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Marine-sediment ^{10}Be and ^{26}Al records of a persistent and dynamic East Greenland Ice Sheet since the Pliocene

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Limited data constrain the multi-million year behavior of ice sheets because repeated ice advances over-run and erode the land-based record of prior glaciations[1-3]. Terrestrial deposits of ancient glacial and interglacial periods are rare, isolated, and poorly dated[4]; thus, they cannot provide a continuous record of ice sheet behavior[2]. In contrast, material shed from continents is preserved as marine sediment that can be analyzed to infer glacial process and history. Here we capitalize on the marine record to show that East Greenland experienced deep, ongoing glacial erosion over the past 7.5 Myr. Our conclusions are based on a progressive, order-of-magnitude decline in the concentration of *in situ*-produced cosmogenic ^{10}Be and the changing ratio of ^{26}Al to ^{10}Be of quartz sand, isolated from ice-rafted debris in sediment cores[5,6] and corrected for radiodecay on the sea floor. The ^{26}Al to ^{10}Be ratio indicates that early Pleistocene East Greenland ice cover was dynamic; in contrast, East Greenland was mostly ice-covered during the mid-to-late Pleistocene. At major climate transitions, the ice sheet expanded into previously ice-free terrain, confirming that the eastern Greenland Ice Sheet consistently responded to global climate forcing. Our results have limitations. Due to mixing of sediment from different sources, transport time of sediment through the glacial system, and amalgamation of samples for analysis, extensive deglaciation during short but intense interglacials could be missed or blurred. Because sediment we analyzed is sourced from East Greenland, we are unable to distinguish during interglaciations between a remnant ice sheet in the Greenland highlands and a more extensive, but diminished continent-wide ice sheet. Our analysis of ^{26}Al to ^{10}Be ratios suggests that similar records from a higher deposition rate coring site could provide a clearer constraint on the behavior of ice sheets during past and thus future interglacial warmth.

Understanding of early Greenlandic glaciation remains fragmentary, uncertain, and for some periods contradictory (Figure 2 and references therein); much of what is known comes from marine sediment. The first presence of ice-rafted debris (IRD) at Ocean Drilling Program (ODP) Site 918 suggests that East Greenland glaciers initially reached the coast ~7.5 Myr[6], whereas the surface texture of quartz at Site 918 suggests that glaciation there began at 11 Myr[7].

IRD data suggest the first large-scale glaciation of Greenland occurred in the latest Pliocene (3.3 Myr), and multiple IRD records indicate expansive Greenland glaciation by 2.7 Myr[8]. Poorly dated shallow marine deposits indicate periods of warmth in Greenland during the later Pliocene or early Pleistocene, some after the initial onset of glaciation[4]. Geochemical and pollen data suggest that southern Greenland was at least partially deglaciated and forested during Marine Isotope Stage (MIS) 11 (~ 400 kyr) and perhaps MIS 5e (~ 130 kyr). In contrast, meteoric ^{10}Be data from silt at the base of the GISP 2 ice core (east-central Greenland) are consistent with continuous cover there by cold-based, non-erosive ice for millions of years[9], an assertion supported by noble gas measurements[10].

The concentration of cosmogenic nuclides in Earth materials reveals near-surface history[11,12]. In non-glaciated terrain, cosmic rays bombard Earth and produce ^{26}Al and ^{10}Be in mineral lattices. Production rates and nuclide concentrations decrease exponentially within a few meters of the surface; weak muon interactions continue at lower rates for tens of meters[13] (Figure 3a). Covering a landscape with ice stops cosmogenic nuclide production in the underlying rock. Subsequent glacial erosion first removes the most highly dosed, near-surface material before excavating rock from depths containing progressively lower isotope concentrations (Figure 3b).

Thermal conditions at the ice sheet bed control its ability to erode, incorporate, and transport rock and sediment. Warm-based ice (at the pressure melting point) can effectively erode rock and transport sediment to and off the coast[14]; thus, the isotopic record we present here is strongly biased toward areas of the ice sheet that were warm based[15]. Cold-based ice, below the pressure melting point, is frozen to the bed and generally non-erosive[16]; it buries and preserves ancient landscapes rather than eroding them.

The ratio of the cosmogenic nuclides ^{26}Al and ^{10}Be provides additional information about burial after initial exposure[12]. Because ^{26}Al ($t_{1/2}$, 0.71 Myr) radiodecays more rapidly than ^{10}Be ($t_{1/2}$, 1.39 Myr), burial of previously exposed material will, over time, lower both the ^{26}Al/^{10}Be ratio and the concentration of both isotopes (Figure 3b). ^{26}Al and ^{10}Be are produced at a ratio of 7.3 ± 0.3 (1σ) near sea level and at high latitude in Greenland (Methods); thus, measured ^{26}Al/^{10}Be ratios <7.3 are diagnostic of burial for significant periods of time ($>\sim200$ kyr).

We measured cosmogenic nuclide records spanning the last 7.5 and 2.6 Myr at ODP Sites 918 and 987 (Figure 1; ED Figures 1, 2). Starting in the Miocene at Site 918, decay-corrected ^{10}Be concentrations generally decrease as sediment gets younger, reflecting the progressive glacial erosion of once-stable regolith and bedrock in East Greenland (Figure 4c). Concentrations of ^{10}Be, corrected for radiodecay on the seafloor, are high (135,000 atoms g^{-1}) in the oldest glacial sediment[6] (~7.5 Myr; sample 918-30) and indicate that the pre-glacial East Greenland landscape was eroding at ~22 m/Myr (Methods). As sediment and rock were removed from the landscape by glacial erosion, material that was deeply shielded in pre-glacial times, and thus less dosed by cosmic radiation, was incorporated into basal ice and carried offshore. We infer that the decrease in ^{10}Be concentration was driven by glacial erosion because IRD at Site 918 and at other sites in East Greenland[2,5] indicates the presence of glaciers eroding rock, extending to the sea, and supplying the coarse, sand-sized sediment that we analyzed[6]. A general increase in the intensity and/or spatial extent of glaciation after 7 Myr is supported by rising accumulation rates of coarse sediment over time at Site 918 (Figure 4b)[6].

By the late Pliocene (~3 Myr), when other records suggest the first major ice expansion over most of Greenland (Figure 2), decay-corrected ^{10}Be concentrations are more than an order of magnitude lower than at the beginning of the record, reaching a minimum of 12,000 atoms g^{-1} at 2.8 Myr. These data are consistent with warm-based glaciers having eroded parts of East Greenland for much of the Pliocene, and provide direct evidence for the model- and core-based supposition[8,17] that the Greenland Ice Sheet has been present and eroding East Greenland since at least the Pliocene.

At the dawn of the Pleistocene, decay-corrected ^{10}Be concentration abruptly increases (Figure 4). Sediment deposited at ~ 2.5 Myr had ~140,000 atoms g^{-1} of ^{10}Be, more similar to Miocene-age (7.5 Myr) sediment than to any of Quaternary age (Table SI1). This ^{10}Be-rich quartz suggests Early Pleistocene expansion of the ice sheet into previously unglaciated areas of East Greenland where stable Miocene regolith remained, an interpretation consistent with abundant IRD found at ~ 2.5 Myr both at Site 918[6] and elsewhere in the Arctic[8]. We suspect this spike in ^{10}Be concentration does not represent an interglacial period, such as that indicated by the warm fauna and flora found in Kap København Formation sediment[4], because that interglacial is thought to be short-lived (<20 kyr)[4]. Such a short duration is insufficient to raise ^{10}Be concentrations to levels attained by continuous pre-glacial exposure[18] (Figure 3c).

From 2.5 Myr to 0.8 Myr, the decay-corrected concentration of ^{10}Be generally declines (Figure 4c), reflecting continued erosion of rock and regolith by warm-based areas of the eastern Greenland Ice Sheet. The decay-corrected ^{10}Be record at Site 918 shows parallels with core sedimentology (Figure 4a,b); an overall inverse correlation exists between ^{10}Be concentration and sand content (ED Figure 3), and several previously noted IRD pulses line up with prominent drops in decay-corrected ^{10}Be concentration, such as pulses near 7, 2.8, 1.9, and 0.8 Myr ago (Figure 4b)[19]. These patterns are consistent with periods of intensified glacial erosion that excavated deeper-sourced material containing less ^{10}Be and delivered it offshore.

An abrupt, four-fold drop in ^{10}Be concentration occurs across the mid-Pleistocene transition at 0.8 Myr (Figure 4), a time when the duration and magnitude of glaciations increased[20]. This drop might reflect reduced interglacial exposure and/or increasing erosivity of the ice sheet. ^{10}Be concentrations over the past 0.8 Myr are similar to those in sediments issuing from the western, southern, and eastern ice margin of Greenland today[18,21] (Figure 1 and 4c), except for one brief increase in a sample spanning 400 to 200 kyr. This higher concentration of ^{10}Be could reflect erosion of sediment exposed during the preceding significant interglacial at MIS 11, a hypothesis supported by a slight, coincident rise in the ^{26}Al/^{10}Be ratio, consistent with interglacial exposure (Figure 4c,d). With this exception, the consistently low ^{10}Be concentrations of the mid- to late Pleistocene sediment indicate the existence of a large, stable ice sheet in East Greenland for most of the last million years.

The shorter ^{10}Be record at ODP Site 987 (2.2 Myr) is consistent with the latter part of the Site 918 record. Decay-corrected concentrations are steady and low at Site 987 (Figure 4c, Table SI1), which suggests that IRD there was derived from glacial erosion of material that was deep below the land surface before East Greenland was ice-covered (Figure 3a). Such efficient erosion of the former Miocene land surface is consistent with the location of Site 987 at the outlet of Scoresby Sund, a large fiord complex that is the major outlet for glaciers sourced in the highlands of East Greenland. Glacial erosion likely began early here because this part of East Greenland is thought to be where the ice sheet nucleated in the Pliocene[6] and where it survived even the most extreme Pliocene warm periods[22]. The four-fold higher ^{10}Be concentrations at Site 918 during the early Pleistocene compared to Site 987 suggests that at least some of the sediment delivered to Site 918 over this interval was sourced from southeast Greenland.

Decay-corrected ^{26}Al/^{10}Be data from Site 918 provide additional information about the history of Greenlandic sediment (Figure 4d). The lack of correlation between ^{10}Be concentration and ^{26}Al/^{10}Be (ED Figure 4) suggests that changes in the ^{26}Al/^{10}Be ratio are not driven by long periods of surface exposure, because such exposure would raise both the ratio and ^{10}Be concentration. This lack of correlation is, however, consistent changing sediment source areas and erosion of material with different ^{26}Al/^{10}Be ratios. Changes in sediment source area may be driven by changes in the basal thermal regime of the ice sheet instead of, or in addition to, changes in ice extent. Spatial and temporal differences in ice sheet behavior likely drive where and when subglacial erosion occurs because remote sensing data show that areas of warm (erosive) and cold (non-erosive) ice are closely juxtaposed[23].

^{26}Al/^{10}Be does not decline steadily over time as would be expected if the entire source area of sediment had been completely and continually covered by ice, which would cause preferential loss of shorter-lived ^{26}Al (Figure 4d). Rather, between 2.6 and 1.7 Myr, the decay-corrected ^{26}Al/^{10}Be ratio is mostly consistent with surface exposure (~7.3). Decay-corrected ^{26}Al/^{10}Be ratios, similar to the production ratio, imply that ice did not cover the sediment source area for most of each glacial/interglacial cycle, because ^{26}Al/^{10}Be ratios change significantly only when surfaces are buried for several times longer than they are exposed and for at least several hundred thousand years in total[24]. In contrast, most of the Site 918 record younger than 1.4 Myr has decay-corrected ^{26}Al/^{10}Be ratios ~ 5 with no decrease over time, which is consistent with the excavation of sediment mostly buried under ice. However, at least some of the material eroded from 1.4 Myr to present must have been intermittently exposed during interglacials in order to prevent the ^{26}Al/^{10}Be ratio from falling steadily because of radiodecay. For a short period just after the mid-Pleistocene transition and the largest IRD spike (samples 918-6 and 918-7), the decay corrected ^{26}Al/^{10}Be rises to ~7.3; we infer that the change in climate cyclicity allowed ice in East Greenland to expand into and erode areas not recently ice covered.

The best hope for detecting short periods of deglaciation is the ^{26}Al/^{10}Be ratio. Contemporary Greenlandic river sand, both glacially and non-glacially sourced, has a ^{26}Al/^{10}Be ratio of 7.6±2.1 (1σ, n=5), which is likely the result of landscape re-exposure during substantial mid-Holocene retreat[25] (SI Table 2). Sand deposited in the Keglen delta at Kangerlussuaq during the end of the last glaciation ~7 kyr ago [26] has a lower than production ^{26}Al/^{10}Be ratio (Table SI2) of 4.54±0.58, fully consistent with ratios we measured in marine cores over the last million years. Thus, high precision ^{26}Al/^{10}Be ratio measurements of quartz extracted from a well-dated, high deposition rate core may reveal glacial/interglacial cycles and could be used to better assess the lag time between exposure and marine deposition[27].

Cosmogenic isotopes preserved in marine sediment record progressive erosion of the pre-glacial landscape in East Greenland from ~7.5 to 2.7 My, the first growth of a full ice sheet at ~2.5 Myr, and a significant change in ice sheet behavior at the 0.8 Myr mid-Pleistocene transition. Measuring ^{26}Al with ^{10}Be demonstrates that erosion under the East Greenland Ice Sheet varied over time and space, and suggests that during the early and mid-Pleistocene, the ice sheet in East Greenland expanded into previously ice-free terrain. Considered along with isotopic measurements of contemporary Greenlandic sediment[18], the lack of repeated increases in ^{10}Be concentration or ^{26}Al/^{10}Be associated with interglacial periods during the past Myr suggests that warming was sufficiently short-lived or modest that it seldom caused significant and lengthy reductions in East Greenland Ice Sheet extent.

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Supplementary Information

We include an EXCEL file of tables with all isotopic and core data.

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Author contributions

PRB and JDS designed the experiment. JDS oversaw core sampling. PRB and LBC did and oversaw laboratory work and DHR, SRZ, and PRB made isotopic analyses. PRB, JDS, LBC, and DHR interpreted the data and all authors contributed to the preparation of the paper.

Author information

The authors declare that all data supporting the findings of this study are available within the paper, the methods, the extended data section, and supplementary information. Reprints and permissions information is available at www.nature.com/reprints. The authors have no competing financial interests. Correspondence and requests for materials should be addressed to pbierman@uvm.edu.

Figure legends

Figure 1. **Map of Greenland**. Shown are Ocean Drilling Program sites used in this study, the East Greenland Current, and locations where contemporary sediment samples were collected (K = Kangerlussuaq, N = Narsarsuaq, T = Tasilaq; italicized numbers give average ^{10}Be concentrations in these sediments[18] as well as a value at Scoresby Sund[21]). Ocean bathymetry contoured in 500 m intervals; derived from MMap (https://www.eoas.ubc.ca/~rich/map.html).

Figure 2. **Compilation of findings that constrain the long-term history of the Greenland Ice Sheet.** Red indicates smaller ice sheet; blue indicates larger ice sheet. References for this figure are included in a second version of Figure 2 available in the Extended Data section as ED Figure 5 with the PDF and html versions of this paper.

Figure 3. **Cosmogenic nuclide systematics and sensitivity to erosion, burial, exposure, and mixing.** (**a**) Estimated steady state concentrations of ^{10}Be and the ^{26}Al/^{10}Be ratio as a function of depth below the preglacial surface of Greenland (sea level) for various erosion rates. (**b**) ^{10}Be concentration (solid) and ^{26}Al/^{10}Be (dashed) starting from the 20 m/Myr steady state profile

following 1 Myr of cold-based ice cover and warm-based ice eroding at 20 m/Myr. (**c**) ^{10}Be (solid) concentration and ^{26}Al/^{10}Be (dashed) (starting from the 20 m/Myr steady state profile) following 1 Myr of warm-based ice eroding at 20 m/Myr, but interrupted by either a 10 or 200 kyr interglacial exposure with 20 m/Myr erosion. (**d**) ^{10}Be concentration (solid) and ^{26}Al/^{10}Be (dashed) of sediments mixed from varying proportions of an eroded and long-buried end member (^{10}Be = 10,000 atoms/g; ^{26}Al/^{10}Be = 4.5) and a long-exposed end member (^{10}Be = 50,000 atoms/g, 200,000 atoms/g; ^{26}Al/^{10}Be = 7.3 in both cases). Muon production rates in all panels calculated using the MATLAB implementation in Balco et al.[13,28]. See Methods for details.

Figure 4. **Seven million years of sediment cosmogenic nuclide values from offshore East Greenland.** (**a**) Simplified lithostratigraphy at Site 918[19]. (**b**) Coarse sand (0.25-2 mm) mass accumulation rate at Site 918[19]. Arrow indicates the oldest dropstones in the core[6]. (**c**) Decay-corrected ^{10}Be concentrations at Sites 918 (blue) and 987 (red) with 1σ uncertainties. Black error bar shows the 1σ range of ^{10}Be in modern sediment from the Greenlandic margin[18]. (**d**) Decay-corrected ^{26}Al/^{10}Be at Site 918 with 1σ uncertainty. (**e**) Global marine δ^{18}O record[29]. Note hiatus from missing core section between 1.7-1.4 Myr.

Methods

Hypothesis testing

We use new isotopic data, in conjunction with sensitivity tests, a forward model, and other extant records, to evaluate three hypotheses about the behavior of the East Greenland Ice Sheet that previous data have not been able to address conclusively. For the last 7.5 Myr, we test whether 1) East Greenland Ice Sheet behavior mirrored global climate/ice volume as represented by the marine δ^{18}O record; 2) the efficacy of erosion under the East Greenland Ice Sheet varied over time and space; 3) most interglacial periods were sufficiently short-lived or cool enough that they did not cause significant reductions in East Greenland Ice Sheet extent.

Compilation of deep time history of the Greenland Ice Sheet

Most of what is known about long-term ice sheet history comes from marine sediment records interpreted as global or regional proxies for ice volume or glacial activity. For example, stable oxygen isotope measurements of foraminifera isolated from marine sediment track global ice volume and ocean temperature, but provide little information about the individual behavior of each of the world's major ice sheets[20]. Global sea-level history reflects total ice volume, but in a complex fashion[30] because the record is aliased by local tectonic and glacioisostatic adjustment of land levels[31]. The most robust inferences about the comings and goings of now-vanished ice sheets are based on the presence and provenance in marine sediment of ice rafted debris (IRD) shed from melting icebergs that originated on glaciated continents[2,8]. IRD records are illustrative of when sediment-bearing glacial ice reached the coast, but with few exceptions[32] do not otherwise constrain ice extent[33]. Figure 2 presents our compilation of references relevant to understanding the history of ice on Greenland since the Miocene. References for Figure 2 are included with references in this methods section.

Determining sediment source area

Making accurate inferences about ice sheet behavior on the basis of terrestrial sediment recovered from marine archives requires knowledge of the sediment source area. Multiple lines of evidence indicate that the quartz we isolated was sourced from East Greenland. The East Greenland current (Figure 1) drifts icebergs from north to south over both Sites 918 and 987, which suggests that the IRD we analyzed is dominantly from East Greenland[34]. IRD composition downcore at Site 918 consistently indicates eastern Greenland sediment sourcing for millions of years[35-37]. While there may be some contribution from gravity flows off the continental shelf, sedimentological evidence suggests that most sand at Site 918 comes from ice rafting rather than turbidites[6,19,38,39]. At Site 918, sand is compositionally similar to larger dropstones, which is consistent with an IRD source for the sand[40]. Sediment at Site 987 is likely more locally sourced because drilling was done on the toe of a large subaqueous fan[41]; although some of the 987 sediment may come from the north, most was presumably delivered directly from ice flowing east through Scoresby Sund (Figure 1). In summary, the cosmogenic data we present reflect the history of and processes active in eastern Greenland.

Thermal conditions at the base of the ice sheet are not well known and change over time[42] and space[23,43]. Warm-based ice (the ultimate source of the sediment we analyzed because it is required to erode the material) is most likely to be found in deep troughs, near some ice margins, and where geothermal heat flux is high[23,44,45]. Models suggests that 20-30% of the pre-industrial Holocene Greenland Ice Sheet was warm-based[44], but during the Last Glacial Maximum up to 50% of ice on Greenland may have been warm-based perhaps due to increased thickness[42].

Paired ^{26}Al/^{10}Be approach and the ^{26}Al/^{10}Be production ratio of 7.3

Because cosmogenic nuclides with different half-lives decay at different rates after production ceases, multiple nuclides can be measured in tandem (e.g., ^{10}Be and ^{26}Al) to provide insight about periods of burial. A multi-nuclide approach can thus constrain the timing and duration of burial by non-erosive, cold-based ice[46], which is a process that has likely occurred variably in Greenland over both space and time.

When exposure begins on a fresh surface, the ^{26}Al/^{10}Be ratio is the production ratio of the two nuclides. If a previously exposed surface is buried and shielded from further nuclide production, the ^{26}Al/^{10}Be ratio drops because the 0.71 Myr half-life of ^{26}Al[47] (ref. 47) is shorter than the 1.39 Myr half-life of ^{10}Be[48-50] (refs. 48-50). If a sample is exposed again following burial, production resumes and the ^{26}Al/^{10}Be ratio increases because the production rate of ^{26}Al is greater than that of ^{10}Be. It is important to note that relatively short burial durations (<100 kyr) and/or re-exposure following burial can result in ^{26}Al/^{10}Be ratios that are indistinguishable from the production ratio[51,52] even though the surface has experienced periods of burial lasting tens of thousands of years.

Any inferences stemming from ^{26}Al/^{10}Be ratios are largely dependent upon the assumed ^{26}Al/^{10}Be production ratio, which is a direct function of the production rates of the two nuclides. Although nuclide production rates have long been known to vary across latitude and elevation[11,53], it has generally been assumed that ^{26}Al and ^{10}Be production rates scale similarly, with a resulting production ratio of 6.75 for all locations on Earth's surface[54]. However, recent work has suggested that the production ratio is itself dependent on latitude and elevation because each isotope's production rate scales differently across space[55-57]. Argento et al.[58] used numerical models to estimate a $^{26}Al/^{10}Be$ production ratio of 7.0-7.1 at sea level and high latitude, which is in agreement with the median value of 7.16 calculated from low-elevation (<2000 m) calibration samples presented in the same study. Sites from a range of latitudes and elevations have production ratios ranging from 7.0-7.3, scaled to sea level and high latitude, and using seven different scaling schemes[57]. Atmospheric mass drives the differences in production between nuclides, with elevation likely being more important than latitude[59], although comprehensive studies of global $^{26}Al/^{10}Be$ production have not yet been conducted.

In this study, we place more emphasis on the relative rather than the absolute $^{26}Al/^{10}Be$ in marine sediment over time, such that the assumed $^{26}Al/^{10}Be$ production ratio is less important than in studies inferring absolute exposure and burial durations. However, we base our assumed production ratio on the work of Corbett et al.[60], who quantified $^{26}Al/^{10}Be$ in 24 continuously-exposed bedrock and boulder surfaces at four high-latitude sites in Greenland that were deeply eroded during the last glaciation. They determined a $^{26}Al/^{10}Be$ ratio of 7.3 ± 0.3 (slope of a York linear regression fit to all data with errors in both variables, 1σ), supporting recent modeling work that the production ratio exceeds 6.75. Although the geographic variability of the production ratio is still unclear, we choose to employ the production ratio of Corbett et al. here because the source of the Site 918 and 987 sediments is similar to the latitude range of the calibration samples in their data set.

Sample measurements

We measured ^{10}Be in 30 samples and ^{26}Al in 22 samples spanning the last 7.5 and 2.6 Myr respectively, in sediment cores at Site 918, located in the Irminger basin 110 km southeast of Greenland (63.1°N, 38.6°W, 1800 m water depth). This site was previously used to define the onset of Greenland glaciation based on the earliest occurrence of IRD[6], which is included in our oldest sample. We also measured ^{10}Be in 16 samples from Site 987 spanning the last 2.2 Myr of deposition 130 km offshore of Scoresby Sund and 1200 km northeast of Site 918 (70.5°N, 17.9°W, 1670 m water depth)[61].

Core samples were obtained from the Bremen Core Repository. We disaggregated and wet-sieved sediments isolating the 0.125 to 0.750 mm grain size fraction and used weak acid ultrasonic leaching (0.25 to 0.5 % HF and HNO_3) to slowly dissolve all minerals other than quartz[62]. We amalgamated quartz from subsamples taken over an interval of core until we had sufficient quartz mass (7.8 to 25.3 g) from which to extract and reliably measure ^{10}Be. Thus, samples represent the average ^{10}Be content of quartz present in core sections ranging in length from 0.04 to 91 m (median = 6 m, st. dev. = 19 m). All uncertainties reported in this paper are 1σ.

Age spans for samples range from 0.001 to 2.9 Myr (median = 0.1 Myr, st. dev. = 0.5 Myr). Our marine sediment record of ^{26}Al and ^{10}Be concentrations does not have the temporal resolution to clearly reflect major high frequency changes in Plio-Pleistocene climate, such as the significant interglacials at MIS 11, 9, or 5e. The need to amalgamate sufficient quartz for measuring very low isotope abundances meant that integration of core sediment over depth (and thus time) mixed sand deposited during glacial and interglacial periods; analysis of a core more proximal to the continental shelf might overcome this limitation.

After purifying quartz, samples were dissolved using HF in the presence of ^{9}Be carrier produced from beryl. Sample were processed in batches of 12 including 2 full chemistry process blanks[63]. ^{10}Be measurements were made at the Center for Accelerator Mass Spectrometry at Lawrence Livermore National Laboratory[64,65] and referenced to standard 07KNSTD3110[66] assuming a ^{10}Be/^{9}Be ratio of 2850×10^{-15}. ^{26}Al measurements were made at the Scottish Universities Environmental Research Centre[67] and normalized to the Z92-0222[47] standard with nominal ^{26}Al/^{27}Al ratio of 4.11×10^{-11} . The average blank ratio (^{10}Be/^{9}Be = $4.6 \pm 1.0 \times 10^{-16}$, n= 6; group 1, ^{26}Al/^{27}Al = $8.5 \pm 2.1 \times 10^{-16}$, n= 4; group 2 ^{26}Al/^{27}Al = $14.9 \pm 4.5 \times 10^{-16}$, n= 4) was subtracted from measured ratios, and uncertainties in sample and blank ratios were propagated in quadrature (Table S1).

Replicate preparation of sample 918-17 (918-17X) indicates reproducibility within measurement uncertainty (Table SI1). Statistically identical measured concentrations of ^{10}Be in four samples (987-E to 987-H) collected from different depths in a 70-cm thick IRD-rich layer (4250±370 to 4460±300 atoms g^{-1}) also demonstrate the reproducibility of our measurements (Table SI1).

In all samples, measured ^{10}Be concentrations are low (2,100 to 40,000 atoms g^{-1}), but well above procedural backgrounds. Because of the shorter half-life of ^{26}Al, it is detectable only in younger samples (< 2.6 Myr), and was measured only at Site 918; concentrations of ^{26}Al are also low (9700 to 118,000 atoms g^{-1}; Table SI1), but similarly well above background. Cosmogenic ^{26}Al/^{10}Be ratios at the time of deposition (corrected by core depth-age models) range from ~3.9 to ~7.5 [Table SI1].

Age models for ODP sites 918 and 987 and decay correction procedure

For Site 918, we used established age-depth constraints from St. John and Krissek[19] who applied ages from the timescale of Cande and Kent [68] to magnetostratigraphic[69] and biostratigraphic datums[70,71]. Ages were linearly interpolated between these control points (ED Figure 1). Note that there is an erosional hiatus at 71.1 mbsf, which is estimated to span 1.71-1.39 Myr[19]. We also developed a planktonic δ^{18}O record (N. pachyderma, left-coiling) to refine the age model above the Brunhes-Matuyama reversal (780 kyr) at 45.9 mbsf[69]. 168 stable isotope measurements were made at Lamont-Doherty Earth Observatory, and 11 to 15 tests were used per sample. The δ^{18}O record clearly displays the Holocene and last interglacial, but there is some ambiguity in the identification of other marine isotopes stages, such as MIS 11 and 13 (ED Figure 2).

For Site 987, we developed an age model by linearly interpolating between the age control points reported by the Leg 162 shipboard scientific party[61], which are primarily based on paleomagnetic events (ED Figure 1).

Measured ^{10}Be and ^{26}Al concentrations (Table SI1) were corrected for decay since deposition on the seafloor using these age models and assuming half-lives of ^{10}Be $t_{1/2} = 1.39$ My [50] (ref. 50) and ^{26}Al $t_{1/2} = 0.71$ Myr [72] (ref. 72). Since our cosmogenic nuclide samples were amalgamated from subsamples spanning 0.001 to 2.9 Myr (Table SI3), we used the sand mass-weighted mean age of these subsamples to derive a single integrated age for each cosmogenic sample. Age model uncertainties can alter the absolute value of decay-corrected ^{10}Be concentrations and change the timing of some isotopic shifts, but have minimal impact on the overall structure of the record.

Sensitivity tests

We examined the sensitivity of ^{10}Be concentrations and ^{26}Al/^{10}Be ratios to erosion, burial, exposure, and mixing (Figure 3), assuming sea-level, high-latitude production rates, including production from muons calculated using the MATLAB implementation in Balco et al.[13,28]. Depth profiles were first run to secular equilibrium, which was reached when nuclide production balanced loss via radiodecay and erosion; the latter was simulated by shifting the profile upward each time step in proportion to the prescribed erosion rate (5, 20, or 50 m/Myr). Steady state profiles with higher erosion rates have lower ^{10}Be concentrations since nuclides are shed more rapidly but they have higher ^{26}Al/^{10}Be ratios since nuclides are brought to the surface more quickly and thus have less time to decay in the subsurface (Figure 3a). We simulated cold-based ice cover for 1 Myr by halting production and allowing the 20 m/Myr steady state profile to decay in place, whereas an analogous simulation for warm-based ice cover continued to erode at 20 m/Myr. Surface nuclide concentrations fall much faster under the erosive warm-based ice, and ^{26}Al/^{10}Be ratios also decline more quickly since the erosive ice brings deeper, and thus longer, buried nuclides to the surface. The 1 Myr long warm-based ice simulation was performed again, but interrupted by either a 10 or 200 kyr episode of interglacial exposure (with erosion continuing at 20 m/Myr) halfway through the simulation. In these simulations, because nuclide concentrations were very low prior to the interglacials, both were able to quickly reset the ^{26}Al/^{10}Be ratio to pre-glacial values; however, only the very long (200 kyr) interglacial had sufficient time to fully rebuild nuclide concentrations. Lastly, we modeled the mixing of sediment from low-concentration, low-ratio (eroded and long-buried) and high-concentration, high-ratio (long-exposed) end members to understand how the values we measured in marine sediments might reflect contributions from multiple source areas on Greenland. Nuclide concentrations mix linearly: $C_{mixed} = C_1F_1 + C_2F_2$, where C_1 and C_2 and F_1 and F_2 are the nuclide concentrations and mixing fractions ($F_1+F_2=1$) of the two end members. ^{26}Al/^{10}Be ratios, however, exhibit nonlinear mixing that is weighted by the ratio of the end members' nuclide concentrations, because the greater the number of nuclides one end member contributes relative to the other, the more it influences the mixed nuclide ratio: $R_{mixed} = R_1 (C_1F_1/(C_1F_1+C_2F_2)) + R_2(C_2F2/(C_1F_1+C_2F_2))$.

Our sensitivity tests demonstrate how progressively deeper erosion, interglacial exposure, burial by cold-based ice, and sediment mixing from different sources affect the concentration of ^{26}Al and ^{10}Be in terrestrial sediment exported from Greenland (Figure 3). Such modeling shows that covering a landscape with non-erosive, cold-based ice for 100s of kyr lowers the ^{26}Al/^{10}Be ratio but does not significantly change ^{10}Be concentration because of the long half-life of ^{10}Be in relation to the burial duration (Figure 3b). In contrast, cover by erosive, warm-based ice not only lowers the ^{26}Al/^{10}Be ratio by shielding the bed from cosmic ray exposure, but also lowers nuclide concentrations because it erodes material with previously produced nuclides and incorporates rock or sediment that was once deeply shielded from cosmic radiation. After the upper several meters of rock and soil are eroded by warm-based ice, isotopic concentrations in the resulting sediment are low and relatively insensitive to continued erosion. This is because the concentration of ^{10}Be in sediment produced by glaciers is controlled primarily by the extent of sub-ice erosion into the deep, muon-dominated production zone that extends tens of meters below the pre-glacial land surface where nuclide concentration changes only gradually with depth (Figure 3a). When sediment is the result of mixing of components with different burial and erosion histories, the history of the sediment may be constrained by considering possible end members with different nuclide concentrations and ^{26}Al/^{10}Be ratios, mixed in different proportions (Figure 3d). The ^{10}Be concentrations we measured reflect the erosion-weighted average ^{10}Be concentration of the areas from which they were sourced, while ^{26}Al/^{10}Be ratios are biased toward source areas that had relatively high nuclide concentrations.

^{26}Al and ^{10}Be concentrations measured in contemporary terrestrial Greenlandic sediment

To better constrain the interpretation of cosmogenic nuclide measurements in marine sediment, we collected sediment samples from Greenlandic rivers, moraines, and river terraces and measured their ^{10}Be (ref. 18), and in some cases,^{26}Al concentrations (Table SI2). Sediment sourced from the ice sheet in eastern, western, and southern Greenland both today[18] and at the end of the last glaciation (sampled in well-dated terraces)[18,21] has very low concentrations of ^{10}Be of only thousands of atoms per gram. Sediment in streams draining only areas outside the current ice margin has on average several times more ^{10}Be, which reflects exposure of the land surface to cosmic radiation during the Holocene[18]. Isotope and mass balance calculations indicate that most sediment now being delivered to the Greenlandic margin originates from beneath the ice sheet and not from the deglaciated margin[18].

To complement existing ^{10}Be data[18], we measured ^{26}Al in 4 samples of contemporary river sediment as well as sediment from the Keglen Delta terrace at Kangerlussuaq[26] (sample GLX-08) and another terrace deposited near Narsarsuaq[73] (GLX-34). Sediment in the Keglen Delta was deposited during the deglaciation (~7 kyr ago)[26] and has a ^{26}Al/^{10}Be ratio substantially lower than production (GLX-08, 4.54± 0.58, 1σ). All sediment from modern streams as well as that in the terrace at Narsarsuaq (GLX-34) deposited during a neoglacial readvance about 1.5 kyr[73] (after mid Holocene retreat and exposure of the landscape) has an average ratio of 7.62±2.12 (1σ; n=5) similar to the production ratio. These data imply that at deglaciation, sediment leaving the ice sheet ~7 kyr ago had a lower than production ^{26}Al/^{10}Be, and that exposure during the mid-Holocene, when the Greenland Ice Sheet retreated km to 10's of km inland of the current margin, raised the ^{26}Al/^{10}Be to or near that of production as suggested by our sensitivity tests (Figure 3).

These results imply that short periods (~10-20 ky) of subaerial interglacial exposure, primarily at the margins of the ice sheet, matter little because they only change the nuclide concentration substantially in the uppermost few meters of rock or soil via shallow neutron spallation reactions. However, even short interglacial re-exposure can effectively raise the ^{26}Al/^{10}Be ratio if initial nuclide concentrations are very low when exposure begins (Figure 3c).

Inherent method limitations

The glacial sediment system itself may limit the resolution of the record. Sediment tracing using ^{10}Be unambiguously shows that most sediment delivered to the current-day Greenland margin during the Holocene interglacial is derived from under the ice, has very little ^{10}Be and ^{26}Al, and is not sourced from deglaciated peripheral area[18]. Sediment currently being shed from deglaciated terrain has several times higher concentrations of ^{10}Be than glacially-derived material, but the marginal area is small in comparison to the area still covered by ice[18]. During glacial advances, sediment from previously exposed margins will be incorporated by ice and eventually mixed with long-shielded material and moved offshore. Even though the marginally sourced material has higher concentrations of ^{10}Be, it is overwhelmed volumetrically by material coming from areas that have been long covered by ice and thus limits the marine record's sensitivity to interglacial cosmic-ray exposure.

Calculation of background, pre-glaciation erosion rate for East Greenland

We determined a pre-glacial erosion rate for southeast Greenland from the decay-corrected ^{10}Be concentration in our oldest sample at site 918 (135,000±10,900, 1σ; Table SI1), which integrates sediments from the 20 m of core immediately below the oldest dropstone at 918 identified by Larsen et al.[6]. Assuming the ^{10}Be in this sample was produced at the surface at sea level directly onshore from 918 and experienced no topographic shielding, we obtain an erosion rate of ~22 ± 3 m/Myr using the CRONUS calculator v.2.3[28]. This estimate is relatively insensitive to these assumptions, except for elevation, which would, for instance, double the erosion rate if production occurred at 1000 m asl rather than at sea level.

Site 918 ^{10}Be comparison to core sand content

We measured Site 918 sand (>63 μm) content and binned data over the same depth intervals as the ^{10}Be samples to facilitate comparison. We similarly binned values in the marine benthic δ^{18}O record[29] over the same time intervals as the 918 ^{10}Be record. Regressions, using logarithmic scaling for the 918 ^{10}Be and sand records, show pronounced relationships, with lower ^{10}Be associated with higher sand content and more enriched marine δ^{18}O ($r^2 = 0.52$ in both cases, $p < 0.001$) (ED Figure 3). As 918 sand concentrations likely reflect glacial erosion on land and marine δ^{18}O is a proxy for global ice volume, these relationships are broadly consistent with intensified glacial activity yielding lower ^{10}Be concentrations in East Greenland-derived sediments.

Forward modeling and code availability

As a first-order attempt to reproduce the ODP 918 and 987 cosmogenic nuclide records, we constructed a simple model of Greenland Ice Sheet dynamics and cosmogenic nuclide concentrations driven by three different plausible ice volume reconstructions over the past 5.3 Myr. The model consists of two sets of 10 parallel cosmogenic nuclide depth profiles, as described in the Sensitivity Tests section above, and was initialized using 20 m/Myr steady-state nuclide depth profiles reflecting pre-ice sheet conditions as indicated by the deepest sample in ODP 918. Ice sheet extent was modeled from 0 to 100% in 10% increments by turning nuclide production on or off for the corresponding number of depth profiles at a given time step; e.g., production was on for all depth profiles when ice cover was 0%, but only nine profiles when ice cover was 10%, etc. The time step is 2 kyr. Since the actual GIS extent through time is poorly constrained, we tried parameterizing it with three different time series: a sea level record derived from δ^{18}O variations in the semi-enclosed Mediterranean Sea basin[74], the marine δ^{18}O record of global ice volume and deep ocean temperature[20], and a simulated history of the GIS from an ice sheet model forced by the marine δ^{18}O record[75]. The last time series explicitly gives ice sheet extent; the relationship between the first two series and GIS extent was calibrated by assuming that ice cover was 100% at 12 kyr, 80% today, 50% during MIS 11, 20% during the mid-Pliocene, and 0% in the Miocene.

We used a simple formulation of basal temperature regimes beneath the modeled ice. Because the GIS has roughly equal areas of cold- and warm-based ice today[45], we set the modeled ice cover to also have equal fractions by making one set of depth profiles warm based (erosion rate = 20 m/Myr) and the other set cold based (erosion rate = 0 m/Myr). Spatial variability in basal temperature regimes was introduced by switching the regime of the two sets of depth profiles every 500 kyr; this is not meant to be realistic, but rather to simply help assess the role of this variable in driving cosmogenic nuclide concentrations given that the basal thermal history of the GIS is not known. Erosion rates were 20 m/Myr in ice-free areas. The simulated cosmogenic nuclide values shown in ED Figure 4 represent the material shed from ice-covered, warm-based depth profiles in the model, and assume instantaneous transport to the deep sea.

This forward modelling illustrates the limitations in the approach we present here as well as the uncertainty of assumptions underlying the model (ED Figure 4). Our model reproduces the overall ^{10}Be record for both Sites 918 and 987, but does not capture the fine structure of the 918 data. The ice sheet extent from a model simulation[75] consistently underestimates ^{10}Be and ^{26}Al/^{10}Be, likely because it does not accurately reflect Greenland Ice Sheet dynamism in the Pleistocene. The marine δ^{18}O proxy[20] and sea level proxy[74] generate more realistic ^{26}Al/^{10}Be; the sea level proxy generates the best fit to the ^{10}Be record. We interpret the fine structure (^{10}Be peaks at 2.5, 1.9 and 1.1 Myr), which we cannot model, as changes in the sediment source area to which cosmogenic nuclides are singularly sensitive; most likely these peaks represent expansion of warm-based areas of the ice sheet into terrain that had not previously been eroded.

The MATLAB code files used to generate the forward model are available at https://github.com/shakunj/Bierman-et-al-2016-Nature. The three versions of the model are provided as MATLAB code files with the forcing series representing Greenland Ice Sheet extent through time designated in the model file name (*forward_model_XXXXX.mat*). These input driving series are the deep sea δ^{18}O record[20] (*LR04.mat*), the Mediterranean Sea sea-level record[74] (*med.mat*), and simulated ice sheet extent based on the modeling[75] (*deboer.mat*), all given at 2 kyr resolution over the past 5.3 Myr. Initialized bedrock profiles with steady state ^{10}Be and ^{26}Al concentrations at 1 cm depth increments below the surface assuming a sea-level high-latitude production rate and 20 m/Myr erosion rate are given in *steadystate_10Be_20mMyr.mat* and *steadystate_26Al_20mMyr.mat*. Sea-level high-latitude ^{10}Be and ^{26}Al production rates in 1 cm depth increments below the surface are given in *P10.mat* and *P26.mat*. The file *er_half_Ma.mat* determines which set of bedrock profiles are beneath erosive warm based ice (1) or nonerosive cold based ice (0) at each time step.

Data Availability Statement

All data generated and analysed during this study are included in this published article and its supplementary information files. MATLAB forward model code is available at: https://github.com/shakunj/Bierman-et-al-2016-Nature

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Extended data legends

ED Figure 1. Age-depth models for Sites 918 and 987. Chronostratigraphic constraints[19,40]are identified by symbols.

ED Figure 2. Site 918 planktonic δ^{18}O stratigraphy. (**a**) The global benthic LR04 δ^{18}O stack on its timescale[20]. (**b**) A planktonic (N. pachyderma, left-coiling) δ^{18}O record from Site 646 off southern Greenland, also on the LR04 timescale[76]. (**c**) The planktonic (N. pachyderma, left-coiling) δ^{18}O record from Site 918 on its depth scale. Notable interglacials in the LR04 stack and their interpreted correlatives at Site 918 are numbered, and the location of the Brunhes-Matuyama magnetic reversal in each record is denoted by the red line. The well-resolved Site 646 δ^{18}O record is shown to provide a nearby planktonic record for comparison to Site 918.

ED Figure 3. Comparing Site 918 decay-correct ^{10}Be concentrations to 918 sand (>63 µm) concentrations and marine δ^{18}O over the past 7.5 Myr. All data have been binned to the same age intervals as the ^{10}Be data.

ED Figure 4. A simple forward model of Greenlandic cosmogenic nuclide concentrations and ratios over the past 5 million years. Simulated (colored lines) (**a**) ^{26}Al/^{10}Be ratios and (**b**) ^{10}Be concentrations of glacially eroded material from a box model with ice extent parameterized as a function of (**c**) GIS extent from a full ice sheet model[75], (**d**) marine δ^{18}O (ref. 20), and (**e**) sea level[74]. The colors of the simulated records in panels a and b correspond to the associated driver of the model in c, d, and f. The ice extent parameterization is represented by the blue shading in panels c, d, and e. ODP Sites 918 and 987 cosmogenic nuclide records are shown by 1σ gray shading in panel a and b, and simulated records have been binned to the same resolution. (**f**) ^{26}Al/^{10}Be-^{10}Be relationships in the simulated (colors) and ODP (black) 918 records. See Methods for model details and https://github.com/shakunj/Bierman-et-al-2016-Nature for computer code.

ED Figure 5. Fully referenced version of Figure 2 from paper.

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