Rapid thinning of the Laurentide Ice Sheet in coastal Maine, USA during late Heinrich Stadial 1

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Abstract

Few data are available to infer the thinning rate of the Laurentide Ice Sheet (LIS) through the last deglaciation, despite its importance for constraining past ice sheet response to climate warming. We measured 31 cosmogenic ¹⁰Be exposure ages in samples collected on coastal mountainsides in Acadia National Park and from the slightly inland Pineo Ridge moraine complex, a ~100-km-long glaciomarine delta, to constrain the timing and rate of LIS thinning and subsequent retreat in coastal Maine. Samples collected along vertical transects in Acadia National Park have indistinguishable exposure ages over a 300 m range of elevation, suggesting that rapid, century-scale thinning occurred at 15.2 ± 0.7 ka, similar to the timing of abrupt thinning inferred from cosmogenic exposure ages at Mt. Katahdin in central Maine (Davis et al., 2015). This rapid ice sheet surface lowering, which likely occurred during the latter part of the cold Heinrich Stadial 1 event (19-14.6 ka), may have been due to enhanced ice-shelf melt and calving in the Gulf of Maine, perhaps related to regional oceanic warming associated with a weakened Atlantic Meridional Overturning Circulation at this time. The ice margin subsequently stabilized at the Pineo Ridge moraine complex until 14.5 ± 0.7 ka, near the onset of Bølling Interstadial warming. Our ¹⁰Be ages

are substantially younger than marine radiocarbon constraints on LIS retreat in the coastal lowlands, suggesting that the deglacial marine reservoir effect in this area was ~1,200 ¹⁴C years, perhaps also related to the sluggish Atlantic Meridional Overturning Circulation during Heinrich Stadial 1.

1. Introduction

The largest uncertainty surrounding future sea-level rise concerns dynamic thinning of the polar ice sheets (Pfeffer et al., 2008), which at present may be driven most strongly by ocean warming at their marine margins (Pritchard et al., 2009, 2012; Straneo and Heimbach, 2013). A variety of processes can determine behavior along marine margins, such as unstable retreat of the grounding line on reverse-sloping beds (marine ice sheet instability; Joughin and Alley, 2011), hydrofracturing of ice shelves and unstable ice cliff collapse (marine ice cliff instability; Scambos et al., 2004; Bassis and Walker, 2012), ocean thermal forcing of basal ice-shelf melt (Straneo and Heimbach, 2013), relative sea-level shifts due to changes in ice volume and gravity fields influencing grounding-line position (Gomez et al., 2015), and grounding-zone sedimentation stabilizing ice margins (Alley et al., 2007).

The short instrumental record and the relatively modest changes over this interval are insufficient to assess how these and other factors will affect future ice sheet collapse as climate warming continues. Ice sheet models are rapidly evolving as better representations of the underlying physical processes are incorporated (Schoof, 2007; Gomez et al., 2015; DeConto and Pollard, 2016); these models need data sets against which to test their reliability (DeConto and Pollard, 2016). Pleistocene deglaciations can provide useful case studies to better understand the dynamics of a decaying marine-terminating ice sheet and its interaction with the ocean and atmosphere – if the timing and rate of past ice sheet collapse in coastal settings can be well dated. The deglaciation of the Laurentide Ice Sheet (LIS) in coastal Maine serves as one such example.

The timing of LIS lateral retreat in terrestrial New England has been constrained using many radiocarbon ages (e.g., Thompson et al., 1999; Dorion et al., 2001; Dyke, 2004), cosmogenic nuclide exposure dating in a few locations (Balco et al., 2002, 2009; Balco and Schaefer, 2006; Bierman et al.,

2015; Bromley et al., 2015; Davis et al., 2015), and the New England Varve Chronology in several major river valleys (Ridge et al., 2012). The deglacial chronology remains less certain elsewhere in New England, including coastal Maine (Fig. 1), which was below sea level due to isostatic depression until emergence began at ~14 ka (Retelle and Weddle, 2001; Kelley et al., 2010). Deglacial radiocarbon ages from marine shells are subject to uncertainties in marine reservoir corrections (Kaplan, 1999; Dorion et al., 2001; Retelle and Weddle, 2001; Borns et al., 2004; Thompson et al., 2011; Thompson, 2015). Lake and bog basal radiocarbon ages from above the marine limit typically provide minimum-limiting ages, and may not faithfully constrain deglaciation timing due to lags in vegetation colonization and deposition of organic material (Stuiver and Borns, 1975; Davis and Davis, 1980). Moreover, these data provide little information on the timing and rate of ice sheet thinning, which, for former ice sheets, is notoriously difficult to quantify.

Cosmogenic exposure ages can provide a more direct constraint on ice sheet thickness history than minimum-limiting radiocarbon ages. The timing and rate of thinning can be determined by obtaining a series of cosmogenic exposure ages along vertical transects down mountainsides that melted out of the thinning ice sheet, an approach informally known as glacial ‘dipstick’ dating (Stone et al., 2003). The glacial dipstick method has been used successfully in Antarctica (Stone et al., 2003; Johnson et al., 2014), Europe (Goehring et al., 2008), Greenland (Corbett et al., 2011; Winsor et al., 2015) and Patagonia (Boex et al., 2013), but has yet to be applied in North America with high-resolution sampling.

Here, we present 18 cosmogenic ^{10}Be surface exposure ages of boulder and bedrock pairs down mountainsides in Acadia National Park and six ^{10}Be exposure ages from the Jordan Pond moraine in the park in order to provide direct information about the timing of vertical ice-sheet thinning in coastal Maine. In addition, we also report seven ^{10}Be exposure ages of boulders from the Pineo Ridge moraine complex (PRMC), a large moraine/glaciomarine delta, to date the timing of ice-sheet margin retreat from coastal Maine (Fig. 1). These data indicate that the LIS thinned rapidly in this area during late Heinrich Stadial 1, and that the ice persisted for several more centuries near the coast before abandoning the PRMC approximately coincident with Bølling warming. The ^{10}Be ages are younger than calibrated radiocarbon-

based estimates of deglaciation, suggesting that the deglacial marine reservoir effect in the waters off New England was $\sim 1,200$ ^{14}C years; this value may help refine marine radiocarbon chronologies elsewhere in the region and inform interpretations of paleoceanographic dynamics.

2. Background

2.1 Exposure Dating: Applications and Assumptions

In situ produced ^{10}Be is formed primarily by cosmic ray spallation of oxygen in rock and soil by neutrons, which mostly attenuate within a meter or two below the surface (Gosse and Phillips, 2001). The buildup of ^{10}Be in rock surfaces can thus be inverted to measure how long ago ice retreated and exposed a landscape. Some geological complications can reduce the accuracy of this geochronometer and lead to either overestimates or underestimates of the true age of deglaciation. For instance, nuclides produced during previous episodes of exposure can be inherited if there was insufficient glacial erosion to remove many meters of material, biasing deglaciation ages so that they are too old (Gosse and Phillips, 2001; Briner et al., 2016). Bedrock may be more likely to contain inherited nuclides than boulders, since the latter have been transported and abraded (Bierman et al., 1999); the collection and age coincidence of boulder-bedrock pairs is thus a means for testing the veracity of exposure ages (Marsella et al., 2000; Briner et al., 2003; Corbett et al., 2013). On the other hand, post-glacial shielding by soil and snow cover can shield the underlying surface from cosmic rays, which lowers the production rate of ^{10}Be , causing the apparent age of the sample to underestimate the timing of deglaciation (Schildgen et al., 2005). Likewise, post-glacial subaerial erosion will lead to a younger apparent age because surface material, rich in nuclides, is preferentially lost (Heyman et al., 2011).

2.2 Paleoclimate context and New England deglaciation

The last Northern Hemisphere deglaciation was punctuated by several abrupt warming and cooling events well expressed in the Greenland ice core record (Andersen et al., 2004). These climate events were related to variations in the strength of northward heat transport by the Atlantic Meridional Overturning Circulation (AMOC), thought to be linked to meltwater events from ice sheets (Clark et al., 2001; Thornalley et al., 2010). For instance, freshwater forcing related to initial retreat of the Northern

Hemisphere ice sheets may have weakened the AMOC beginning ~19 ka, triggering Heinrich Stadial 1 (HS1; 19-14.6 ka), a cooling event in the North Atlantic (McManus et al., 2004; Clark et al., 2009). Following the recovery of the AMOC at 14.6 ka, the Northern Hemisphere experienced abrupt warming into the Bølling-Allerød interstadial. The deglaciation of the LIS is inherently linked to these perturbations of the overall warming climate (Balco et al., 2002; Kaplan, 2007; Ullman et al., 2014), though the detailed dynamics of the connections remain uncertain.

Over the last several decades, glacial geologic mapping and cosmogenic exposure dating has revealed the maximum extent and pattern of subsequent retreat of the LIS in New England (Balco et al., 2002, 2009; Ridge et al., 2012). At the LGM, the southeastern margin of the LIS extended into the Gulf of Maine ~600 km beyond the present coastline and onto Georges Bank (Pratt and Schlee, 1969; Schnitker et al., 2001). Cosmogenic exposure dating indicates that the LIS margin initially pulled back from its maximum extent in southeastern Massachusetts at Martha's Vineyard, dated to 27.5 ± 1.6 ka ($n = 8$), and southern New Jersey, dated to 25 ± 1.3 ka ($n = 16$; Corbett et al., in press; these ages and all those reported below are recalculated using the northeastern North American production rate from Balco et al. (2009) and give uncertainties as the standard error of boulder ages and the production rate uncertainty, added in quadrature). Several recessional moraines in southernmost New England also have been dated with ^{10}Be , including the Buzzard's Bay moraine in Massachusetts (20.3 ± 1.0 ka, $n = 10$; Balco et al. (2002)), and the Old Saybrook (20.7 ± 1.1 ka, $n = 7$; Balco and Schaefer (2006)) and Ledyard (20.7 ± 1.0 ka, $n = 7$; Balco and Schaefer (2006)) moraines in southern Connecticut. These ages indicate that retreat was slow and spatially limited until after ~20 ka when Northern Hemisphere summer insolation increased (Clark et al., 2009). It took about 7,000 years for the ice margin to retreat to northern New England and deposit the Littleton-Bethlehem moraine (13.8 ± 0.7 ka, $n = 4$; Balco et al. (2009); Thompson et al. (2017)) in central-northern New Hampshire, and the Androscoggin moraine (13.2 ± 0.7 ka, $n = 7$; Bromley et al., 2015) in northeastern New Hampshire and western Maine.

The New England varve record also has provided significant insight into the retreat of the LIS in central New England, based on the progressive deposition of glacial varves in proglacial ice-dammed

lakes up the length of the Connecticut Valley (Ridge et al., 2004, 2012). The record indicates that retreat began slowly (~50 m/year) in southern New England, but steadily increased to 300 m/year between 14.7 and 14.1 ka during the Bølling in northern New England (Ridge, 2004, 2012). The ice margin receded from the Connecticut valley into Canada by 13.5 ka (Ridge et al., 2012).

2.3 Marine Reservoir Correction in coastal Maine

Because there is a difference between the $^{14}\text{C}/^{12}\text{C}$ ratio of the ocean and that of the contemporaneous atmosphere, a marine reservoir correction is needed to make radiocarbon ages of marine samples comparable to those of terrestrial samples (Waelbroeck et al., 2001). The marine reservoir effect (the offset between the $^{14}\text{C}/^{12}\text{C}$ ratio in the ocean and atmosphere) varies over time as a result of changes in ocean circulation (Mangerud, 1972), ^{14}C production rates in the atmosphere (Stuiver et al., 1986), and ventilation between the atmosphere and ocean (Bard et al., 1994). Modern marine reservoir ages are 400 ± 100 years in the North Atlantic (Bard, 1988).

A reservoir correction must be applied to many of the radiocarbon ages in coastal Maine because they come from marine sediments deposited while the region was isostatically depressed. The ocean inundated the coast and transgressed up to 175 km inland and ~70 m above sea level based on the elevation of topset-forset delta contacts (Thompson et al., 1989). Numerous estimates for the deglacial marine reservoir effect in coastal Maine have been made, including >400 years (Dorion et al., 2001), 600 years (Borns et al., 2004), >1,000 years (Thompson et al., 2011), and 2,000 years (Ridge et al., 2001). Dating of glaciated landforms using cosmogenic nuclides helps address the uncertainty surrounding the marine reservoir correction in coastal Maine (Kaplan, 2007).

2.4 Deglacial History in Maine

The deglacial history of the LIS in Maine has primarily been constrained with minimum-limiting radiocarbon ages from both terrestrial and marine settings. However, in a few locations, radiocarbon ages come from glaciomarine sediments that are interbedded with ice-proximal and ice-distal facies (e.g. Dennison Point, Pond Ridge moraine), and thus closely limit deglaciation (Kaplan, 1999; Dorion et al., 2001). As discussed above, marine samples dated by radiocarbon must be corrected for the marine

reservoir effect, though other ages in this region are from terrestrial materials deposited on bog and pond bottoms (Borns et al., 2004) for which no marine reservoir correction is needed. However, radiocarbon ages measured on bulk sediment also should be treated cautiously, because the samples may contain some proportion of material predating the time of deposition (Grimm et al., 2009).

The sedimentary record also reveals how the ice sheet margin evolved during deglaciation. Seismic reflection profiles combined with marine stratigraphy from the Gulf of Maine (Belknap and Shipp, 1991; Schnitker et al., 2001) indicate that a grounded ice sheet transitioned to a marine calving embayment (Borns, 1985; Hughes et al., 1985; Borns and Hughes, 1977) between 20 and 17 ka (Schnitker et al., 2001). The ice margin is estimated to have reached the present coastline between 17 and 16 ka during HS1 (using a 600 year radiocarbon reservoir correction; see Discussion) (Smith, 1985; Kaplan, 1999; Retelle and Weddle, 2001; Borns et al., 2004; Fig. 1), becoming buttressed on underlying bedrock ledges and sills in shallow marine water resembling a modern tidewater glacier (Tucholke and Hollister, 1973; Crossen, 1991; Kaplan, 1999). Analysis of benthic foraminiferal $\delta^{18}\text{O}$ and faunal assemblages from southeastern coastal Maine indicate that water temperatures were arctic to subarctic during the deglaciation (0-2°C; Dorion et al., 2001). Various deglacial landforms were deposited into shallow water (less than 80 m) along the coast, such as stratified moraines, esker-fed ice-contact deltas, and outwash deltas (Ashely et al., 1999; Borns et al., 2004), collectively forming a coastal moraine belt. Ice deformation features such as folds and faults in stratified deposits indicate that ice was actively flowing during retreat (Koteff and Pessl, 1981; Crossen, 1991; Dorion et al., 2001).

Within the coastal moraine belt there are two particularly prominent ice-contact Gilbert-style deltas, the PRMC and the Pond Ridge moraine (LePage, 1982). Radiocarbon dating of *in situ* shells within the Pond Ridge moraine bracket its age to be between 16.7 and 16.5 ka (13.8-13.6 ^{14}C ka; Kaplan, 1999, no reservoir correction), whereas radiocarbon ages from mollusk shells north and south of the PRMC constrain its emplacement age between ~16 and 15 ka (13.8-13.4 ^{14}C ka; Kaplan (2007); author assumed a reservoir correction of 400 ^{14}C years on the older age and 600 ^{14}C years on the younger age), during the latter part of HS1. Davis et al. (2015) reported a ^{10}Be age of 16.4 ± 1.1 , a ^{26}Al age of 19.8 ± 1.8

and an uncertainty-weighted average ^{10}Be - ^{26}Al age of 17.5 ± 1.1 ka for one boulder from the PRMC (PTK-17), older than the radiocarbon constraints but still during the cold HS1. Crosscutting relationships within the moraine belt in eastern Maine indicate that the PRMC and likely the Pond Ridge moraine represent a readvance and/or reorganization of the ice margin, though the western portion may have been a stillstand (Kaplan, 1999).

Following deposition of the coastal moraine belt, the ice margin retreated abruptly after the Bølling warming, corresponding with a geomorphic transition from moraines to esker systems (Dorion et al., 2001; Borns et al., 2004). Once the ice sheet retreated further inland, relative sea level rapidly fell after ~ 14.3 ka (~ 12.2 ^{14}C ka; Retelle and Weddle, 2001, authors used 600 ^{14}C year reservoir correction) due to glacial isostatic adjustment, reaching a relative sea-level low stand of -60 m at 12.5 ka (Barnhardt et al., 1997). Relative sea level rose rapidly after 12.5 ka to -20 m below present level, then remained nearly static between 11.5 and 7.5 ka, before approaching modern levels in the mid-Holocene (Kelley et al., 2010).

Although the deglacial chronology in Maine is highly dependent on radiocarbon ages, recent cosmogenic nuclide exposure dating in central Maine also offers insight into ice dynamics during the last deglaciation. Some samples collected on the summits of Katahdin, Maine (Fig. 1), as well as Mount Washington, New Hampshire, contain high concentrations of cosmogenic nuclides inherited from prior periods of exposure (Bierman et al., 2015). Such inheritance likely indicates that thin, cold-based ice covered these peaks, causing little erosion. Nonetheless, other samples on the Katahdin edifice (~ 1300 - 1600 m) appear to have been deeply eroded before exposure at deglaciation as they yield an average age of 15.3 ± 1.1 ka ($n = 6$), indistinguishable from boulders on the Basin Ponds moraine, part way down the mountain (~ 750 m; 16.1 ± 0.9 ka; $n = 5$), as well as Pockwockamus Rock in the nearby lowlands (150 m; 14.5 ± 0.8 ka). Together, these data suggest rapid drawdown of the ice sheet between 16 and 15 ka in central Maine (Davis et al., 2015).

3. Study Sites

We used ^{10}Be to date deglaciation in three areas of coastal Maine: vertical transects in Acadia National Park, the Jordan Pond moraine in Acadia National Park, and the PRMC, about 50 km northeast of Acadia National Park. All sample locations in Acadia National Park were above the marine limit (70 m).

Acadia National Park is located on Mount Desert Island, along the south-central coast of Maine (Fig. 1). Several peaks rise above 350 m, including Cadillac Mountain (466 m), the highest coastal point in the eastern United States. Most of Acadia National Park is underlain by the Devonian pink coarse-grained Cadillac Mountain Granite (Gilman et al., 1988). Two basal radiocarbon ages have been reported from Sargent Mountain Pond (336 m; above the marine limit) on Mount Desert Island, providing a minimum constraint on when the area emerged from beneath the ice sheet. Lowell (1980) dated bulk sediments 1.33 m above the base of a core to 15.7 ± 0.7 ka (13.25 ± 0.26 ^{14}C ka), while Norton et al. (2011) more recently obtained a similar basal age of 15.7 ± 0.4 ka (13.26 ± 0.05 ^{14}C ka) on unspecified organic sediment 7 cm above refusal from a different core.

Jordan Pond is located in central Acadia National Park between Penobscot Mountain and Pemetic Mountain at an elevation of 83 m. Once the ice sheet reached coastal Maine and thinned sufficiently to expose peaks, ice lobes formed in the major valleys (Gilman et al., 1988). One such lobe deposited an end moraine just above the marine limit that serves as a natural dam at the southern end of Jordan Pond (Graham, 2010). The Jordan Pond moraine is thus the most proximal moraine to the coast that was not affected by marine inundation.

The PRMC is ~100 km long and located ~20 km from the current coastline in southeastern Maine at an elevation of ~70 m, just below the marine limit (Fig. 1; Borns et al., 2004). Parts may have initially been under water, but any submergence was likely brief due to rapid postglacial rebound (Borns et al., 2004; Kelley et al., 2010). Morphologically, the PRMC is a sharp-crested morainal bank that spatially transitions to a flat-topped morainal bank (Hunter and Smith, 2001); it was formed at the grounding line in marine conditions, as demonstrated by the presence of marine silt and clay, a change in sediment facies between the ice-proximal and ice-distal side of the moraine, and stratified beds (Kaplan, 1999).

4. Methods

4.1 Study Design and Field Methods

We collected nine paired samples of boulder and bedrock along elevation transects from just above the valley floors to the highest peaks in Acadia National Park, including Cadillac (466 m), Sargent (418 m), and Pemetic (380 m) Mountains (Fig. 2 and 3). Multiple mountains were sampled because we did not find enough suitable co-located boulder and bedrock samples on individual vertical transects. We also collected six boulder samples from the northern edge of the Jordan Pond moraine in Acadia National Park. These boulders are partially submerged by water along the shore of the pond but protrude 0.5-1.5 m above modern (2015) lake level. Lastly, we sampled seven boulders taller than 1 m from the PRMC; two are from the delta top (PR-14-6 and PR-14-7), one is from the front edge of the delta on a gentle slope (PR-14-1), and four extend outward from PR-14-1 along the top a small moraine ridge less than 1-2 m high (PR-14-2 through PR-14-5) (Fig. 4). We resampled the boulder from Davis et al. (2015; PRK-17) as our PR-14-1.

At each sampling site, we targeted surfaces that exhibited characteristics of glacial erosion (e.g., striations and/or polish) to reduce the likelihood of measuring ^{10}Be inherited from prior episodes of exposure and loss of ^{10}Be by erosion since deglaciation. We also sought to avoid boulders on slopes and bedrock with signs of previous soil cover. We recorded latitude and longitude with a handheld GPS, measured topographic shielding using an inclinometer, determined surface orientation with a Brunton compass, photographed boulders, and measured sample thickness as well as boulder dimensions.

4.2 Sample Preparation and Isotopic Analysis

Samples were processed at the University of Vermont following the procedures in Corbett et al. (2016). Samples were crushed and pulverized to individual grains, underwent magnetic separation, and non-quartz minerals were removed through etching in weak acid (Kohl and Nishiizumi, 1992). Quartz purity was quantified with inductively coupled plasma optical emission spectrometry. Approximately 225 μg of ^9Be (carrier made from beryl at University of Vermont) was added to each sample. Samples were prepared in three batches each containing one process blank and one CRONUS standard (Standard N; Jull

at al., 2015). Beryllium was isolated through anion and cation column chromatography, oxidized, mixed with Nb powder, and pressed into cathodes for measurement. Ratios of $^{10}\text{Be}/^9\text{Be}$ were measured at Lawrence Livermore National Laboratory and were normalized to standard 07KNSTD3110 with an assumed $^{10}\text{Be}/^9\text{Be}$ ratio of 2850×10^{-15} (Nishiizumi et al., 2007). We corrected for background by subtracting the average of three full process blanks ($6.22 \pm 4.91 \times 10^{-16}$, 1SD) from all sample values and propagated uncertainties in quadrature. Blank-corrected $^{10}\text{Be}/^9\text{Be}$ ratios range from 1.31×10^{-13} to 5.57×10^{-14} .

4.3 Exposure Age Calculations

Sample exposure ages were calculated with the CRONUS-Earth online calculator version 2.2 (Balco, 2008) and constants version 2.2.1, using the regional northeastern North American production rate (Balco et al., 2009). The reported ages use the Lal (1991)/Stone (2000) constant production rate model and scaling scheme, though we note that other schemes yield similar ages (to within a century) since the latitude and elevation difference is minimal between the calibration sites and our study area. We assume no post-exposure erosion, no snow cover, a rock density of 2.7 g cm^{-3} , and do not account for production rate changes related to isostatic rebound because these likely would have affected the calibration sites and our study site similarly since they are close in location and age. However, we estimated how much uplift could potentially affect our ages by taking the uplift history of our field site from the ICE-6G model (Argus et al., 2014) to determine its time-averaged elevation, and using this value to recalculate exposure ages, as done by Cuzzone et al. (2016). This uplift correction would increase exposure ages by 2%, which is within our reported uncertainty (see supplementary material).

We report the arithmetic mean age for landforms, excluding outliers based on an interquartile range analysis (which considers any value more than 1.5 times outside the difference between the 25th and 75th percentile of the dataset an outlier; Boslaugh, 2012), and quantify uncertainties as the standard error of ages plus the production rate uncertainty, added in quadrature. The close agreement of the northeastern North American production rate and the more precise Arctic production rate (Young et al., 2013) suggests that our uncertainty estimate may be conservative. We use internal uncertainties when focusing on

samples within our dataset, but show external uncertainties when comparing our exposure ages to other paleoclimate records. In order to generate probabilistic estimates of the rate of ice-sheet thinning implied by the Acadia dipstick data, we followed the Monte Carlo approach of Johnson et al. (2014), fitting 1000 linear regressions through the age-elevation data while varying errors and excluding simulations with negative thinning rates. We ran four sets of Monte Carlo simulations: with the boulder and bedrock population, with just the bedrock, with just the boulders, and with averaged boulder-bedrock pairs (i.e. averaged sample pairs at a given elevation, giving eight groupings), using internal uncertainties in the first three sets and the standard deviation of sample pairs in the last set to drive the simulation.

5. Results

Background-corrected sample ^{10}Be concentrations are $3.74\text{--}8.59 \times 10^4 \text{ atoms g}^{-1}$. Most exposure ages from the Acadia National Park vertical transect (Fig. 5 and 6, Table 1) cluster between 14.5 ± 0.4 and $16.2 \pm 0.5 \text{ ka}$, but two boulders (MDI-23, MDI-25) are significantly younger (11.3 ± 0.2 and $12.4 \pm 0.2 \text{ ka}$, respectively), perhaps due to rolling, snow/soil cover, or erosion, and are excluded as outliers based on the interquartile range analysis. A paired t -test indicates that there is no significant difference between boulder and bedrock mean ages ($p = 0.24$). Similarly, an independent t -test considering boulder and bedrock samples as two distinct populations also indicates similar age distributions ($p = 0.15$). The exposure ages are indistinguishable from top to bottom, showing no significant trend with elevation whether considering the full population, the boulder population alone, the bedrock population alone, or the average of exposure ages at each elevation (Fig. 7). The mean exposure age of the entire dipstick population is $15.2 \pm 0.7 \text{ ka}$ ($n = 16$). Monte Carlo simulations indicate that the ice sheet surface likely lowered at a rate of decimeters to meters per year (Fig 7e); we note that the Monte Carlo analysis cannot constrain the highest possible thinning rates but is able to place quantitative bounds on the lowest possible rates.

The Jordan Pond moraine has an average exposure age of $13.3 \pm 0.7 \text{ ka}$ ($n = 5$), excluding one young outlier MDI-19 ($8.9 \pm 0.3 \text{ ka}$), and displays a multi-modal distribution (Fig. 6c). The average exposure age of boulders on the PRMC (Fig. 4) is $14.5 \pm 0.7 \text{ ka}$ ($n = 7$), with ages ranging from 13.9 ± 0.4

to 15.3 ± 0.4 ka (Fig. 6d). The ^{10}Be and ^{26}Al exposure ages from Davis et al. (2015; dashed grey lines in Fig. 6d) on a single boulder from the PRMC have older central tendencies than our age on the same boulder, but they have wide uncertainties that overlap with our data at 1 and 2σ , most likely because extraction methods and measurement precision have improved since the sample of Davis et al. (2015) was analyzed in 2000. The Jordan Pond moraine lies in a stratigraphic position similar to the PRMC, which suggests that both deposits could represent the same stillstand or readvance of the ice. Therefore, we performed a one-way ANOVA, which shows that the ages of the vertical transect in Acadia and the two moraines (Jordan Pond and PRMC) are all separable from each other ($p < 0.001$), indicating each represents a distinct phase of the coastal deglaciation.

6. Discussion

6.1 ^{10}Be Ages in Coastal Maine

The dipstick exposure ages indicate that the LIS surface lowered in Acadia National Park between 15.5 and 14.5 ka (Fig. 7). These ages are statistically indistinguishable from the rapid thinning identified by Davis et al. (2015) in the Katahdin area of west-central Maine between 16 and 15 ka (t test of Acadia ages versus all Katahdin ages, excluding outliers; $p=0.50$), suggesting that ice collapse was regional in scale. The Monte Carlo simulations indicate that thinning occurred in no more than a millennium, and perhaps substantially less, implying thinning rates of decimeters to meters per year ($15\text{--}680\text{ cm yr}^{-1}$ at 2σ ; Fig. 7e). This is broadly comparable to the thinning rates (average = 84 cm yr^{-1}) of the fastest flowing glaciers ($>100\text{ m yr}^{-1}$) in Greenland and the fastest thinning glaciers in Antarctica (9 m yr^{-1}) between 2002 and 2007, which appear to have been driven by dynamic thinning at marine margins (Pritchard et al., 2009).

The Monte Carlo approach we used to quantify thinning rates has been employed in several recent dipstick studies (Johnson et al., 2014; Winsor et al., 2015), but the broad range of possible thinning rates it simulates points to two factors that control the ability of dipsticks to constrain abrupt rates of ice collapse – namely, the precision of the thinning rate will vary as a function of the elevation range spanned by the dipstick relative to the variance of the ages. Thus, all else being equal, the thinning rate distribution

will be tighter for taller dipsticks; our study spans only 300 m of relief. Moreover, geologic scatter of cosmogenic ages also tends to bias the result to slower thinning rates because it allows a wider range of possible thinning rates. For instance, identical ages along a dipstick with zero uncertainty would imply instantaneous thinning, but wider error bars on the ages would also allow for simulations consistent with more gradual thinning. This limitation could potentially be overcome by averaging multiple ages per elevation to reduce noise. Given that the boulder and bedrock populations from Acadia National Park are indistinguishable – consistent with effective erosion by warm-based ice – we applied this averaging approach, and find that it indeed yields a clearer age-elevation relationship than considering the samples individually (Fig. 7d).

Our exposure age for the PRMC, 14.5 ± 0.7 ka, is slightly younger than previous estimates of ~16-15 ka based on radiocarbon constraints that assumed a 400-600 ^{14}C year reservoir correction (Kaplan, 2007). Our PRMC age is also several centuries younger than the nearby Acadia dipstick ages, indicating that the LIS margin may have stabilized at the PRMC as suggested by Thompson and Borns (1985), and perhaps represents an example of the effect of morainal banks on grounding line stability (Alley et al., 2007).

The Jordan Pond moraine age that we obtained of 13.3 ± 0.7 ka is roughly 1,000 years younger than the PRMC and the youngest Acadia dipstick ages (Fig. 6), which is difficult to explain glaciologically because Jordan Pond is outboard of the PRMC and only slightly lower than the bottom of the dipstick. Moreover, the multi-modal distribution of Jordan Pond ages suggests that the boulders do not share a simple history of exposure (Fig. 6c). Several processes may explain why these ages are younger and more scattered than others in our study. Jordan Pond initially drained through its southeast corner following deglaciation, until a new outlet was cut through till to bedrock at the southwest corner of the pond, stabilizing the lake at its current level. LiDAR imagery of the original sluiceway's elevation indicates that the former lake level was ~1 m higher than present, which would have shielded the shortest boulder we sampled (MDI-20, 13.4 ka, 0.5 m above lake level) by ~30%, but likely had little to no effect on the four taller boulder samples (1.0-1.5 m, 12.2-14.3 ka), as evidenced by no relationship between

boulder height-age relationship ($r^2 = 0.28$, $p = 0.36$). In addition, the current setting of the boulders on the edge of the moraine, several meters below its crest and partially submerged in lake water, suggests that the boulders may have initially been covered by till before wave action exhumed them. Lastly, deep snow drifts typically accumulate at the southern end of Jordan Pond in modern times during winter, suggesting that snow shielding may further contribute to the young apparent exposure ages of the boulders. Because of these issues, we consider the Jordan Pond moraine mean exposure age a minimum estimate, though we note that our oldest boulder (14.3 ± 0.3 ka; MDI-17) lines up with the youngest Acadia dipstick ages and may better represent the moraine's age.

6.2 Comparison of ^{10}Be and ^{14}C Ages in Coastal Maine

A comparison of the ^{10}Be ages with nearby radiocarbon ages from land and ocean sediments reveal similarities and differences. On the one hand, the terrestrial radiocarbon ages from the base of Sargent Mountain Pond (15.7 ± 0.7 ka, Lowell, 1980; 15.7 ± 0.4 ka, Norton et al., 2011; Fig. 5) agree well with the Acadia dipstick average exposure age (15.2 ± 0.7 ka), suggesting that vegetation became established rapidly following deglaciation. On the other hand, the ^{10}Be dipstick ages are ~1,000 years younger than the 16.1 ka isochron drawn through Mount Desert Island by Borns et al. (2004; Fig. 1) assuming a 600 ^{14}C year reservoir correction for marine radiocarbon ages. Likewise, previous radiocarbon estimates of the PRMC based on marine shells assuming a 400-600 ^{14}C year reservoir correction (Kaplan, 2007; 16.1-15.2 ka, recalibrated with MARINE13) are slightly older than our mean exposure age on the PRMC (14.5 ± 0.7 ka), and require a 1,200 ^{14}C year marine reservoir correction to be brought into better agreement (14.9-14.2 ka; see supplementary material). These exposure and marine radiocarbon age discrepancies therefore suggest that the deglacial marine reservoir effect was over 1,000 ^{14}C years in coastal Maine, which confirms earlier suggestions of large reservoir ages in this area based on co-existing marine and terrestrial organics in the Presumpscot Formation at Portland, Maine (Thompson et al., 2011) and varve chronologies of deglaciation to the west (Ridge et al., 2001). However, we note that although our exposure ages are relatively precise (average analytical uncertainty of 2.3%), there is the possibility of a systematic offset if the production rate is incorrect on the order of ± 700 years (i.e., given the 4.8%

uncertainty of the northeastern North American production rate), or if all samples have experienced similar histories of snow cover or erosion.

6.3 Paleoclimate Implications of LIS deglaciation in Coastal Maine

Our cosmogenic exposure ages in Acadia National Park (15.2 ± 0.7 ka) indicate that the LIS thinned in coastal Maine during late HS1 (Fig. 8), an association bolstered by several arguments. First, the ^{10}Be production rate is now well anchored in this region with the northeastern North American calibration (Balco et al., 2009), and latitude/elevation scaling uncertainties are minimal since our field site is spatially (≤ 400 km) and temporally (deglacial) close to most of the calibration sites. Second, extensive radiocarbon dating in coastal Maine, perhaps most notably the terrestrial basal ages from Sargent Mountain Pond (Norton et al., 2011), also place ice retreat from the coast during HS1. Lastly, given the multi-millennial duration of HS1, the only other climate interval during which coastal ice retreat plausibly occurred is the Bølling/Allerød after 14.6 ka. However, a Bølling/Allerød age would imply that our exposure ages are too old by upwards of 1,000 years and that marine radiocarbon reservoir corrections would need to be $\geq 2,000$ ^{14}C years, which together seem unlikely.

A more difficult question to answer is why did rapid collapse of the southeastern LIS happen during the cold HS1 interval? We suggest two possible scenarios. The first scenario relies on oceanic heat buildup in the Gulf of Maine due to AMOC weakening. Recent high-resolution climate model simulations that capture details of coastal bathymetry demonstrate a relationship between AMOC weakening and warming in the northwest Atlantic Shelf of several degrees Celsius. These simulations use modern boundary conditions, and thus may not apply to the deglaciation, but they highlight the potential for Atlantic temperature anomalies to be transmitted into the Gulf of Maine via the Northeast Channel (Saba et al., 2015). Furthermore, paleoclimate modeling studies suggest that AMOC weakening during Heinrich stadials produced subsurface warming in the North Atlantic (Alvarez-Solas et al., 2010, 2013; Marcott et al., 2011). Modern melting rates of outlet glaciers around Antarctica show a positive relationship of an extra 1 m per year of basal melt for each 0.1°C ocean temperature increase (Rignot and Jacobs, 2002). Thus, even slight increases in ocean temperatures in the Gulf of Maine could have led to substantially

increased sub-ice shelf melting and/or faster calving rates in the marine embayment, which could have increased ice discharge and surface drawdown further upstream. Alternatively, rather than bottom-up oceanic forcing, a second scenario calls on top-down ice sheet melt by the atmosphere. Denton et al. (2005) suggested that HS1 cooling was mainly a wintertime phenomenon associated with extensive sea ice in the North Atlantic and that summers continued to warm due to rising insolation and CO₂, which could have driven an increasingly negative ice surface mass balance over a large portion of the ice sheet surface.

Our PRMC mean age (14.5 ± 0.7 ka) corresponds within uncertainty to the beginning of the Bølling interstadial at 14.6 ka (Rasmussen et al., 2014), suggesting that the LIS may have abandoned the PRMC in response to abrupt warming at the Bølling onset associated with strengthening of the AMOC (Dorion et al., 2001; Borns et al., 2004; Fig. 8). An additional possibility is that the ice margin was furthermore forced off of the PRMC by the rapid ~15 m jump in global sea level associated with meltwater pulse 1A that also occurred at this time (14.6-14.3 ka; Deschamps et al., 2012). This sea-level ratcheting would only be plausible if meltwater pulse 1A was largely sourced from Antarctica rather than North America, since the decreased self-gravitation from the shrinking ice sheet would cause a sea-level fall in the near field (Clark et al., 2002).

Our new cosmogenic ages, in the context of previous work (Fig. 1), reveal the broad pattern of deglaciation in Maine. There was slow net retreat in coastal Maine during HS1, punctuated with multiple readvances and stillstands (e.g., PRMC, Pond Ridge moraine), followed by more rapid retreat inland during the Bølling (Borns et al., 2004; Ridge et al., 2004), which was interrupted during the Allerød at 14-13 ka with deposition of the Littleton-Bethlehem and Androscoggin moraines in western Maine/northern New Hampshire (Bromley et al., 2015; Thompson et al., 2017). A similar shift from slower HS1 to accelerated Bølling retreat, both punctuated by shorter events, is also evident in the Connecticut valley varve record to the west, and suggests that these patterns were at least regional in scale (Ridge et al., 2012).

7. Conclusions

Sixteen ^{10}Be exposure ages down mountainsides in Acadia National Park are indistinguishable within uncertainty over a 300 m range in elevation, suggesting that the LIS in coastal Maine thinned rapidly at 15.2 ± 0.7 ka during the latter part of HS1, perhaps related to regional oceanic heat buildup or summertime warming. We also provide the first direct and precise dating of the Pineo Ridge moraine complex, with seven boulders yielding a mean exposure age of 14.5 ± 0.7 ka. This later timing suggests that the ice margin stabilized at PRMC for several centuries after the rapid thinning in Acadia, and then resumed retreat coincident within uncertainty of Bølling warming or meltwater pulse 1A sea-level rise. A comparison of these exposure ages to nearby radiocarbon ages indicates that the deglacial marine reservoir effect in this area was $\sim 1,200$ ^{14}C years, considerably higher than many, though not all, previous estimates. This discrepancy suggests caution in interpreting marine radiocarbon ages of deglaciation and isostatic rebound in coastal New England and highlights the potential of cosmogenic dating to improve coastal chronologies in this region.

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Figure 1: Calibrated radiocarbon ages that constrain deglaciation throughout Maine from Dyke et al. (2003). Terrestrial samples (triangles) were calibrated with IntCal13, and marine samples (circles) were calibrated with MARINE13 using the built-in reservoir correction of 405 years (Stuiver et al., 2016). The shaded grey zone depicts the area of postglacial marine submergence. The isochrons are based on Fig. 1 in Borns (2004) with reservoir correction of 600 ^{14}C years, and show radiocarbon ages followed by calibrated ages in parentheses. The inset map shows the location of Maine within New England and the LGM margin of the LIS as thick black line. PRMC = Pineo Ridge moraine complex, indicated by dashed line. The black box shows the location of Fig. 3. Full radiocarbon data are provided in Table 1 in the supplementary material.

Figure 2: Photographs of boulder and bedrock samples from the (a) Jordan Pond moraine (MDI-20), (b) Acadia National Park vertical transect (MDI-06 (bedrock in rear) and MDI-05 (boulder in foreground)), and (c) Pineo Ridge moraine complex (PR-14-4 (boulder on left) and PR-14-5 (boulder on right)).

Figure 3: A LiDAR digital elevation model with a transparent hillshade of Mount Desert Island showing ^{10}Be exposure ages with 1σ internal uncertainties (in ka) in Acadia National Park. The calibrated basal radiocarbon age from a Sargent Mountain Pond sediment core (SMP) is shown for comparison (Norton, et al., 2011). Boulder ages are gray and bedrock ages are italicized green. Full exposure age data are provided in Table 1 and Supplementary Table 2. (U.S.G.S., 2016; LiDAR data from Gordon Longworth, pers. comm.).

Figure 4: (a) A 1/3 arcsecond LiDAR digital elevation model (USGS, 2013) with a transparent hillshade showing the location of the Pineo Ridge moraine complex in coastal Maine (black dashed line). Sample locations are represented by white dots. The black line on the Maine inset shows the location of PRMC. Black box shows the location of panel b. (b) ^{10}Be ages with 1σ internal uncertainty (in ka) of seven boulders on Pineo Ridge. Full exposure age data are provided in Table 1 and Supplementary Table 2.

Figure 5: ^{10}Be ages of boulder (circles) and bedrock (triangles) sample pairs taken at the same elevations from Acadia National Park. Calibrated basal radiocarbon age from Sargent Mountain Pond (SMP, Fig. 3) is the open diamond (Norton et al., 2011). Error bars show 1σ internal uncertainty.

Figure 6: Probability distribution functions (camel plots) of ^{10}Be ages of (a) Acadia National Park boulders, (b) Acadia National Park bedrock, (c) Jordan Pond moraine boulders, and (d) Pineo Ridge moraine complex boulders. Thin lines represent individual samples with internal uncertainties, and thick black lines show summed probability distributions. The plots exclude two Acadia boulder outliers (MDI-23, MDI-25) and one Jordan Pond outlier (MDI-19). The two grey dashed lines in panel d represent the ^{10}Be and ^{26}Al ages from Davis et al. (2015) on a Pineo Ridge moraine complex boulder (PTK-17), resampled for this study. Vertical gray bars show mean age and standard deviation.

Figure 7: ¹⁰Be ages from Acadia National Park with 1000 age-elevation regressions generated by randomly resampling boulders (grey circles) and bedrock (black triangles) ages across their uncertainty ranges (considering internal rather than external uncertainties, since production rate and scaling uncertainties are common to all samples). Two outlier boulder samples are excluded (MDI-23, MDI-25). Age-elevation regressions using (a) boulders and bedrock, (b) boulders, (c) bedrock, and (d) boulder-bedrock pair averages at each elevation. We prefer model d since averaging helps to reduce noise. Error bars show 1σ internal uncertainty in panels a-c, and standard deviation of boulder-bedrock pairs in panel d. (e) Cumulative frequency curves of the thinning rates implied by the regressions shown in panels a-d, and colored accordingly. The purple bars represent the 1 and 2 standard deviation range of the thinning rates from the boulder-bedrock pair averages.

Figure 8: Paleoclimate records from the Northern Hemisphere compared to our ^{10}Be dipstick ages. (a) 43°N June insolation curve (Laskar et al., 2004), (b) North Atlantic sediment $^{231}\text{Pa}/^{230}\text{Th}$, a proxy for Atlantic Meridional Overturning Circulation strength (McManus et al., 2004), (c) Greenland ice core $\delta^{18}\text{O}$ (NGRIP dating group, 2006) a proxy for North Atlantic temperature, (d) ^{10}Be dipstick ages of boulder-bedrock pair averages and the Pineo Ridge moraine complex age, with uncertainties combining the standard error of ages and the production rate uncertainty added in quadrature.

Table 1: ^{10}Be ages from Acadia National Park and the Pineo Ridge moraine complex.

Sample	Location	Lat (DD)	Lon (DD)	Elev (m)	Sample type	Thickness (cm)	Shielding	^{10}Be conc (10^4 atoms g^{-1}) ¹	^{10}Be unc (1σ) (10^4 atoms g^{-1})	Age (ka) ²	Internal uncertainty (1σ) (ka)	External uncertainty (1σ) (ka)
MDI-03	Pemetic Mt.	44.33703	68.24551	374	boulder	2	0.99	8.12	0.16	14.9	0.3	0.8
MDI-04	Pemetic Mt.	44.33703	68.24552	374	bedrock	2	0.99	8.03	0.26	14.7	0.5	0.9
MDI-05	Pemetic Mt.	44.33477	68.24505	374	boulder	1	1	8.59	0.18	15.5	0.3	0.8
MDI-06	Pemetic Mt.	44.33477	68.24505	374	bedrock	5	0.99	7.76	0.19	14.6	0.4	0.8
MDI-07	Pemetic Mt.	44.33211	68.24359	351	boulder	1.5	0.98	8.39	0.16	15.8	0.3	0.8
MDI-08	Pemetic Mt.	44.33211	68.24359	351	bedrock	1.5	0.99	7.86	0.15	14.7	0.3	0.8
MDI-09	Pemetic Mt.	44.32898	68.24300	285	boulder	1.5	0.99	7.82	0.15	15.5	0.3	0.8
MDI-10	Pemetic Mt.	44.32898	68.24300	285	bedrock	4	0.99	7.43	0.15	15.0	0.3	0.8
MDI-11	Pemetic Mt.	44.32645	68.24426	245	boulder	1.5	0.97	7.27	0.14	15.2	0.3	0.8
MDI-12	Pemetic Mt.	44.32645	68.24426	245	bedrock	3	0.98	7.21	0.24	15.1	0.5	0.9
MDI-13	Pemetic Mt.	44.32366	68.24363	178	boulder	3	0.95	6.84	0.14	15.7	0.3	0.8
MDI-14	Pemetic Mt.	44.32372	68.24401	170	bedrock	1.5	0.98	6.51	0.16	14.5	0.4	0.8
MDI-15	Jordan Pond	44.32426	68.25177	86	boulder	1.5	0.98	5.76	0.11	13.9	0.3	0.7
MDI-16	Jordan Pond	44.32426	68.25177	86	boulder	2	0.99	5.08	0.11	12.1	0.3	0.6
MDI-17	Jordan Pond	44.32389	68.25298	81	boulder	4	0.99	5.86	0.13	14.3	0.3	0.8
MDI-18	Jordan Pond	44.32390	68.25244	87	boulder	1	0.98	5.42	0.12	13.0	0.3	0.7
MDI-19	Jordan Pond	44.32391	68.25242	86	boulder	1.5	0.99	3.74	0.10	8.9	0.2	0.5
MDI-20	Jordan Pond	44.32388	68.25264	85	boulder	1.5	0.99	5.59	0.14	13.3	0.3	0.7
MDI-21	Cadillac Mt.	44.35244	68.22646	460	boulder	2	0.99	8.83	0.19	15.0	0.3	0.8
MDI-22	Cadillac Mt.	44.35244	68.22646	460	bedrock	2	0.99	9.50	0.27	16.2	0.5	0.9
MDI-23	Sargent Mt.	44.34106	68.27230	404	boulder	2.5	0.99	6.31	0.13	11.3	0.2	0.6
MDI-24	Sargent Mt.	44.34106	68.27230	404	bedrock	2.5	0.99	8.51	0.22	15.3	0.4	0.8
MDI-25	Penobscot Mt.	44.32363	68.26296	217	boulder	2	0.99	5.86	0.12	12.4	0.2	0.6
MDI-26	Penobscot Mt.	44.32236	68.26296	217	bedrock	5	0.99	7.04	0.14	15.3	0.3	0.8
PR-14-1	Pineo Ridge	44.67295	67.82355	74	boulder	1.5	1	6.43	0.19	15.3	0.4	0.9
PR-14-2	Pineo Ridge	44.67295	67.82355	74	boulder	4.0	1	5.90	0.12	14.3	0.3	0.7
PR-14-3	Pineo Ridge	44.67305	67.82213	72	boulder	1.0	1	5.94	0.12	14.1	0.3	0.7
PR-14-4	Pineo Ridge	44.67253	67.81900	62	boulder	1.0	1	6.21	0.12	14.9	0.3	0.8
PR-14-5	Pineo Ridge	44.67253	67.81900	62	boulder	1.0	1	6.08	0.18	14.6	0.4	0.8
PR-14-6	Pineo Ridge	44.67255	67.82676	77	boulder	2.5	1	5.97	0.16	14.3	0.4	0.8
PR-14-7	Pineo Ridge	44.67293	67.82713	77	boulder	1.5	1	5.85	0.17	13.9	0.4	0.8

¹Be measurements were normalized to standard 07KNSTD3110 (assumed ratio of 2.850×10^{-15}) and corrected for internal blanks ($6.22 \pm 4.91 \times 10^{-16}$, n=3)

²Exposure ages were calculated with the CRONUS-Earth online calculator v.2.2 and constants v.2.1 (Balco et al., 2008) using the northeastern North American production rate (Balco et al., 2009) and the Lal (1991)/Stone (2000) constant production rate and scaling scheme, assuming no erosion, no inheritance, and a density of 2.7 g cm^{-3}

9. References

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