation of remnant vegetation trimlines (Fig. 2), and assign an age of 1880 CE to that snowline, as discussed in section 5.1.

During the Common Era snowline declined more than 400 m, from an initial elevation above 900 m asl to the sparsely vegetated hills 450 to 500 m asl. To evaluate how a declining snowline elevation translates to the area of permanent ice, we calculate the distribution of area across elevation for the region of low-vegetation cover outlined in Fig. 1. Reconstructed snowline elevations for the three kill-date clusters are mapped on to the area/elevation relationship (Supplemental Fig. 3). To calculate the area of permanent ice-cover at peak-LIA-cold, we measured the area of sparsely vegetated regions (Supplemental Fig. 4). Fig. 15 illustrates the highly non-linear relationship between snowline decline and resultant areal extent of permanent ice. Although we lack quantitative snowline data between ice-cap advances, snowline could not have risen much above the preceding low snowline for long or all ice from earlier advances would have melted and no dead vegetation would have been preserved.

5.4 “Instantaneous glacierization” and the initiation of the Laurentide Ice Sheet

The large area of the north-central Baffin Island uplands ice covered at the peak of the Little Ice Age relative to the area currently glacierized (e.g., Fig. 2) led to the concept of “Instantaneous Glacierization” (Ives, 1957; Andrews et al., 1975; Ives et al., 1975) when a modest decline in snowline resulted in a dramatic increase in the area beneath ice caps. Subsequent simulations of conditions leading to the initiation of the Laurentide Ice Sheet from an initial interglacial state (e.g., Andrews and Mahaffy, 1975; Clark et al., 1993; Kleman et al, 2002; Birch et al. 2017; 2018) all show initial snow accumulation on Baffin Island with the albedo feedback providing an important acceleration of ice-sheet growth.

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The dramatic increase in the area under permanent snow/ice-cover in our field area between 1480 and 1880 CE (Fig. 15), during which time snowline dropped 165 m, from ~665 to ~500 m asl, is due to the large area of land between those two elevations (Supplemental Fig. 3). Assuming a standard free air lapse rate of 0.6°C per 100 m, this drop in snowline required a 1°C decrease in mean summer temperature. The direct evidence of "instantaneous glacierization", constrained by the limits provided by moss kill-dates and climate modeling illustrates how a significant albedo increase may occur following a modest decrease in summer temperature, where the resultant snowline descent results in a dramatic increase in permanently ice-covered lands.

5.5 Climate forcing that led to 20th Century ice-cap recession

The combination of direct evidence and modeled summer temperatures points to rapid recession from maximum Common Era ice-cap coverage across north-central Baffin Island beginning in the late 1800s CE. The past2k model simulates all summers after 1900 CE warmer than the mean summer temperature of the 19th Century, with particularly warm summers between 1920 and 1960 CE (Figs. 14 and 16), similar to Arctic-wide temperatures in multiple climate model simulations with all forcings (Fyfe et al., 2013). The dominant forcings that contribute to warming from 1900 to 1960, by which time remote imagery shows that almost all peak LIA ice caps in the field area (Fig. 2) had melted, include a slight increase in solar irradiance through the early 1900s, significant increases in greenhouse gases (GHG: CO_2, CH_4, and N_2O; Fig. 4), partially compensated by a cooling effect from increased anthropogenic aerosols and land-use change, and a slight decline in Northern Hemisphere summer insolation from orbital terms (e.g., Fyfe et al., 2013).

An underlying question is whether the cold century from 1780 to 1880 CE and resultant dramatic increase in ice-covered terrain across north-central Baffin Island was the prelude to a transition into a new ice age, or just an aberrant century that even without anthropogenic trace gas inputs would have returned to warmer summers and ice-cap retreat. The primary impacts on Earth's planetary energy balance over the past century are the opposing effects of anthropogenic aerosols and greenhouse gases, with modest impacts from solar irradiance and volcanism (Myhre et al. 2013; Zhao et al. 2019).

Fyfe et al. (2013) compared pan-Arctic mean annual temperature anomalies over the 20th Century in a range of models that included those using only "Natural" forcings (solar irradiance and volcanism) and those using "Historical" forcings (Natural forcings plus greenhouse gas and aerosol forcings). They found that "Natural" simulations resulted in no trend in circum-Arctic mean annual temperatures, whereas "Historical" simulations produced a modest early 20th Century temperature rise, a slight decline in temperature centered on the 1960s, and a steady warming thereafter. However, the Fyfe et al. (2013) results cannot be compared in detail to our reconstructed changes in the North Atlantic sector, which behave differently from the pan-Arctic average in paleoclimate models (Zhong et al., 2018). Their conclusions also were for mean annual temperatures, whereas we focus on mean summer temperature, the primary control on glacier mass balance (Koerner, 2005).

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To test the potential role of anthropogenic forcing in the observed Baffin Island snowline rise after 1880 CE we compare 15-year running means of summer (JJA) temperature over the North Atlantic Arctic lands in a series of CESM1 ensembles beginning in 1850 CE with "Historical" (4 runs) and CESM1 ensembles with only "Natural" forcings (3 runs; Hurrel et al. 2013, Meehl et al. 2020, Xu et al 2022) in Fig. 16. The "Historical" and "Natural" ensembles track closely until ~1920 CE, when the "Historical" ensemble continues a long increase in simulated summer temperature, with a temporary reduction in the 1950s and 1960s, whereas the "Natural" forcing simulations fail to show additional increases in summer temperature after 1920. Based on this comparison, it is likely that some rise in snowline in the early 1990s would have occurred without any anthropogenic disturbance in the planetary energy balance, but the significant increase in summer temperatures after 1920 CE seen in both the past2k and "Historical" simulations is primarily a response to anthropogenic impacts. This suggests that without Anthropogenic impacts, Baffin Island ice caps would still be widespread at elevations well below 600 m asl, the lowest elevation ice cap in 1985 CE imagery (Fig. 2).

6. CONCLUSIONS

• Small ice caps in the cold, low-relief landscapes of Baffin Island, are frozen to their beds, allowing them to act as preservation agents. Ice cap dimensions are set by summer temperature, with small changes in summer temperature driving significant changes in glacier dimensions. As ice caps expand, they entomb vegetation *in situ*, preserving them until the vegetation is re-exposed when ice caps recede. Radiocarbon dates on moss collected within a few meters of ice-cap margins date times of consistently cold summers resulting in ice-cap expansion through the Common Era, but are restricted to expansions for which subsequent summers never warmed sufficiently and/or long enough to completely melt earlier ice-cap expansions. By inference, summers currently are warmer than any series of multi-decadal summers since the dated ice-cap expansions, placing contemporary warming in a millennial perspective. Kill-dates that define changes in ice cap dimensions are a summer paleo-temperature proxy.

• None of the 107 kill-dates from plateau ice caps ice-marginal dates within the area inscribed in Fig. 1 pre-date the Common Era. Consequently, if earlier Neoglacial episodes of summer cold resulted in long-lived ice caps across the uplands, subsequent summers were warm enough and/or long enough to completely melt those ice caps before 1 CE.

• The composite PDF of 186 moss kill-dates collected from seventy ice caps over a region covering more than 50,000 km^2 yields three multi-century clusters, defining times of widespread ice-cap expansion 250-450 CE, 850-1000 CE, and 1240-1480 CE, continuing without preserved vegetation to a peak snowline lowering between 1780 and 1880 CE,

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resulting in over ~11,000 km^2 covered by ice caps, across a region where ice caps currently cover <100 km^2.

• Clusters of kill dates defining episodes of multi-decadal cold summers, align with an annually resolved continuous record of ice-cap expansion from Iceland through the Common Era.

• The annually-resolved record of varve thicknesses in a lake draining Longjøkull records the irregular expansion of the ice cap during the same time intervals as Baffin kill dates, with both regions reaching maximum ice dimensions 1780-1880 CE, suggesting that the kill-dates likely reflect the evolution of summer temperature throughout the North Atlantic Arctic.

• We compare the times of inferred cold summers that led to snowline lowering and ice-cap expansion with the past2k fully coupled climate model for the Common Era with all major forcings. The model simulates decadally averaged summer temperatures that reveal a first-order decline in summer temperatures over the Common Era, consistent with steadily lower snowline estimates derived from the dated ice caps, and centennial intervals of cold summers that align with the clusters of kill-dates. Peak Common Era modeled cold summers between 1780 and 1880 led to a snowline decline that dramatically increased the snow-covered regions of the Baffin Island uplands (Instantaneous Glacierization) and brought the Eastern Canadian Arctic perilously close to conditions anticipated at the inception phase of the Laurentide Ice Sheet.

• Climate modeling over the interval from 1850 CE to present suggests that rapid ice-cap retreat over Baffin Island beginning ~1880 CE, but especially after 1920 CE, was driven primarily by anthropogenic forcing. We may never know with certainty whether the dramatic snowline decline, and resultant aerial increase in permanently snow-covered lands in Arctic Canada at the peak of the Little Ice Age heralded the onset of the next Northern Hemisphere ice age, only to be reversed by the increase in greenhouse gases and other forcing agents related to the Industrial Revolution. Summer insolation across the Arctic was near its orbitally driven minimum at the peak of the Little Ice Age, and will continue to decrease slightly through the 21st Century. The strong positive correlation between Arctic summer temperatures and summer insolation (Kaufman et al, 2009) suggests that only the additional anthropogenic forcings since ~1850 CE reversed the natural transition into a new ice age.

FIGURES

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Figure 1 Field area, northern Baffin Island. Solid line defines regions for which the area-elevation relation is quantified in Supplemental Figure 1 and used to reconstruct the area covered by descending snowlines (Fig. 15). Dotted white square is the area covered by Fig. 2. Encircled dots show locations of all 186 moss kill-dates (some dots include multiple samples). A portion of the Barnes Ice Cap is in lower right corner. Basemap © Google Earth.

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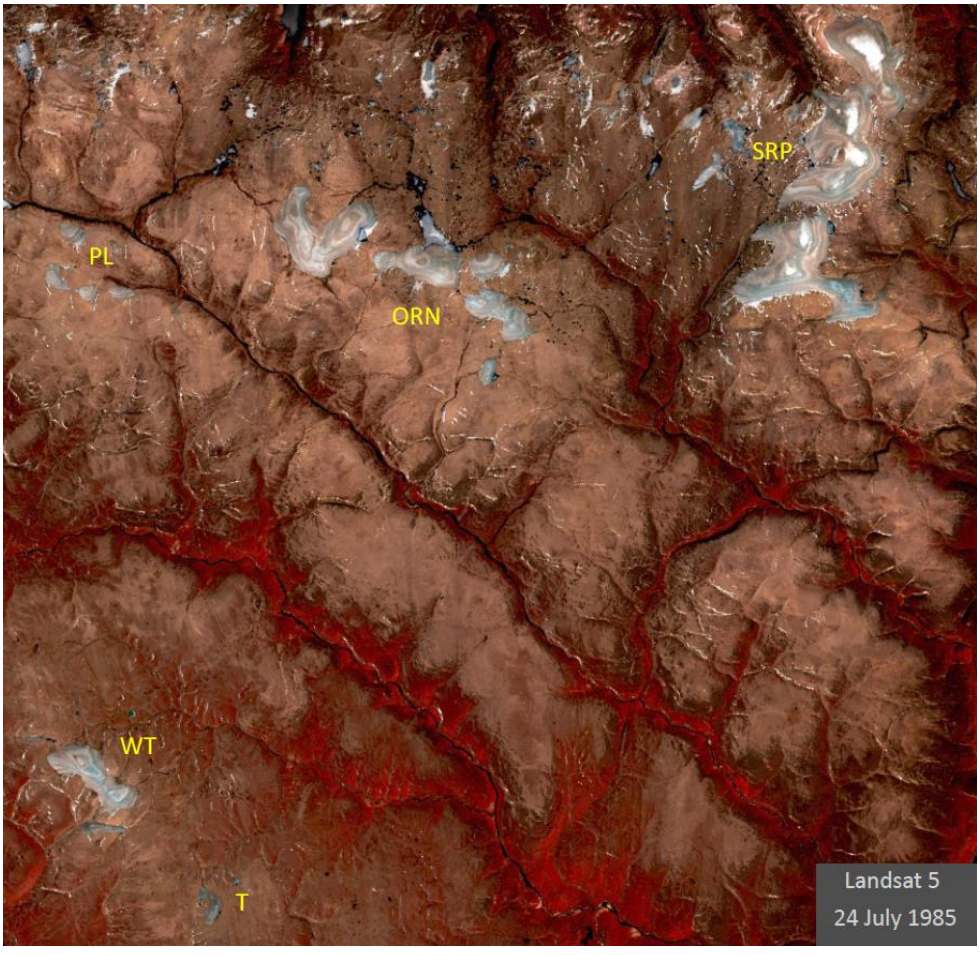

Figure 2 Landsat 5 image from summer 1985, color-enhanced to maximize the spectral contrast that highlights lighter-toned regions that are sparsely vegetated. Those regions were beneath ice caps for a century or more during peak summer cold of the Little Ice Age. Labels are for the Serpens (SRP) and Orion (ORN) ice-cap complexes, and White Tiger (WT) and Tiger (T) ice caps, and Pleiades ice-cap cluster (PL). Location of image shown in Figure 1-1.

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Figure 3 Delicately perched boulder left by the receding Laurentide Ice Sheet that has recently emerged by down-wasting of the local ice cap in background. The preservation of such delicate features demonstrates the lack of internal deformation in the cold-based, ice caps that mantle the low relief hills of the Baffin Island uplands. Photo by GHM.

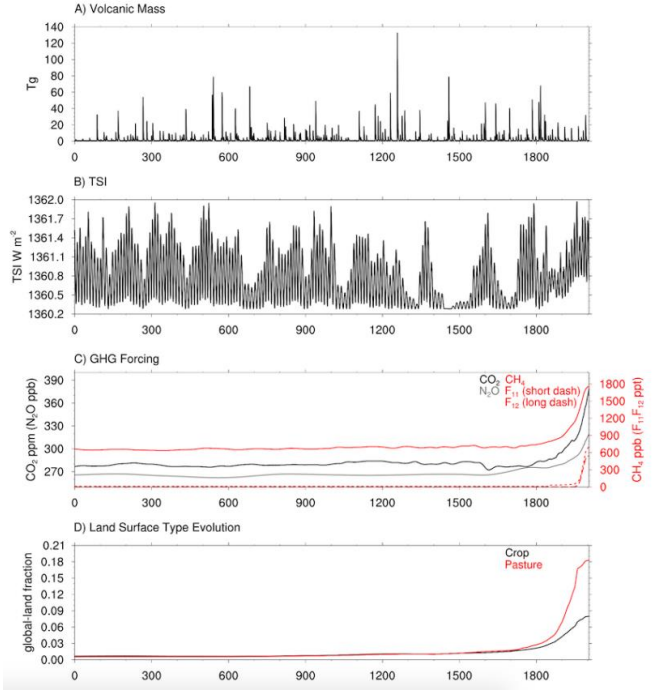

Figure 4 Forcing used for the past2k climate model discussed in the text. A) Mass of sulfuric acid injected into the stratosphere by explosive volcanism in teragrams. B) Total solar irradiance (TSI) in watts /m^2. C) Concentrations of anthropogenic greenhouse gasses released through the Common Era. D) Anthropogenic changes in global land cover through the development of crop (upper line) and pasture (lower line) through the Common Era.

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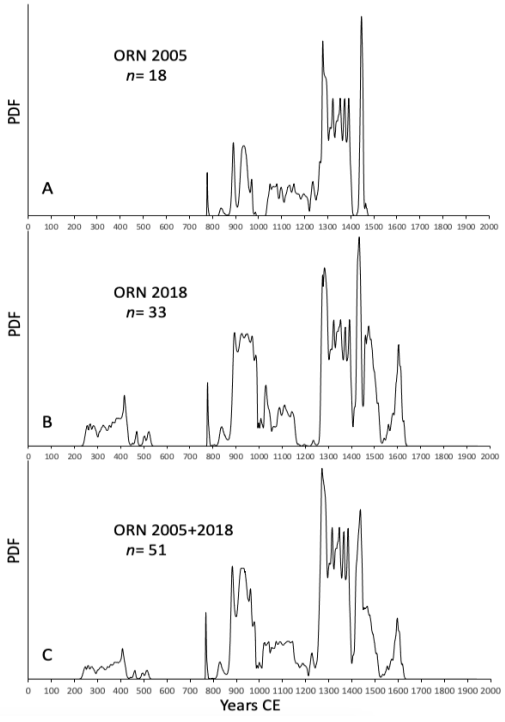

Figure 5 Probability density functions (PDF) for calibrated radiocarbon dates on dead moss collected around the margins of the Orion Ice Complex (ORN, Fig. 2-2), which consists of several adjacent ice caps. (A) The PDF of 18 samples collected in 2005. (B) The PDF of 33 samples collected in 2018 mirrors the distribution from 2005, except for the addition of early First Millennium samples in the latter group, that reflect a buried ice cap now emerging from beneath a younger ice cap that grew on top of it. (C) PDF of all dates from ORN from all years

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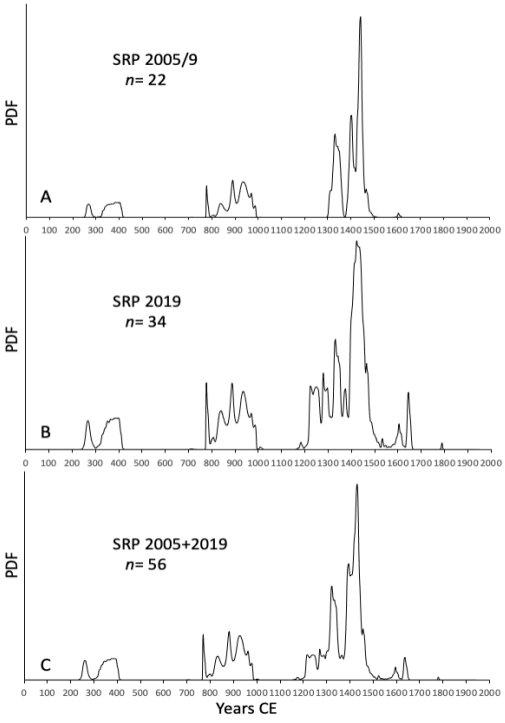

Figure 6 Probability density functions (PDF) for calibrated radiocarbon dates on dead moss collected around the margins of the Serpens Ice Complex (SRP, Fig. 2-2), which consists of three adjacent ice caps. (A) The PDF of 22 samples collected in 2005 (3 from 2009). (B) The PDF from 34 samples collected in 2019. (C) The composite PDF for all kill dates from SRP form three discrete age clusters that exhibit little change between collection dates, despite significant (~250 m) ice-margin recession.

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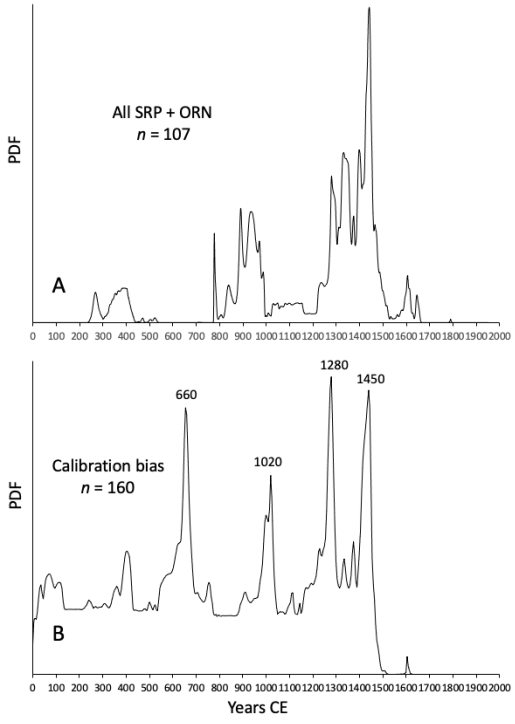

Figure 7 A comparison between probability density functions (PDF) for all calibrated radiocarbon dates on dead moss collected around the margins of ORN (A) and SRP (B) ice complexes (Fig. 2-2) showing the similarities between the two data sets. The composite PDF for the 107 moss kill-dates from both ice complexes (C) defines tightly clustered episodes of ice cap expansion through the Common Era 250-400 CE, 850-1000 CE, and between ~1240 and 1480 CE.

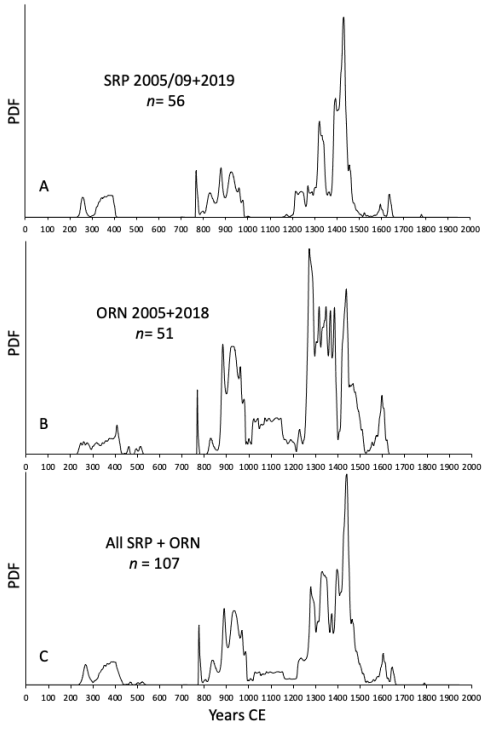

Figure 8 The composite kill-date PDF for SRP+ORN (top panel) compared to the PDF of 160 randomly generated ^{14}C ages between 1 and 1500 CE (bottom panel) to test whether changes in atmospheric $^{14}CO_2$ concentrations and resultant bias in the calibrated ages contributed to the clustering of kill dates in the SRP and ORN composite PDFs. Neither of the earlier calibration peaks (660 and 1020 CE) have any power in the SRP or ORN PDF. However, both the 1280 and 1450 CE peaks coincide peaks in the SRP and ORN PDFs, suggesting that at least some of the magnitude of those two spikes, especially the 1450 CE spike, is due to natural variations in atmospheric $^{14}CO_2$ concentrations rather than an abundance of ice-expansion. However, the calibration artifact does not affect the conclusion that ice caps were advancing at about those times.

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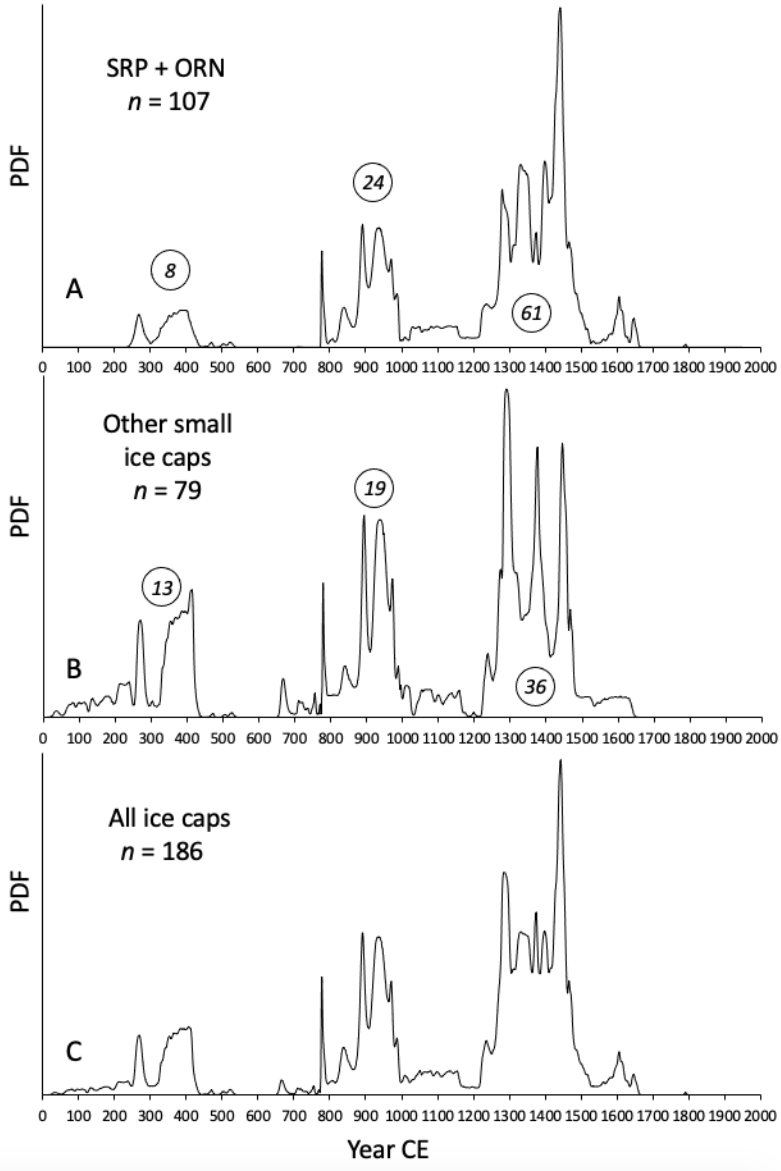

Figure 9 The PDF for all moss kill-dates from SRP+ORN (A) and the PDF for all kill-dates from the margins of all other small ice caps (B) are composited in Panel C. Moss kill-dates that make up Panel B are from 62 different ice caps (Fig. 1-1), all within 250 km of the SRP and ORN ice complexes. The number of dates in each of the clusters is given as a circled number. The similarities between PDFs in Panels A and B is striking, suggesting ice-cap response to climate forcing over the Common Era was uniform across northern Baffin Island, and lends confidence to a comparison of ice-cap response and modeled climate forcing. Location data, Lab-ID, ^{14}C ages and the PDF for each sample in Panel C are available in Supplemental Table S-1. The spike at ~780 CE is a calibration artifact. The height of the spikes at 1280 and 1450 CE may also be amplified by natural biases in the calibration process.

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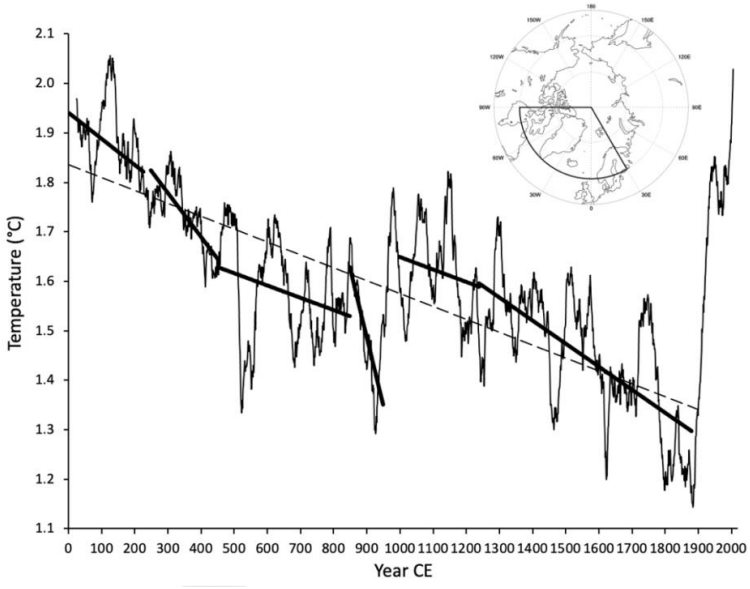

Figure 10 Fifty-year running means of modeled annual summer (J, J, A) temperature over land areas north of 60° and between 30° and 90° west latitude between years 1 and 2005 CE in our CESM past2k simulation. Dashed black line is a least-squares linear regression for the data between 1 and 1890 CE, showing a first order trend of declining summer temperature through the Common Era. Solid lines are least-squares linear regressions through times defined by clusters of kill dates, and the times between kill-date clusters in Figure 9C. The regressions allow a direct comparison of summer temperature trends and relative magnitudes for intervals characterized by widespread ice-cap expansion (predicted cold summers), and intervals lacking evidence of sustained ice-cap expansion (predicted mild summers). The time-intervals defined by these trends can be compared to the reconstructed snowline elevations through the Common Era (Fig. 15).

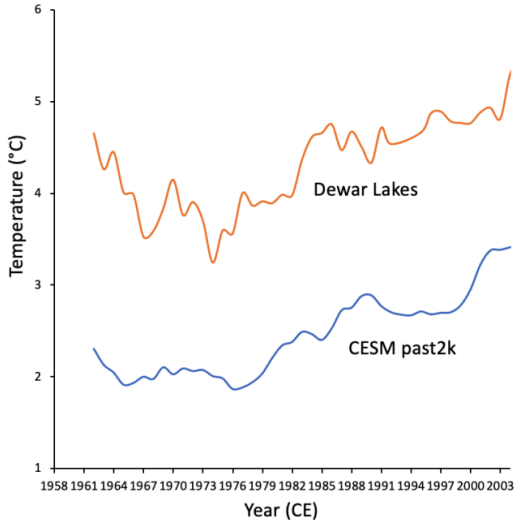

Figure 11 10-year running mean for JJA temperature measured daily at Dewar Lakes since 1958 (upper line) and 5-year running means of simulated JJA 2-m temperature over land north of 60°N and between 30°E and 90°W in our CESM past2k model run (lower line). Dewar Lakes temperature are more strongly smoothed because site-specific records have greater variability than past2k, which is averaged over a large area that naturally smooths the data.

https://doi.org/10.5194/egusphere-2023-737
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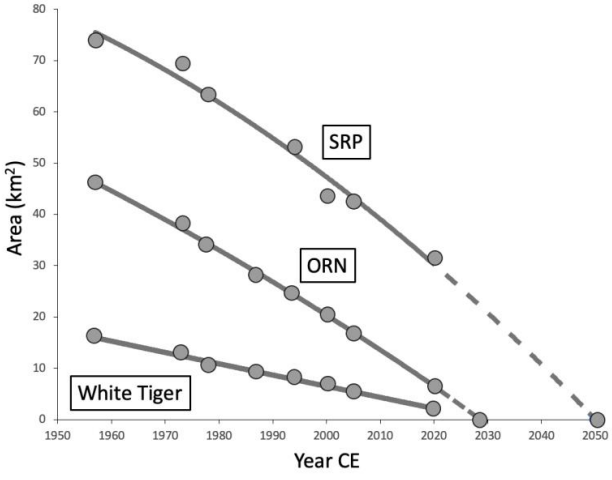

Figure 12 The area of three named caps in Figure 2, between 1957 and 2020. Serpens Ice Complex (SRP; 3 ice caps), Orion Ice Complex (ORN; 6 ice caps, 4 remaining in 2020 CE), and White Tiger (WT, one ice cap). All three show decreased recession rates between 1957 and 1972 CE, consistent with cold summers recorded at Dewar Lakes over the decade between 1958 and mid 1970s CE, followed by steady, or increasing rates of mass loss after 1972,. Areas are derived from vertical aerial photography from 1957 CE and satellite imagery subsequently. Updated from Anderson et al. (2008).

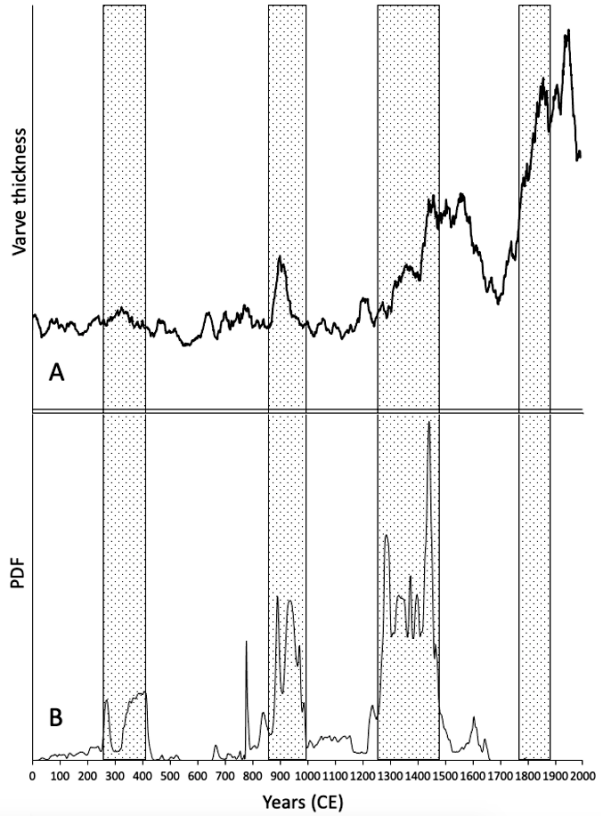

Figure 13 Comparison of 15-year running means of annual varve thickness in glacial lake Hvítárvatn (A), a proxy for the size of Langjökull, second largest glacier in Iceland, and moss kill-date clusters from Baffin Island (B). Stippled boxes show the close correspondence between growth of Langjökull and expansion of Baffin Island ice caps. The varves also confirm that Langjökull attained its greatest dimension in the late 1800s CE, the same time Baffin Island snowline was at its lowest of the Common Era. That close correspondence suggests Baffin kill-dates and Hvítárvatn varves provide a North Atlantic Arctic temperature history for the Common Era.

https://doi.org/10.5194/egusphere-2023-737
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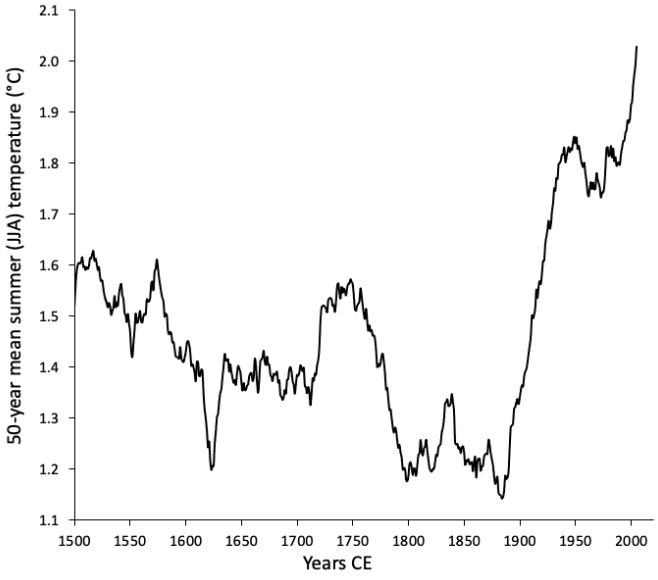

Figure 14 Fifty-year running means of summer (JJA) temperatures in our past2k simulation illustrating the strong trend in declining summer temperatures after 1500 CE (our youngest kill dates), the peak cold century (1780-1880 CE) and rapid summer temperature rise after 1900 CE.

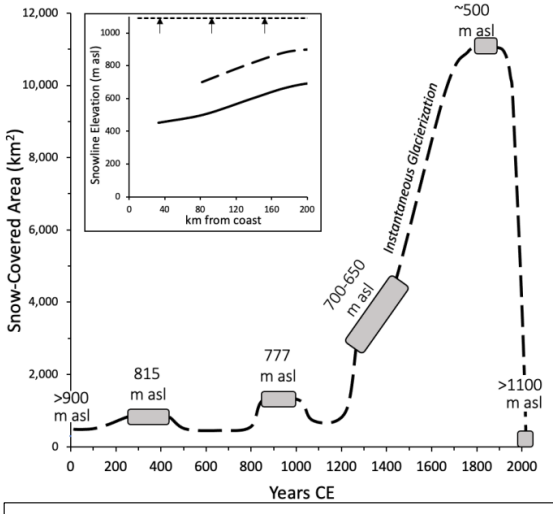

Figure 15 Changes in ice-covered area through the Common Era within our prescribed field area (Fig. 1). Episodes of ice-cap expansion derived from moss kill-dates (Fig. 9C); square boxes) and their associated snowline elevation listed above each box, are plotted against the ice-covered area based on the area-elevation relation within the field area as shown in Supplemental Figure 3. The area covered by ice caps 1780-1880 CE is the summed areas defined by vegetation trim lines (Supplemental Figure 4), with the age derived from the past2k modeled temperatures (section 5.2). Inset show the snowline elevations through our field area for 1880 CE (solid line), 1960 CE (dashed line) and snowline everywhere is above 2020 CE (dotted line), The dramatic increase in area of permanently covered ice as snowline descended through the Little Ice Age was originally postulated by Ives (1957) to result in "instantaneous glacierization" due to the significant albedo increase that might lead to the onset of an ice age (Andrews and Mahaffy, 1975; Clark et al., 1993; Birch et al., 2017).

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7. **SUPPLEMENTAL DOCUMENTS**

8. AUTHOR CONTRIBUTIONS

GHM conceived the project. GHM, SLP, SJL, JPB, JHR, HB, MR collected samples, AJ and YZ were responsible for climate modeling with assistance from ÁG and GHM. SJL and JRS were responsible for the ^{14}C dating. GHM wrote the manuscript with help from all authors.

9. COMPETING INTERESTS: The authors declare that they have no conflict of interest.

10. ACKNOWLEDGEMENTS

We gratefully acknowledge the Qikiqtaani Inuit and the Government of Nunavut for permission to access their land through research permits issued by the Nunavut Research Institute. We thank the Inuit of Qikiqtarjuaq for friendships, advice, and field guides over many decades, and members of the Nunavut Research Institute for assistance in Iqaluit. Battelle ARO and Polar Continental Shelf Program, Government of Canada, provided logistical support. We thank the INSTAAR Laboratory for AMS Radiocarbon Preparation and Analysis for sample preparation, and The Keck Carbon Cycle AMS Laboratory University of California, Irvine, for radiocarbon measurements. Digital Globe acquired high-resolution imagery for some of our ice caps through a cooperative agreement with the US-NSF. The field collections and dating were supported by the US National Science Foundation through grants 0454662, 0909347, 1204096, 1418040, 1556627, and 1737715. Past2k climate modeling was funded jointly by a Grant of Excellence (141573-053) from the Icelandic Research Center (RANNIS) and the US National Science Foundation 1204096. The CESM project is supported by the National Science Foundation and the Office of Science (BER) of the U.S. Department of Energy. We acknowledge supercomputing resources for all used CESM simulations, provided by NSF/CISL/Yellowstone and the Oak Ridge Leadership Computing Facility supported by the Office of Science (BER) of the Department of Energy under Contract DE-AC05-00OR22725.

The past2k model output is available at https://www.earthsystemgrid.org/dataset/ucar.cgd.ccsm4.past2k_transient.html while the natural forcing CESM simulations (b.e10.B20NATC5CN.f09_g16) are available here (https://portal.nersc.gov/archive/home/c/ccsm/www/CESM-CAM5-SF-No/atm/proc/tseries/monthly/TREFHT) and the full forcing simulations (b.e10.B20TRC5CN.f09_g16) are available at https://www.earthsystemgrid.org/search.html?Project=CMIP5&q&page=1&rpp=20&Model=CESM1-CAM5&Experiment=historical.

Dewar Lake climate data are available from https://www.canada.ca/en/environment-climate-change/.

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