Quaternary Science Reviews 169 (2017) 344–356

Contents lists available at ScienceDirect

Quaternary Science Reviews

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

Age of the Pineo Ridge System: Implications for behavior of the Laurentide Ice Sheet in eastern Maine, U.S.A., during the last deglaciation

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A R T I C L E I N F O

Article history:
Received 19 April 2017
Received in revised form
5 June 2017
Accepted 13 June 2017

Keywords:
Laurentide Ice Sheet
Maine
Exposure-age dating
Pineo Ridge
Heinrich Stadial 1
Glacial history
North America

A B S T R A C T

The Laurentide Ice Sheet was a major driver of global sea-level change during the last deglaciation and may have impacted both atmospheric and oceanic circulation. An understanding of past changes in the ice sheet is important for constraining its interaction with other components of the climate system. Here, we present the geologic context and chronology for ice-sheet fluctuations in eastern Maine, adjacent to the North Atlantic Ocean, thought to be a key player in the termination of the last ice age. Retreat of the Laurentide Ice Sheet through coastal Maine first produced a series of lobate grounding-line moraines, followed by deposition of the prominent Pineo Ridge System, which crosscut the earlier moraine set and which is characterized by extensive ice-contact deltas, closely spaced parallel moraines, and association with eskers. Our new ^{10}Be surface exposure ages indicate that the Pineo Ridge System, which extends for more than 100 km in eastern Maine and Atlantic Canada, dates to ~15.3 ka, ~800 years older than recent estimates. Our data are in accord with inboard minimum-limiting radiocarbon ages of terrestrial materials, which indicate deglaciation as early as 15.3 ka, as well as of marine shells that are as old as 15.0 ka. Both the deglaciation that produced the lobate moraines and the short-lived readvance that led to the Pineo Ridge System occurred during Heinrich Stadial 1. Given that faunal and isotopic evidence indicates that the ocean remained cold during deglaciation of coastal Maine, we infer that ice recession was due to rising summer air temperatures that gave way briefly to cooling to allow minor readvance. Glacial deposits north of the Pineo Ridge System display evidence of ice stagnation and downwasting, suggesting rapid ice retreat following deposition of the delta-moraine complex, coincident with the onset of the Bølling.

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1. Introduction

The Laurentide Ice Sheet (LIS) contained more than 80 m of global sea-level equivalent at the last glacial maximum (LGM) and was a major influence not only on sea level but also on Northern Hemisphere climate through its effect on atmospheric circulation, albedo, and temperature (e.g., Clark and Mix, 2002). Yet, despite its size and importance, many regions of the former ice sheet remain without detailed chronologies, leading to only generalized understanding of ice-sheet history and dynamics. The eastern margin of the LIS, spanning ~1500 km from the Gulf of Maine to Labrador, is one such location where existing chronologies have not kept pace with the growing need for precision. This is in contrast, for example, with some areas of the ice sheet, such as the Connecticut River Valley to the west, where varve sequences have led to a detailed history of ice recession (Ridge et al., 2012).

In Maine, the focus of this study, roughly 75 radiocarbon ages constrain the timing of deglaciation (Borns et al., 2004). However, nearly all afford only minimum ages for deglaciation. Radiocarbon-dated samples typically fall into one of two categories: 1) shells found in raised marine clay (Presumpscot Formation; Bloom, 1963) and 2) terrestrial organic materials found within sediment cores from lakes and bogs, some of which were inundated by the sea for an unknown amount of time before lacustrine organic sedimentation began. With the marine samples, there is the added complication of a marine reservoir effect, the size of which has been

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http://dx.doi.org/10.1016/j.quascirev.2017.06.011
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debated (Dorion et al., 2001; Borns et al., 2004; Thompson et al., 2011), but which is universally agreed to have been large (~600–1000 years). Altogether, uncertainty in existing chronologies restricts our ability to pin down the precise timing of ice-sheet fluctuations during deglaciation.

Recent studies have suggested a direct link between cold periods seen in the GISP2 ice core in Greenland and LIS stillstands/readvances in New England, at least during certain time periods (Ridge et al., 2012; Hooke et al., 2016; Thompson et al., 2017). However, with the exception of the Connecticut River Valley, there are only limited data with which to test this hypothesis. Moreover, the GISP2 isotope-inferred temperature records are now thought to pertain primarily to winter (i.e., Buizert et al., 2014), leading one to speculate on the degree to which these cold periods would affect summer melt of the LIS.

The primary goal of our project was to date a very prominent, but somewhat enigmatic, delta-moraine complex in eastern Maine in order to improve our understanding of 1) ice-sheet fluctuations in this sector and 2) the relationship between ice-sheet history and climate events recorded in Greenland ice cores and other proxies in the North Atlantic region. Known as the Pineo Ridge system (PRS; Borns, 1973), the former ice margin extends from a few tens of kilometers east of the Penobscot Lowlands eastward into Atlantic Canada (Figs. 1 and 2). The PRS cannot be traced unequivocally farther west, and its correlation (if any) with other LIS moraines,

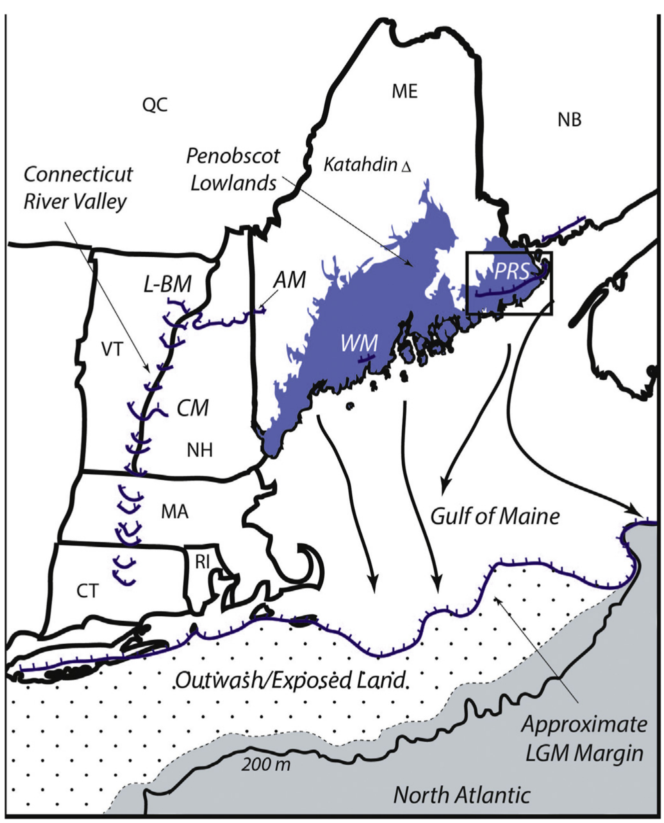

Fig. 1. Geographic context for the Pineo Ridge System, eastern Maine, showing inferred Last Glacial Maximum ice extent in the Gulf of Maine modified after Schnitker et al. (2001) and Schlee (1973). ME = Maine, NH = New Hampshire, VT = Vermont, MA = Massachusetts, CT = Connecticut, RI = Rhode Island, NB = New Brunswick, QC = Quebec, L-BM = Littleton-Bethlehem moraine, CM = Charlestown moraine, AM = Androscoggin moraine, WM = Waldoboro moraine, PRS = Pineo Ridge System. Moraines in the Connecticut River Valley are drawn after Ridge et al. (2012). Blue color shows extent of the post-glacial marine submergence in Maine, drawn after Thompson and Borns (1985); former islands are not shown. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

such as those in the Connecticut River Valley, remains uncertain. The PRS is arguably the most prominent moraine belt in the region and differs significantly in morphology from other landforms. Based on cross-cutting relationships with older moraines, the PRS is thought to represent either a local readvance (e.g., Borns, 1973; Kaplan, 1999; Borns et al., 2004) or stillstand (Smith and Hunter, 1989) during deglaciation. Borns et al. (2004) suggested an age of about 15,800 yr BP, based on minimum-limiting radiocarbon ages derived largely from shells. Davis et al. (2015) produced a single exposure age of 16,350 ± 1140 yr BP for the PRS. Recently, this boulder was redated to 15,300 ± 400 yr BP by Koester et al. (2017). The same authors added six additional dates of the same landform, together yielding an average of 14,500 ± 700 yr BP (*external error*), which they proposed indicates the age of the PRS.

2. Methods

We used a combined geomorphologic and chronologic approach toward determining the geologic significance and age of the PRS. Using satellite images, vertical air photographs, and extensive ground survey, we constructed a geomorphologic map (Figs. 3 and 4) of what is generally considered to be the 'type area' of the PRS system at the Pineo Ridge ice-contact delta and moraine, located between the towns of Cherryfield and Epping.

To obtain chronologic information for the formation of the PRS, we relied on ^{10}Be surface exposure-age dating. Recently, the application of cosmogenic surface-exposure age dating has shown promise in Maine (Bierman et al., 2015; Bromley et al., 2015; Davis et al., 2015; Koester et al., 2017). The potential advantage of this method is that one can date the actual timing of deglaciation and land exposure. One caveat, however, is that those sites covered by the marine inundation that accompanied deglaciation in Maine and that extended to a maximum limit of ~130 m elevation (less locally; Thompson et al., 1989) must either be precluded from analysis or the effect of the inundation (and potential associated exhumation and erosion) constrained.

Using maps and elevation data, we selected sites for ^{10}Be surface exposure-age dating from regions that were above former sea level, considered to be the delta surface elevation at the top of the foreslope. Elevation of this feature (77.6 m) is well-known, being part of a baseline used for a coastal geodetic survey. All samples >78 m elevation are assumed to have remained above sea level since deposition, an assumption supported by the presence of numerous subaerial meltwater channels on the delta surface and by the regional relative sea-level curve (Kelley et al., 1992), which shows continuous emergence during the early stages of deglaciation. We preferentially chose large boulders (Table 1) near the ice-contact slope of the delta in stable positions, typically on small moraines nested against the ice-contact slope. Most sites were under open, scrubby tree cover, although three samples were from fields. We chose glacially molded boulders where available and sampled them with hammer and chisel or with a drill and wedges. Elevations were obtained by handheld GPS with an averaging feature that takes advantage of repeat measurements for improved accuracy and precision. Samples from 2006 also were cross-checked with a Novalynx barometric altimeter (model Precision Absolute Manometer 355) calibrated to a local benchmark. We estimate elevation precision for all samples to be better than 5 m. Shielding was obtained using a clinometer. We do not correct for snow cover. While winter snow cover is unlikely to always be zero, Pineo Ridge is known for being a wind-swept area and total annual snowfall in many cases is less than boulder height. Boulders were visible in satellite imagery taken in early March. Moreover, there is no relationship between boulder age and height. Therefore, we conclude that snow cover has had an insignificant effect on the accumulation

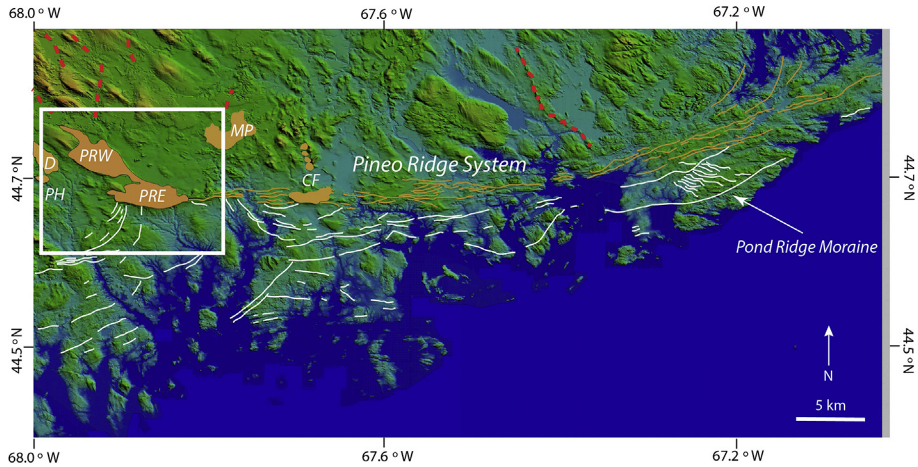

Fig. 2. Map of the PRS in eastern Maine. Orange lines and shading refer to the moraines and ice-contact deltas, respectively, of the PRS. White lines mark some of the more prominent lobate moraines, whereas red dashes mark approximate esker locations. Box shows location of Fig. 3. CF = Columbia Falls delta, MP = Montegail Pond delta, PRE = Pineo Ridge East delta, PRW = Pineo Ridge West delta, PH = Poplar Hill delta, D = Deblois sand plain. Digital elevation map from USGS (2013). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

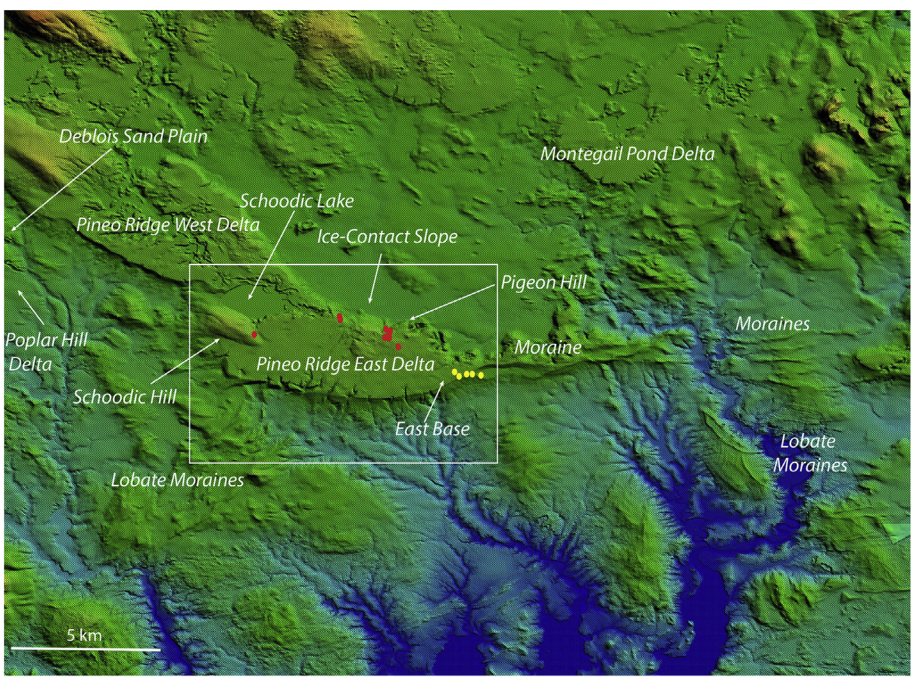

Fig. 3. The Pineo Ridge field area, with place names shown. Red dots show our samples. Yellow dots are samples from Koester et al. (2017). Box shows location of Fig. 4. Digital elevation map is from USGS (2013). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

of cosmogenic isotopes.

^{10}Be samples were prepared in the University of Maine cosmogenic isotope laboratory using standard procedures. We separated and purified quartz from the 250–500 μm fraction using surfactants and weak hydrofluoric acid (Kohl and Nishiizumi, 1992). This was followed by ion-exchange chemistry protocols to isolate beryllium (Ditchburn and Whitehead, 1994). Cathodes were processed at Lawrence Livermore National Laboratory and are normalized to 07KNSTD (Nishiizumi et al., 2007). Full procedural blanks accompanied each batch and were commonly two orders of magnitude below the sample concentrations ($^{10/9}$Be blank ratios range from 8.8236 E^{-16} to 2.72375 E^{-15}). We corrected samples by subtracting the number of ^{10}Be atoms in the corresponding process blanks. Uncertainty in the blanks was propagated in quadrature. A CRONUS standard was included in one batch (Jull et al., 2015). Final ages were calculated using the CRONUS online calculator (Balco et al., 2008), constant Lal (1991)/Stone (2000) scaling, and the regionally calibrated Northeastern North America ("New England") production rate of Balco et al. (2009). Use of this local production rate reduces uncertainty due to choice of scaling scheme and to some extent mitigates changes in cosmogenic isotope production due to postglacial uplift, as both our site and the calibration site are assumed to have generally similar uplift histories. Individual dates are cited with 1σ analytical uncertainties only. Ages of landforms are given with the standard error of the arithmetic mean, unless otherwise noted.

Throughout the text, we refer to previously published radiocarbon ages. All radiocarbon dates are quoted in the text as calendar

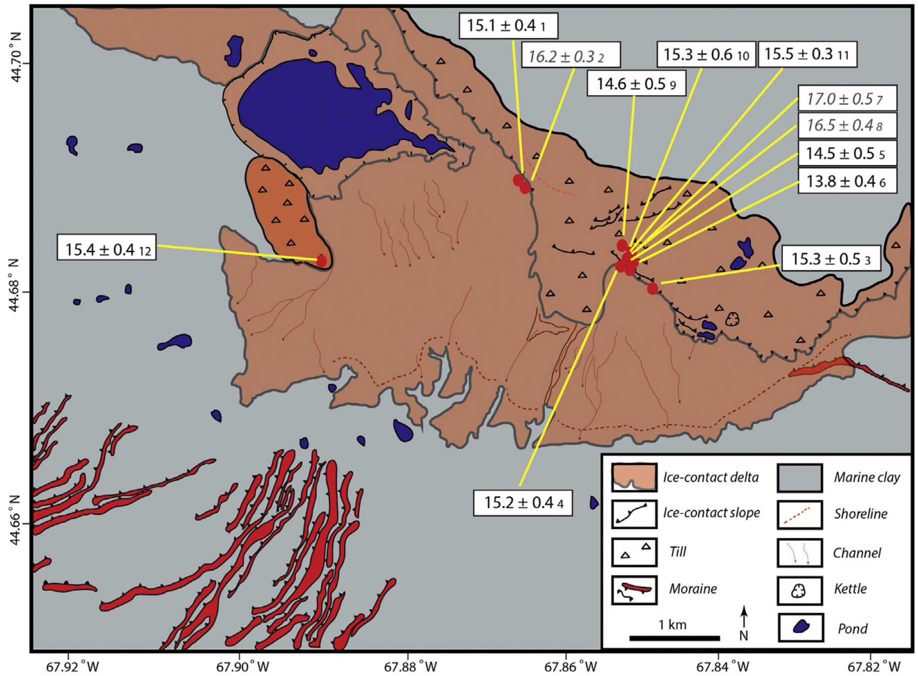

Fig. 4. Simplified geomorphologic map of the Pineo Ridge field area. Boxes show our surface exposure ages in thousands of years. Subscripts refer to Table 1. Map was drawn over Google Earth imagery, which shows more detail than other available images.

Table 1
Surface exposure age data from this study. All samples are normalized to 07KNSTD (Nishiizumi et al., 2007). Ages were calculated with the CRONUS calculator v. 2.2 (Balco et al., 2008), the Northeastern North America production rate (Balco et al., 2009), Lal (1991)/Stone (2000) scaling, a sample density of 2.7 g/cm^3, standard treatment of atmospheric pressure, and an assumption of zero erosion. Superscripts in the Sample column are keyed to Fig. 4.

Sample	Dimensions (L x W x H m)	Latitude	Longitude	Elev. (m)	Thickness (cm)	Shielding	^{10}Be (at/g)	Error (at/g)	Age (yr BP)	Error (yr BP)
PR-06-2[1]	6 × 2 x 1.5	44.69030	−67.86427	79	3.8	0.999987	6.265E+04	1.655E+03	15,120	400
PR-06-4[2]	1.5 × 1.5 x 2	44.68973	−67.86387	79	4.2	0.999978	6.698E+04	1.347E+03	16,210	330
PR-06-5[3]	4 × 3 x 2.5	44.67980	−67.84533	88	1.7	0.999999	6.516E+04	2.228E+03	15,330	530
PR-14-1[4]	2 × 2 x 1.5	44.68229	−67.84970	91	2.0	0.998754	6.451E+04	1.579E+03	15,180	370
PR-14-2[5]	2.5 × 1 x 2	44.68243	−67.84884	92	2.6	1	6.144E+04	1.926E+03	14,500	460
PR-14-3[6]	2.5 × 1.8 x 1.1	44.68210	−67.84845	91	3.5	1	5.818E+04	2.244E+03	13,840	540
PR-14-4[7]	2 × 1.2 x 0.9	44.68292	−67.84870	92	2.0	1	7.242E+04	2.287E+03	17,020	540
PR-14-5[8]	4 × 3.5 x 2	44.68275	−67.64819	91	2.1	1	7.030E+04	1.427E+03	16,540	340
PR-14-6[9]	2.2 × 2 x 1.8	44.68419	−67.84944	93	2.7	1	6.179E+04	2.183E+03	14,580	520
PR-15-2[10]	4 × 1.5 x 2	44.68366	−67.84819	92	1.5	1	6.544E+04	2.652E+03	15,310	620
PR-15-3[11]	3 × 2 x 1.5	44.68366	−67.84819	92	3.6	1	6.492E+04	1.386E+03	15,450	330
PR-15-7[12]	4 × 2 x 2.2	44.68300	−67.89165	88	6.4	1	6.321E+04	1.465E+03	15,440	360

years BP and were calibrated using the MARINE13 database (or INTCAL13, when referring to terrestrial samples) and CALIB 7.1 (Reimer et al., 2013). We applied a delta R value to the marine samples of 600 years, based on the updated reservoir calculation of Thompson et al. (2011) of ~1000 years [Reservoir calculation (~1000 years) = mean ocean value (~400 yrs) + delta R (~600 years)]. This calculation derives from a landslide deposit in southern Maine which contained both marine shells and spruce trees. We round all ages (radiocarbon and exposure age) to the nearest decade and quote them with 1σ uncertainty.

3. Results

3.1. Geomorphology

The PRS (Fig. 2) consists of a chain of interconnected deltas and tightly spaced parallel moraine segments that extends in an east-west direction across numerous river drainages and intervening highlands for ~100 km in eastern Maine and adjacent New Brunswick. Several esker segments, part of the Katahdin esker system (Shreve, 1985) terminate just inboard of the PRS and probably were the source of the meltwater and sediments that produced the deltas. In the Cherryfield region, our field area, there are several ice-contact deltas (Fig. 2), the most prominent of which is Pineo Ridge East (Figs. 3 and 4), a broad (as much as 2.9 km from back to front) ice-contact delta that extends for more than 5 km in an east-west direction. The ice-contact slope is generally steep, with more than 20 m of relief. In some locations (e.g., Pigeon Hill), several small moraines are banked up against the ice-contact slope. The delta displays well-preserved relict meltwater channels that extend across a very gently sloping surface from the ice-contact area to the foreslope. The foreslope, which rises above a swampland floored by marine clay, has 10–40 m of relief and is cut by a prominent erosional shoreline notch about 6 m below the delta top. This notch is thought to have formed during regression of the sea and is a common feature on similar ice-contact deltas associated with the

PRS in eastern Maine and New Brunswick (Thompson et al., 1989). Near the west end of the field area, a low, till-covered hill (Schoodic Hill) rises above the delta plain. Depressions near the rear of the delta, the largest of which is Schoodic Lake, likely are from ice block melt. The distance between the ice-contact slope and the delta foreslope decreases eastward to a point just east of East Base, where the delta terminates and merges into a broad moraine. For simplicity, we refer to the entire landform (Pineo Ridge East delta and the contiguous moraine) as Pineo Ridge.

Other ice-contact deltas occur in the Cherryfield region (Figs. 2 and 3), including at least two that are in recessional positions relative to Pineo Ridge. To the west of Schoodic Lake and immediately inboard of Pineo Ridge, a broad, horseshoe-shaped, ice-contact delta (Pineo Ridge West of Thompson et al., 1989) displays large areas of rocky, hummocky topography, characteristic of ice stagnation. Morphology indicates that one lobe of this delta prograded eastward into an area vacated by ice when it retreated from Pineo Ridge, thus confirming that this is a recessional landform. Another recessional ice-contact delta (Montegail Pond delta) occurs about 5 km northeast of Pineo Ridge, and a remnant of another delta forms Poplar Hill. All of these deltas, regardless of position, were graded to the same former sea level. Moreover, all display the same prominent erosional shoreline on their foreslopes. This notch also is cut into parts of the ice-contact slope of Pineo Ridge, indicating that the shoreline formed after ice recession had begun. One final large deposit of sand and gravel, the Deblois sand plain, occurs adjacent to Poplar Hill and consists of an elongate (N-S) gently sloping deposit of fluvial sediments that terminates in an abrupt drop, interpreted to be a delta foreslope (Lurvey, 1999), at ~59 m elevation.

The PRS crosscuts an older belt of moraines (Borns, 1973; Borns et al., 2004, Fig. 4). Initial deglaciation outboard of the PRS appears to have consisted of relatively steady, active retreat, which produced tens if not hundreds of low (2–10 m), relatively evenly spaced, grounding-line moraines (Smith and Hunter, 1989; Kaplan, 1999; Dorion et al., 2001). Detailed examination of these moraines in eastern Maine, including within our field area, indicates that they contain not only till, but also turbidite deposits, current-sorted sand and gravel, and marine clay (Bingham, 1981). These moraines have very rocky surfaces, covered thickly by large granite boulders, and display what we interpret as a wave-washed surface lag (Fig. 5). Moraine geometry indicates the ice margin was lobate, with deep reentrants (Fig. 2; Borns et al., 2004). The PRS crosscuts these lobate ridges, some of which can be seen projecting out from under the front of Pineo Ridge (Figs. 4 and 5). In addition, continuations of these lobate moraines are only partly buried by delta sediments and in places stand as much as 10 m above the delta surface. Even those that extend above the level of the delta are rocky and exhibit a wave-washed appearance, suggesting that sea level when the lobate moraines formed may have been higher than when the delta was deposited. Overall, the PRS differs significantly from the crosscut lobate moraine belt in landform size, orientation, ice-margin geometry, and association with deltas. Moreover, unlike the PRS, the lobate moraine belt typically is not associated with eskers.

3.2. Chronology

We obtained 12 ^{10}Be surface exposure ages from large, glacially transported boulders on Pineo Ridge (Table 1; Figs. 4 and 5). Boulders show glacial molding and rare polish. There is minimal evidence of post-depositional erosion (generally sub-centimeter-scale pits, as well as centimeter-scale, localized water erosion hollows). Nine boulders came from the ice-contact slope in the vicinity of Pigeon Hill. Four of these were set into the surface of moraines, with the rest coming from the till surface. Two additional boulders were located at the rear of the delta at the top of the ice-contact slope about a kilometer west of Pigeon Hill. The remaining sample was taken from Schoodic Hill, a till-covered knob that projects above the western end of the delta. There is no significant difference in ages among the sites.

Overall, the entire set of ages ranges from 13,840–17,020 yr BP, has a peak age of 15,280 yr BP, and shows a probability distribution with a distinct shoulder (Fig. 6). Based on comparison of calculated chi-squared statistics with theoretical values (a 'goodness of fit' test) evaluated at 95% confidence, this shoulder, comprised of three samples, likely represents a separate population of boulders ranging from 16,130–17,020 yr BP. We consider them outliers and remove them from further analysis. The remaining nine samples produce an unweighted arithmetic mean age of 14,970 $\pm$ 180 yr BP and a peak age of 15,270 yr BP, with a reduced chi-squared value of 1.39. The youngest sample skews the probability distribution and mean age, but we include it in the analysis, because it is within 2σ of the mean. Because of this skewing, we consider that the peak age, 15,270 yr BP, best represents the age of Pineo Ridge. This peak age is robust and is largely unaffected by inclusion or removal of outliers (10-yr difference; Fig. 6).

4. Discussion

4.1. Interpretation of surficial geomorphology

The geomorphology of deposits in eastern Maine represents a relative history of ice-marginal fluctuations. The oldest features on the landscape are lobate moraines that are crosscut and buried by the PRS, which is characterized by extensive sand and gravel deltas. Without chronologic control, it is impossible to determine whether the age difference between the lobate moraines and the PRS was of short or long duration. If we are correct in our assessment that even those portions of the lobate moraines that project above Pineo Ridge sediments are wave-washed, then sea level when they formed was at least ~10 m higher than that which existed during delta construction. Based on the regional sea-level curve (Kelley et al., 1992), it would have taken only a few hundred years for 10 m of relative sea-level fall during that time period. This affords a minimum estimate of the time between formation of the lobate moraines and the PRS.

All but one of the deltas in the field area are at the same elevation, despite several of them marking recessional ice positions that are younger than the PRS. This is also true for deltas to the east at Columbia Falls. This similarity in elevation suggests that the deltas may have formed rapidly and in quick succession, as there is at present no independent evidence of a stillstand in deglacial sea-level rise (Kelley et al., 1992). The speed of this formation is consistent with rapid melting and sediment delivery to the ice margin probably via the eskers that feed each delta. Overall, the volume of sediment deposited, as well as the subsequent deposition of extensive ice-stagnation landforms, suggests very quick melt of the ice-sheet margin during deglaciation from Pineo Ridge.

The very prominent shoreline notch carved in the ice-contact deltas associated with the PRS (i.e., at Pineo Ridge East and West, Poplar Hill, Montegail Pond, Columbia Falls, Pennfield) is widespread. One possibility is that this shoreline feature represents a brief period when land uplift and sea-level rise occurred at about the same rate, causing the shore to remain fixed. Alternatively, the feature could be the result of a series of large storms, although the amount of erosion in some locations is considerable. This shoreline feature formed at a time when ice already had begun to pull back from the PRS position. Given it occurs only ~5–6 m below the delta surface elevation, the shoreline probably postdates delta formation

Fig. 5. Photographs of the field area. A. Grounding-line moraine, part of the lobate moraine set that extends beneath the west end of Pineo Ridge East delta. We interpret the rocky surface as a lag deposit caused by winnowing of fine materials by ocean currents and waves; B. Similar lobate grounding-line moraine projecting obliquely from Pineo Ridge and the location of Koester et al. (2017) samples, including their KPR14-1, the largest boulder in the right of the photograph. View is to the southeast. This lobate moraine is crosscut and partially buried by the PRS and was underwater when Pineo Ridge formed; C. Sampling a glacially molded erratic near Pigeon Hill. This boulder is typical of the size of boulders sampled in our study; D. Surface of the Pineo Ridge East delta; E. Our boulder PR-14-6 in the Pigeon Hill area; F. Boulder PR-06-04, located ~1 km west of our Pigeon Hill samples.

by less than a few hundred years.

The Deblois sand plain appears to be graded to a lower sea level at ~59 m elevation (Lurvey, 1999). Our interpretation is that this outwash plain prograded as sea level fell; it is not coeval with the PRS, but rather extends back to ice-contact deposits located farther north.

4.2. Age of Pineo Ridge

Based on the peak age in our surface exposure-age data, we conclude that Pineo Ridge and, by extension, much of the PRS, was deposited at 15,270 yr BP (Fig. 6). Use of the mean age of the population (14,970 ± 180 yr BP without outliers, 15,380 ± 260 yr BP with outliers) does not alter our interpretations. Because our boulders represent some of the last rocks deposited on the ice-contact slope before ice recession, we conclude that the event that produced Pineo Ridge culminated at ~15,270 yr BP.

Our proposed age for the PRS is ~500–800 years older than a recent estimate provided by Koester et al. (2017), from seven ^{10}Be exposure ages from within our field area (Figs. 3 and 7; Table 2; they calculated ages using the same production rate and scaling as used in our study). Their dates, including a reanalysis of the sample dated in Davis et al. (2015), produce a mean age of 14,500 ± 180 yr BP (analytical uncertainty only). While the two age estimates barely overlap within 2σ error, we note that their samples come not from the PRS but from one of the lobate moraines that projects from beneath Pineo Ridge (Fig. 3). Thus, their dates should be older, not younger than ours, which were taken from the ice-contact slope of Pineo Ridge itself. One possibility is that the ages from both studies are essentially the same and that the older age of the lobate moraine cannot be resolved from that of the PRS. An alternate possibility is that the ages from the Koester et al. (2017) study were affected by the marine inundation, which covered their sample sites. This effect would not have been limited just to shielding of cosmic rays by seawater, but also to the potential erosion and exhumation of boulders from the moraine while it was within the zone of wave action. For example, we interpret the characteristic rocky surfaces of the lobate moraines, including those projecting through the delta, as being indicative of a winnowed lag. Moreover, formation of the prominent shoreline notch on the delta front may have resulted in significant erosion of the sampled moraine. It may be that the samples are dating emergence of the moraine from the

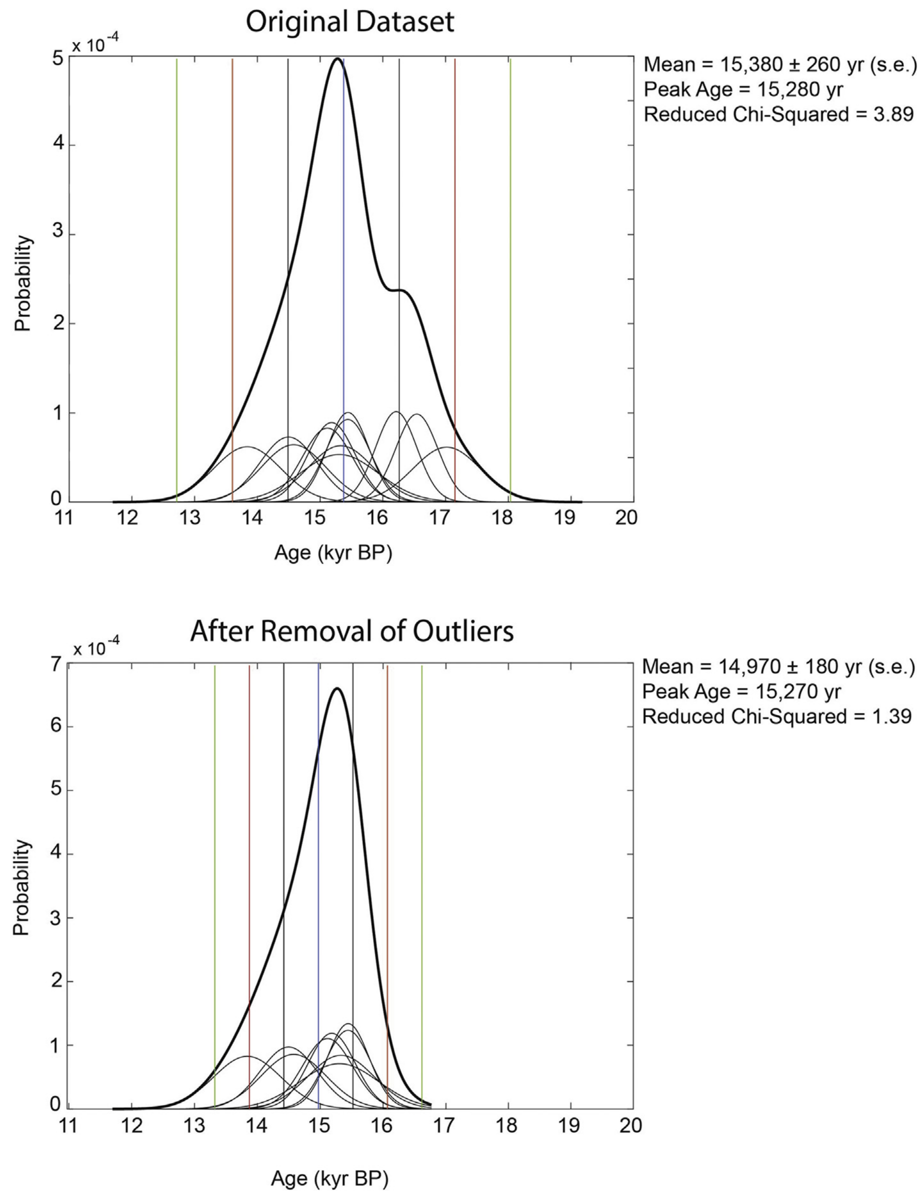

Fig. 6. Camel plots showing the probability distribution of our data. Upper plot shows original dataset and lower plot shows data after removal of three samples from consideration (see text). Vertical lines refer to the mean (blue) and 1σ (black), 2σ (red), and 3σ (green) uncertainties. Camel plots were generated using a MATLAB script modified after one developed by G. Balco (https://cosmognosis.wordpress.com/2009/07/13/matlab-code-for-camel-diagrams/). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

ocean, rather than deposition.

Overall, our exposure ages show good agreement with nearby radiocarbon dates (Fig. 7). Because deglaciation (and exposure) must have occurred before lakes formed and plants populated the landscape, surface exposure ages must be either indistinguishable from or older than inboard radiocarbon dates. Our peak age (15,270 yr BP) is nearly identical to the oldest radiocarbon age of terrestrial materials (Mountain Pond, Lurvey, 1999, Table 3) inboard of the PRS and only ~20 km away. Our mean age of 14,970 ± 180 yr BP also is indistinguishable, particularly when uncertainties in production rate are propagated (±740 yr BP). We choose the oldest radiocarbon age for comparison, because all of the inboard dates afford only minimum ages for deglaciation. However, several ages are in the ~15,000 yr BP range (Table 3, Fig. 8). Data from the Koester et al. (2017) afford a poor fit to the radiocarbon data, reinforcing the suggestion above that those dates may be too young. We also note that calculation of our exposure ages using higher production rates, such as the 'global' rate of Borchers et al. (2016), also result in a poorer fit to the radiocarbon constraints.

Although we rely on terrestrial data for the comparison between surface exposure and radiocarbon dates mentioned above, we note that our results are also consistent with the marine data, when a delta R value of 600 years is used (Thompson et al., 2011). We do not find any support or need for a larger reservoir correction to bring terrestrial and marine dates, as well as exposure ages, into alignment.

Consistent with the geomorphological evidence, radiocarbon ages suggest rapid ice recession immediately following the culmination of the event that produced the PRS at ~15,270 yr BP. In addition to the dates in the ~15,000 yr BP range directly inboard of the PRS, referred to above, the ice retreated quickly through the adjacent Penobscot Lowlands. Deglaciation had proceeded to near

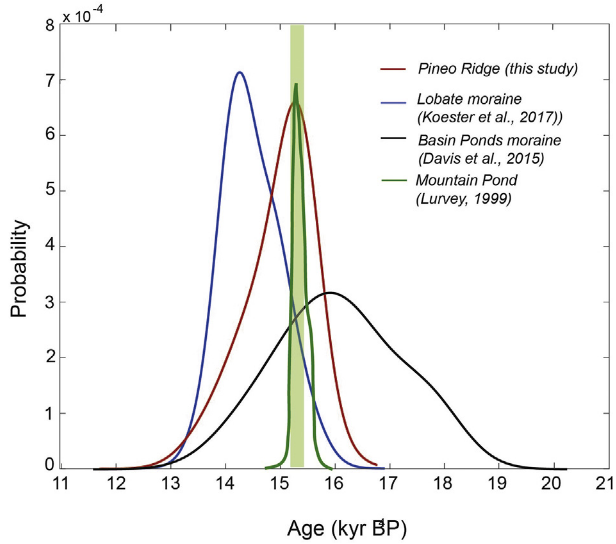

Fig. 7. Comparison of surface exposure ages for Pineo Ridge (this study), a lobate moraine projecting from the front of Pineo Ridge (Koester et al., 2017), and Basin Pond moraines near the base of Mt. Katahdin (Davis et al., 2015). Also shown is the calibrated probability age distribution for a radiocarbon date of terrestrial macrofossils from the base of Mountain Pond (Lurvey, 1999), inboard of the PRS near our field area. Note, the vertical height of this latter probability distribution is shown schematically only, as its probability is several times higher than that of the exposure ages. The green band marks the mean age and 1σ uncertainty of the radiocarbon data. Exposure ages from this study and Koester et al. (2017) should all fall to the right of the radiocarbon curve, because the PRS and lobate moraines were deposited prior to deglaciation of the Mountain Pond basin. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

the inland marine limit in the Penobscot River valley by at least 14,560 ± 300 yr BP (Mattaseunk Lake; Dorion et al., 2001) and in the Piscataquis River valley by 14,820 ± 280 yr BP (Dover-Foxcroft; Dorion et al., 2001). There is no evidence in the form of laterally extensive moraines to indicate significant pause of the ice margin at any time during this retreat. Rather, deposits north of Pineo Ridge show widespread evidence of ice stagnation, typical of rapid deglaciation (Thompson and Borns, 1985), and may mark the end of the LIS in eastern Maine.

4.3. Ice configuration during the Pineo Ridge event

The PRS differs significantly in size and orientation from moraines produced during the initial phase of deglaciation. Both the cross-cutting relationships with earlier moraines, as well as the size of the landforms, indicate not only that there was at least a minor readvance of the ice margin, but also that the ice must have maintained that more extensive position long enough to produce large deltas and moraines. Moreover, the ice margin shifted from being lobate in nature to a straighter position that cut across drainage divides (Thompson and Borns, 1985; Borns et al., 2004; Kaplan, 2007).

One puzzling feature of the PRS is that although it is easily recognized in eastern Maine and adjoining Atlantic Canada, it cannot be traced unequivocally west of the Pineo Ridge region. There is no similar moraine belt crossing central Maine. Here, we lay out two competing hypotheses for the westward extent of the PRS. The first is that the PRS corresponds to the prominent Waldoboro Moraine and associated landforms (Stone, 1899) in coastal southwestern Maine. This correlation is based largely on the observation that both the PRS and the Waldoboro Moraine system mark the termination point of eskers (Thompson and Borns, 1985); eskers are rare to absent distal to these ice positions. The Waldoboro Moraine generally has been thought to relate to an earlier phase of deglaciation, as it is one of a series of subaqueous grounding-line ridges (Smith and Hunter, 1989) with an orientation similar to those that extend beneath the west end of Pineo Ridge. This hypothesis would require that the moraines that extend beneath the Pineo Ridge delta are actually part of the PRS, and thus must be very close in age to the delta. If this correlation is correct, then the ice margin changed orientation significantly (~45°) west of Cherryfield, resulting in Pineo Ridge occupying an interlobate position. However, the Waldoboro Moraine is not dated. A date of organic-rich sediment from Kalers Pond in Waldoboro suggests a minimum age of 15,910 ± 280 yr BP (GX-23803; Voisin, 1998) for the moraine, although this date needs to be confirmed with samples of macrofossils. Terrestrial macrofossils from Ledge Pond (primarily *Dryas integrifolia* leaves and coleoptera parts, but also *Oxyria digna* and *Betula glandulosa*), located inboard of the moraine, produced an age of 17,230 ± 150 yr BP (OS-4130; Donner, 1995), significantly older than our dates for the PRS. If either of these dates is correct, it would refute the proposed PRS-Waldoboro Moraine

Table 2
Previously published exposure-age data for Basin Pond moraines, Mt. Katahdin (Davis et al., 2015) and a lobate moraine projecting from the front of Pineo Ridge (PTK- Davis et al., 2015; KPR-Koester et al., 2017 - we added a 'K' to the Koester et al. sample names to distinguish them from ours). Ages were calculated assuming a density of 2.7 g/cm^3 and negligible shielding or erosion.*Re-analysis of same boulder.

Sample	Location	Latitude	Longitude	Elev. (m)	Thick. (cm)	^{10}Be conc. (10^{-5}) at/g	Standard	^{26}Al conc. (10^{-5}) at/g	Standard	^{10}Be age	^{26}Al age
PTK-05	Basin Pond recessional	45.92154	−68.88766	713	3	1.43 ± 0.06	LLNL3000	8.75 ± 0.52	KNSTD9919	16.5 ± 0.6	17.3 ± 1.0
PTK-11	Basin Pond moraine	45.92381	−68.89444	748	5	1.26 ± 0.05	LLNL1000	9.14 ± 0.55	KNSTD9919	15.5 ± 0.7	17.8 ± 1.1
PTK-12	Basin Pond moraine	45.92381	−68.89414	749	2	1.33 ± 0.06	LLNL1000	9.22 ± 0.42	KNSTD9919	16.0 ± 0.7	17.5 ± 0.8
PTK-13	Basin Pond moraine	45.91180	−68.87747	744	4	1.56 ± 0.06	LLNL3000	8.73 ± 0.36	KNSTD9919	17.7 ± 0.6	16.9 ± 0.7
PTK-14	Basin Pond moraine	45.91201	−68.87839	750	3	2.28 ± 0.06	LLNL3000	14.89 ± 0.60	KNSTD9919	25.6 ± 0.7	28.7 ± 1.2
PTK-15	Basin Pond moraine	45.91180	−68.87870	750	4	1.29 ± 0.07	LLNL3000	7.22 ± 0.32	KNSTD9919	14.6 ± 0.7	13.9 ± 0.6
PTK-17*	Pineo lobate moraine	44.67335	−67.82246	60	0.5	0.79 ± 0.05	LLNL3000	5.63 ± 0.49	KNSTD9919	16.4 ± 1.1	19.9 ± 1.8
KPR14-1*	Pineo lobate moraine	44.67295	−67.82355	74	1.5	0.64 ± 0.02	07KNSTD	–	–	15.3 ± 0.4	–
KPR14-2	Pineo lobate moraine	44.67295	−67.82355	74	4.0	0.59 ± 0.01	07KNSTD	–	–	14.3 ± 0.3	–
KPR14-3	Pineo lobate moraine	44.67305	−67.82213	72	1.0	0.59 ± 0.1	07KNSTD	–	–	14.1 ± 0.3	–
KPR14-4	Pineo lobate moraine	44.67253	−67.81900	62	1.0	0.62 ± 0.1	07KNSTD	–	–	14.9 ± 0.3	–
KPR14-5	Pineo lobate moraine	44.67253	−67.81900	62	1.0	0.61 ± 0.2	07KNSTD	–	–	14.6 ± 0.4	–
KPR14-6	Pineo lobate moraine	44.67255	−67.82676	77	2.5	0.60 ± 0.2	07KNSTD	–	–	14.3 ± 0.3	–
KPR14-7	Pineo lobate moraine	44.67293	−67.82713	77	1.5	0.59 ± 0.2	07KNSTD	–	–	13.9 ± 0.4	–

Table 3
Pertinent radiocarbon ages relating to deglaciation of eastern Maine. All marine dates are calibrated using a delta R value of 600 years (total reservoir value of ~1000 years; Thompson et al., 2011). n.d. = no data. Superscripts after site names are keyed to Fig. 8.

Site	Lab No.	Latitude	Longitude	Material	^{14}C yr	1σ	δ^{13}C	Cal yr	1σ	Reference
Outboard of Pineo Ridge - minimum ages for deglaciation										
Dennison Point[1]	OS-2154	44.642	−67.242	*M. calcarea*	14,000	85	−0.6	15,530	210	Kaplan, 1999
Turner Brook[2]	AA-7461	44.669	−67.250	*Nucula* sp.	13,810	90	−1.8	15,310	230	Dorion et al., 2001
Turner Brook[2]	Y-2217	44.669	−67.250	seaweed	13,720	200	n.d.	15,110	440	Stuiver and Borns, 1975
Turner Brook[2]	OS-1314	44.669	−67.250	*M. calcarea*	13,650	55	−2.0	15,000	230	Dorion et al., 2001
Carrying Place Bluff[3]	OS-2075	44.813	−66.979	*N. tenuis*	13,800	80	1.4	15,290	220	Dorion et al., 2001
Carrying Place Bluff[3]	OS-7135	44.813	−66.979	*H.arctica*	13,350	50	0.6	14,390	250	Dorion et al., 2001
Carrying Place Bluff[3]	AA-8213	44.813	−66.979	*H.arctica*	13,150	150	0.2	14,070	290	Dorion et al., 2001
Hadley Pond Esker[4]	OS-2155	44.750	−67.420	*H. arctica*	12,800	50	0.6	13,650	130	Kaplan, 1999
Sprague Neck[5]	AA-7462	44.664	−67.319	*P. arctica*	13,370	90	−1.5	14,430	290	Dorion et al., 2001
Sargent Mt. Pond[6]	Beta 240351	44.334	−68.270	Org. Sed.	13,260	50	n.d.	15,940	100	Norton et al., 2011
Sargent Mt. Pond[6]	SI-4042	44.336	−68.269	Org. Sed.	13,230	360		15,830	540	Lowell, 1980
Cape Rosier[7]	Y-2214	44.313	−68.829	*Balanus* sp.	13,180	160	n.d.	14,140	330	Stuiver and Borns, 1975
Long Pond[8]	OS-3466	44.595	−68.023	*N. tenuis/ P. arctica*	12,950	120	−17.9	13,820	190	Dorion et al., 2001
Look's Cannery[9]	OS-2152	44.720	−67.310	*N. expansa*	12,900	50	−7.7	13,760	140	Kaplan, 1999
Ross Pond[10]	Beta 24945	43.923	−69.143	Cedar? twig	11,770	160	n.d.	13,600	160	Kellogg, 1988
Inboard of Pineo Ridge; Minimum ages for retreat from Pineo Ridge										
Beddington[11]	Y-2220	44.828	−68.080	Org. sed.	13,760	100	n.d.	16,630	190	Stuiver and Borns, 1975
Sand Point[12]	OS-2663	45.139	−67.129	*Nucula* sp.	13,700	70	1.9	15,070	230	Dorion et al., 2001
Patrick Lake[13]	OS-3465	44.878	−67.385	Seaweed	13,400	95	−19.3	14,470	300	Dorion et al., 2001
Lily Lake[14]	OS-2151	44.828	−67.102	*H. arctica*	13,350	50	0.2	14,390	250	Kaplan, 1999
Mark's Lake[15]	OS-3161	44.758	−67.505	*Nucula* sp.	13,300	65	1.3	14,310	260	Dorion et al., 2001
Pocomoonshine Lake[16]	OS-2661	45.125	−67.539	*P. arctica*	13,200	60	−0.1	14,090	190	Dorion et al., 2001
Pennfield Ridge[17]	GSC-882	45.097	−66.755	*P. arctica*	13,000	240	n.d.	13,850	350	Gadd, 1973
Lewis Cove[18]	OS-2659	45.034	−67.108	*N. tenuis*	12,900	50	2.3	13,760	140	Dorion et al., 2001
Mountain Pond[19]	OS-15398	44.85	−68.028	Insects/ Terrestrial Macrofossils	12,850	65	−19.2	15,320	120	Lurvey, 1999
Loon Pond[20]	SI-4953	45.039	−68.200	Terrestrial macrofossils	12,615	115	n.d.	14,950	240	Davis and Jacobson, 1985
Basswood R. Lake[21]	GSC-1067	45.255	−67.330	Gyttja	12,600	270	n.d.	14,780	480	Mott, 1975
Bear Pond[22]	OS-2093	44.853	−68.094	Terrestrial macrofossils	12,350	55	−27.2	14,330	170	Borns et al., 2004
Green Lake[23]	OS-4843	45.016	−68.063	Terrestrial macrofossils	12,200	60	−20.1	14,090	80	Borns et al., 2004
N. Beddington[24]	Y-2221	44.920	−68.100	Org. sed./peat	11,840	100	n.d.	13,660	100	Stuiver and Borns, 1975
Middle Unknown L[25]	OS-4844	45.178	−68.064	Terrestrial macrofossils	11,700	50	−25.4	13,510	40	Borns et al., 2004
Pine Ridge Pond[26]	Beta 55257	45.567	−67.100	Terrestrial vegetation	11,490	80	n.d.	13,350	80	Levesque et al., 1994
Duck Lake[27]	OS-4842	45.343	−68.052	Insects	11,000	40	−25.5	12,850	70	Borns et al., 2004
Penobscot Lowlands - minimum ages for deglaciation and likely for retreat from the PRS										
Dover-Foxcroft[28]	OS-11022	45.204	−69.181	*M. balthica*	13,550	60	−1.1	14,820	280	Dorion et al., 2001
Moulton Pond[29]	I-5639	44.628	−68.639	Org. sed.	13,510	300	n.d.	16,290	450	Davis et al., 1975
Mattaseunk Lake[30]	OS-1322	45.590	−68.378	*P. arctica*	13,450	75	−16.8	14,560	300	Dorion et al., 2001
T2R8 NWP[31]	OS-3160	45.368	−68.549	*Nucula* sp.	13,300	65	0.3	14,310	260	Dorion et al., 2001
Gould Pond[32]	AA-7463	44.993	−69.319	*P. arctica*	13,290	85	−0.5	14,300	270	Borns et al., 2004
Boyd Lake[33]	AA-9293	45.170	−68.924	*P. arctica*	13,075	90	0.4	13,990	410	Dorion et al., 2001
Monroe[34]	Y-1479	44.631	−69.010	Peat	12,220	240	n.d.	14,210	420	Stuiver and Borns, 1975
Belfast[35]	Y-1460	44.430	−69.225	Peat/Org. sed.	11,750	160	n.d.	13,590	160	Stuiver and Borns, 1975
Mud Pond[36]	Beta 43725	43.725	−68.029	*Picea* needles	11,620	100	n.d.	13,450	100	Doner, 1995
Minimum ages for Waldoboro moraine										
Ledge Pond[37]	OS-4130	44.439	−69.279	Terrestrial macrofossils	14,150	95	−28.3	17,230	160	Donner, 1995
Kalers Pond[38]	GX-23803	44.109	−69.422	Bulk organics	13,240	190	−28.0	15,910	280	Voisin, 1998
Toddy Pond[39]	OS-2662	44.543	−69.057	*E. excvatum*	13,000	60	−1.2	13,880	140	Dorion et al., 2001

correlation.

As an alternative, one possibility is that the PRS landforms are absent farther west. If a calving bay had started to develop in the Penobscot Lowlands, rapid retreat could have prevented formation of prominent moraines. In addition, whether or not this bay existed, the deeper water at the ice front in the lowlands may have precluded the formation of large ice-contact deltas, which are one of the key features that allows the PRS to be identified and traced. More detailed mapping and chronological work are needed to resolve the westward extension of the PRS.

Our new age for the PRS is also relevant to the interpretation of the extensive Pond Ridge Moraine, which lies outboard of the PRS east of the field area near Machias (Fig. 2). Kaplan (1999, 2007) noted that the Pond Ridge Moraine, similar to the PRS, showed cross-cutting relationships with earlier landforms and to at least some extent represented a reorganization of the ice margin. At Turner Brook, the oldest radiocarbon age of shells interbedded within what are interpreted as ice-proximal sediments afford an age of only 15,310 ± 230 yr BP for construction of the Pond Ridge Moraine. If this date and its interpretation are correct, then the age of the Pond Ridge Moraine is indistinguishable within dating error from that of the PRS in our field area. This similarity in ages suggests that there was very little time between deposition of the distal lobate moraine belt (to which Pond Ridge Moraine belongs) and the PRS. In general, deglaciation of coastal Maine and the formation of the lobate moraine belt appear to have occurred quickly in the millennium prior to deposition of the PRS. For example, a radiocarbon date from the base of Sargent Mountain Pond on Mt Desert Island distal to the PRS (Fig. 8), is only 15,940 ± 100 yr BP (Beta 240,351; Norton et al., 2011). This radiocarbon age is indistinguishable from an earlier date of Lowell (1980) and is in accord with nearby exposure ages that suggest rapid ice thinning at that same time (Koester et al., 2017; Hall and Lowell, unpublished data).

The size and prominence of the PRS have led to speculation that it correlates with the Basin Ponds moraines, which encircle the base of Mt. Katahdin (e.g., Davis et al., 2015; Hooke et al., 2016; Koester et al., 2017). The Basin Ponds moraines are thought to have formed during deglaciation when the mountain projected from the ice

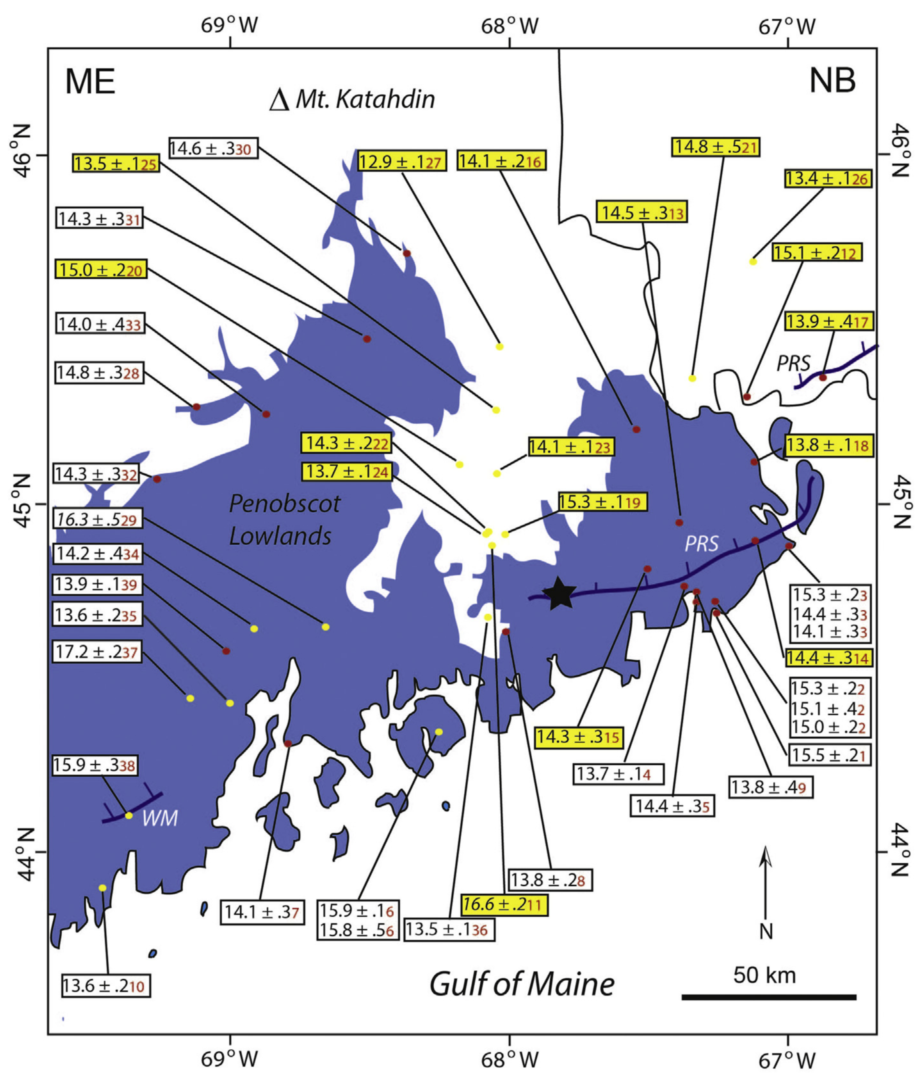

Fig. 8. Map of Maine showing pertinent radiocarbon samples. Ages have been calibrated, as outlined in the text, and are shown in thousands of years. Red subscripts refer to sample numbers on Table 3. Yellow boxes indicate samples that must afford minimum ages for the PRS based on their geographic position. All other dates are shown in white boxes and afford minimum ages for deglaciation of those particular sites. Some of these also may afford minimum ages for the PRS, but as the western extent of the PRS is not known for certain, we have been conservative in our interpretations. Yellow dots indicate dates of terrestrial materials; red dots of marine organisms (calibrated ages take into account delta-R of 600 yrs; Thompson et al., 2011). We place little reliance on the two ages in italic text, because they are bulk sediment dates obtained >40 years ago. Star marks location of Pineo Ridge field area. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

sheet as a nunatak. In favor of the correlation is the fact that both of these moraine belts are prominent landscape features that seemingly represent the only significant stillstand or readvance following the start of deglaciation. Valid correlation of these deposits would allow reconstruction of the former ice-sheet surface profile and confirm that both moraine sets relate to a regional climate event, rather than to local ice dynamics. However, examination of the exposure-age data for the Basin Ponds moraines suggests that this correlation cannot yet be drawn firmly on the basis of chronologic information. Five boulders from the most distal moraine yield ^{10}Be ages ranging from 14,560 to 25,600 yr BP and corresponding ^{26}Al ages of 13,920–28,700 yr BP (Davis et al., 2015, Table 2; Fig. 7), leading Davis et al. (2015) to suggest a mean age (with the oldest date removed) of 16,100 ± 1060 yr BP (^{10}Be) for deposition. A recessional moraine yielded a single age of 16,530 ± 640 yr BP. Thus, although the scatter in the existing ages permits the possibility that the Basin Ponds moraines and Pineo Ridge are coeval, on average, Basin Ponds dates are older, and it is equally plausible that the moraines predate the PRS.

4.4. Correlations to deposits in New England and Atlantic Canada

The PRS can be traced into southern New Brunswick to the Pennfield-Utopia and Pocologan ice-contact deltas ~40–50 km from the Maine border. From the Pocologan delta, the PRS may extend as a series of kames and moraines as far eastward as St. John (Fig. 2; Lougee, 1954; Mott, 1975; Seaman et al., 1993; Seaman, 2004). Based on radiocarbon dates of shells in nearby deposits, Rampton et al. (1984) suggested that ice was at this position at ~ 14,000 yr BP, somewhat younger than our data indicate for Pineo Ridge. However, the exact relationship of the dated material to the former ice position is uncertain.

Correlation of the PRS to moraines in western New England remains elusive. Although due to the New England varve chronology moraines in the Connecticut River valley are well-dated (i.e., Ridge et al., 2012), only a few stand out in terms of prominence or continuity and none can be traced far into Maine. Moreover, the frequency of moraine deposition in the river valley makes it difficult to distinguish which one (if any) correlates to the PRS. The North Charlestown moraines (Fig. 1), dated to ~14,500–14,600 yr BP by the varve chronology (Ridge et al., 2012), are thought to represent a short readvance or stillstand similar to the PRS, but there are also several more distal moraines dating closer to ~15,270 yr BP, the age of Pineo Ridge. Farther north, the Littleton-Bethlehem moraines (~14,000 yr BP; Thompson et al., 1999) represent a significant stillstand. Geomorphologically, this feature resembles the PRS in that it is laterally extensive. However, based on our age data, it appears that the PRS is older than the Littleton-Bethlehem moraines.

4.5. Relationship of ice dynamics to North Atlantic climate

Previous researchers have related the formation of moraines in eastern Maine to cold periods recorded in the Greenland ice cores (e.g., Hooke et al., 2016). In particular, the PRS and Pond Ridge moraines have been correlated to the Oldest Dryas period, also known as Heinrich Stadial 1 (HS-1; 14,700–17,500 yr BP), which was a time of cold temperatures, at least in winter, in parts of the North Atlantic region (Alley, 2000). Our data place the formation of the PRS within the second half of HS-1 at ~15,270 yr BP. At first glance, this age lends support for the proposed correlation between Maine ice-sheet history and cold periods in the Greenland ice-core record. However, exposure-age and radiocarbon data summarized above suggest that deglaciation to the position of the PRS also occurred during HS-1. Thus, HS-1 in eastern Maine appears to have been a time of rapid ice retreat followed by a short-lived stillstand/readvance. If this pattern of ice behavior were due to changes in air temperature, then the inference would be that climate in Maine warmed during this recession, then cooled briefly to produce the PRS. Such a reconstruction is unlike the isotope-inferred temperature record from GISP2, which shows extremely cold conditions throughout HS-1 (Alley, 2000). This discrepancy could result from glaciers responding primarily to summer temperature change (Oerlemans, 2005), whereas the periods of extreme cold in the Greenland ice-core records (such as HS-1) are thought to reflect largely winter (Denton et al., 2005; Buizert et al., 2014). Recent analysis of meltwater discharge events from the European ice sheet via the Channel River also suggests rapid melting and ice-sheet retreat during HS-1 (Toucanne et al., 2015) and provides support for this hypothesis of rising summer air temperatures. An alternate hypothesis is that the rapid deglaciation in coastal Maine during HS-1 reflects marine recession due to hypothesized warmer sea-surface temperatures in the Gulf of Maine (Koester et al., 2017). While possible, we do not favor this hypothesis, because faunal and isotopic data indicate retreat took place under arctic marine conditions with consistently cold water temperatures (Dorion et al., 2001).

Regardless of the reasons for ice retreat during HS-1, the link between LIS recession and rising air temperatures seems clear during deglaciation from the PRS. This deglaciation corresponds closely in time to the start of the Bølling warm period, the largest warming recorded in many North Atlantic region records during Termination I. Radiocarbon dates, as well as the lack of coherent moraines and the presence of widespread ice-stagnation topography north of the PRS, all suggest rapid ice recession and downwasting.

5. Conclusions

- The PRS forms an extensive belt of ice-contact deltas and tightly spaced moraines that extends across eastern Maine into Atlantic Canada and crosscuts a series of lobate moraines that formed during deglaciation of the present coast. Extension of the PRS to the west remains equivocal.
- Our new ^{10}Be surface exposure -age data suggest that the PRS was formed at 15,270 yr BP (peak age; mean age of 14,970 ± 180 yr BP without outliers, 15,380 ± 260 yr BP with outliers). This age is older than that proposed in Koester et al. (2017) from boulders on a cross-cut lobate moraine that projects from the front of Pineo Ridge and that is outboard of (older than) our samples. One possibility is that the boulders in that study were affected by marine inundation and erosion.
- The timing of initial deglaciation to the PRS remains uncertain. However, most data suggest that the period between formation of the lobate moraines during coastal deglaciation and the formation of the PRS was of short duration.
- Regardless of the exact time of initial deglaciation, existing data suggest that both ice recession through coastal Maine and the readvance/stillstand that produced the PRS occurred during HS-1. One hypothesis is that this recession was due to rising summer air temperatures, which gave way to brief cooling near the end of HS-1, leading to the formation of the PRS. This reconstruction is not in accord with the GISP2 isotope-inferred temperature record, which shows extremely cold conditions throughout HS-1. This discrepancy, should it bear up under additional data collection, could be due to the glacial record reflecting summer temperatures, whereas the Greenland record was influenced strongly by winter.
- Recession from the PRS corresponds closely in time with the start of the Bølling warm period, suggesting that the two events

may be linked. Both the geomorphology of the deposits north of the PRS (suggestive of extensive ice stagnation) and the minimum-limiting radiocarbon ages inboard of the PRS (which overlap with our exposure ages from the ice-contact slope of Pineo Ridge) suggest ice recession was very rapid.

Acknowledgments

We thank Wyman's of Maine and Cherryfield Foods for access to their land for field work and their wider support for our glacial geologic research and Maine's Ice Age Trail: Down East, Map and Guide. We also thank K. Overturf for assistance in the field and laboratory and two reviewers whose comments improved this manuscript. This research was supported by grants from the National Science Foundation (ATM-0511226) and the Comer Family Foundation (CP111).

References

Alley, R.B., 2000. Ice core evidence of abrupt climate change. Proc. Natl. Acad. Sci. 97, 1331–1334.
Balco, G., Stone, J.O., Lifton, N.A., Dunai, T.J., 2008. A complete and easily accessible means of calculating surface exposure ages or erosion rates from ^{10}Be and ^{26}Al measurements. Quat. Geochronol. 2, 174–195.
Balco, G., Briner, J., Finkel, R.C., Rayburn, J.A., Ridge, J.C., Schaefer, J., 2009. Regional beryllium-10 production rate calibration for northeastern North America. Quat. Geochronol. 4, 93–107.
Bierman, P.R., Davis, P.T., Corbett, L.B., Lifton, N.A., Finkel, R.C., 2015. Cold-based Laurentide ice covered New England's highest summits during the last glacial maximum. Geology 43, 1059–1062.
Bingham, M.P., 1981. The Structure and Origin of Washboard Moraines and Related Glacial Marine Sediment in Southeastern Coastal Maine. M.S. Thesis. University of Maine, Orono, Maine, 103pp.
Bloom, A.L., 1963. Late Pleistocene fluctuations of sea level and post-glacial crustal rebound in coastal Maine. Am. J. Sci. 261, 862–879.
Borchers, B., Marrero, S., Balco, G., Caffee, M., Goehring, B., Lifton, N., Nishiizumi, K., Phillips, F., Schaefer, J., Stone, J., 2016. Geological calibration of spallation production rates in the CRONUS-earth project. Quat. Geochronol. 31, 188–198.
Borns Jr., H.W., 1973. Late-Wisconsin fluctuations of the Laurentide ice sheet in southern and eastern New England. In: Black, R.F., et al. (Eds.), The Wisconsin Stage, vol. 136. Geological Society of America Memoir, pp. 37–45.
Borns Jr., H.W., Doner, L.A., Dorion, C.C., Jacobson Jr., G.L., Kaplan, M.R., Kreutz, K.J., Lowell, T.V., Thompson, W.B., Weddle, T.K., 2004. The deglaciation of Maine, U.S.A. In: Ehlers, J., Gibbard, P.L. (Eds.), Quaternary Glaciations - Extent and Chronology Part II. Elsevier, pp. 89–109.
Bromley, G.R.M., Hall, B.L., Thompson, W.B., Kaplan, M.R., Garcia, J.L., Schaefer, J.M., 2015. Late-glacial fluctuations of the Laurentide ice sheet in the White Mountains of Maine and New Hampshire, U.S.A. Quat. Res. 83, 522–530.
Buizert, C., Gkinis, V., Severinghaus, J.P., He, F., Lecavalier, B.S., Kindler, P., Leuenberger, M., Carlson, A.E., Vinther, B., Masson-Delmotte, V., White, J.W.C., Liu, Z., Otto-Bliesner, B., Brook, E.J., 2014. Greenland temperature response to climate forcing during the last deglaciation. Science 345, 1177–1180.
Clark, P.U., Mix, A.C., 2002. Ice sheets and sea level of the last glacial maximum. Quat. Sci. Rev. 21, 1–7.
Davis, P.T., Bierman, P.R., Corbett, L., Finkel, R.C., 2015. Cosmogenic exposure-age evidence for rapid Laurentide deglaciation of the Katahdin area, west-central Maine, USA, 16 to 15 ka. Quat. Sci. Rev. 116, 95–105.
Davis, R.B., Jacobson Jr., G.L., 1985. Late glacial and early Holocene landscapes in northern New England and adjacent areas of Canada. Quat. Res. 23, 341–368.
Davis, R.B., Bradstreet, J.E., Stuckenrath, R., Borns Jr., H.W., 1975. Vegetation and associated environments during the past 14,000 years near Moulton Pond, Maine. Quat. Res. 5, 435–466.
Denton, G.H., Alley, R.B., Comer, G.C., Broecker, W.S., 2005. The role of seasonality in abrupt climate change. Quat. Sci. Rev. 24, 1159–1182.
Ditchburn, R.G., Whitehead, N.E., 1994. The separation of ^{10}Be from silicates. In: Hancock, G., Wallbrink, P. (Eds.), Third Workshop of the South Pacific Radioactivity Association. Australian National University, Canberra, pp. 4–7.
Doner, L.A., 1995. Late-pleistocene Environments in Maine and the Younger Dryas Dillemma. M.S. Thesis. University of Maine, Orono, Maine, 68pp.
Donner, J.E., 1995. The Effects of Islands on the Recession of the Late Wisconsinan Ice Sheet Margin in the De Geer Sea, Maine. M.S. Thesis. University of Maine, Orono, Maine, 90pp.
Dorion, C.C., Balco, G.A., Kaplan, M.R., Kreutz, K.J., Wright, J.D., Borns Jr., H.W., 2001. Stratigraphy, paleoceanography, chronology, and environment during deglaciation of eastern Maine. In: Weddle, T.K., Retelle, M.J. (Eds.), Deglacial History and Relative Sea-level Changes, Northern New England and Adjacent Canada, vol. 351. Geological Society of America Special Paper, pp. 215–242.
Gadd, N.R., 1973. Quaternary geology of southwest New Brunswick, with particular reference to Fredricton area. Geol. Surv. Can. Pap. 71–34, 31.
Hooke, R.LeB., Hanson, P.R., Belknap, D.F., Kelley, A.R., 2016. Late-glacial and holocene history of the Penobscot River in the Penobscot Lowland, Maine. Holocene. http://dx.doi.org/10.1177/0959683616670474.
Jull, A.J.T., Scott, E.M., Bierman, P., 2015. The CRONUS-Earth inter-comparison for cosmogenic isotope analysis. Quat. Geochronol. 26, 3–10.
Kaplan, M.R., 1999. Retreat of a tidewater margin of the Laurentide ice sheet in eastern coastal Maine between ca. 14,000 and 13,000 ^{14}C yr BP. Geol. Soc. Am. Bull. 111, 620–632.
Kaplan, M.R., 2007. Major ice sheet response in eastern New England to a cold North Atlantic region, ca. 16-15 cal ka BP. Quat. Res. 68, 280–283.
Kelley, J.T., Dickson, S.M., Belknap, D.F., Stuckenrath Jr., R., 1992. Sea-level change and late Quaternary sediment accumulation on the southern Maine interior continental shelf. In: Fletcher, C., Wehmiller, J. (Eds.), Quaternary Coasts of the United States: Marine and Lacustrine Systems, vol. 48. SEPM Special Publication, pp. 23–34.
Kellogg, D.C., 1988. Problems in the use of sea-level data for archaeological reconstructions. In: Nicholas, G.P. (Ed.), Holocene Human Ecology in Northeast North America. Plenum Press, New York, pp. 81–104.
Koester, A.J., Shakun, J.D., Bierman, P.R., Davis, P.T., Corbett, L.B., Braun, D., Zimmerman, S.R., 2017. Rapid thinning of the Laurentide ice sheet in coastal Maine, USA, during late Heinrich Stadial 1. Quat. Sci. Rev. 163, 189–192.
Kohl, C., Nishiizumi, K., 1992. Chemical isolation of quartz for measurement of in situ-produced cosmogenic nuclides. Geochim. Cosmochim. Acta 56, 3586–3587.
Lal, D., 1991. Cosmic ray labeling of erosion surfaces: in situ nuclide production rates and erosion models. Earth Planet. Sci. Lett. 104, 424–439.
Levesque, A.J., Cwynar, L.C., Walker, I.R., 1994. A multiproxy investigation of late-glacial climate and vegetation change at Pine Ridge Pond, southwest New Brunswick, Canada. Quat. Res. 42, 316–327.
Lougee, R.J., 1954. The role of upwarping in the post-glacial history of Canada, Part II: the Maritime region and the St. John valley. Rev. Can. Géogr. VIII, 3–52.
Lowell, T.V., 1980. Late Wisconsin Ice Extent in Maine: Evidence from Mt. Desert Island and the St. John River Area. M.S. Thesis. University of Maine, Orono, Maine, 180pp.
Lurvey, L.K., 1999. An Investigation of a Glacial Landform Transition in Southwestern Maine. M.S. Thesis. University of Maine, Orono, Maine, 79pp.
Mott, R.J., 1975. Palynological studies of lake sediment profiles from south-western New Brunswick. Can. J. Earth Sci. 12, 273–188.
Nishiizumi, K., Imamura, M., Caffee, M.W., Southon, J.R., Finkel, R.C., McAninch, J., 2007. Absolute calibration of ^{10}Be AMS standards. Nucl. Instrum. Methods Phys. Res. Sect. B 258, 403–413.
Norton, S.A., Perry, R.H., Saros, J., Jacobson Jr., G.L., Fernandez, I.J., Kopacek, J., Wilson, T.A., San Clements, M., 2011. The controls on phosphorus availability in a boreal lake ecosystem since deglaciation. J. Paleolimnol. 46, 107–122.
Oerlemans, J., 2005. Extracting a climate signal from 169 glacier records. Science 308, 675–677.
Rampton, V.N., Gauthier, R.C., Thibault, J., Seaman, A.A., 1984. Quaternary geology of New Brunswick. Geol. Surv. Can. Mem. 416, 77.
Reimer, P., et al., 2013. IntCal13 and Marine13 radiocarbon age calibration curves 0-50,000 years cal BP. Radiocarbon 55, 1869–1887.
Ridge, J.C., Balco, G., Bayless, R.L., Beck, C.C., Carter, L.B., Dean, J.L., Voytek, E.B., Wei, J.H., 2012. The new North American varve chronology: a precise record of southeastern Laurentide Ice Sheet deglaciation and climate, 18.2-12.5 kyr BP, and correlations with Greenland ice core records. Am. J. Sci. 312, 685–722.
Schnitker, D., Belknap, D.F., Bacchus, T.S., Friez, J.K., Lusardi, B.A., Popek, D.M., 2001. Deglaciation of the Gulf of Maine. In: Weddle, T.K., Retelle, M.J. (Eds.), Deglacial History and Relative Sea-level Changes, Northern New England and Adjacent Canada, vol. 351. Geological Society of America Special Paper, pp. 9–34.
Schlee, J., 1973. Atlantic continental shelf and slope of the United States - sediment texture of the northeastern part. U.S. Geol. Surv. Prof. Pap. 529L, 64.
Seaman, A.A., 2004. Late pleistocene history of New Brunswick, Canada. In: Ehlers, J., Gibbard, P.L. (Eds.), Quaternary Glaciations - Extent and Chronology Part II. Elsevier, pp. 151–167.
Seaman, A.A., Broster, B.E., Cwynar, L.C., Lamothe, M., Miller, R.F., Thibault, J.J., 1993. Field guide to the quaternary geology of south-western New Brunswick. New Brunswick department of Natural resources and energy. Mineral Resour. Open File Rep. 93–1, 102.
Shreve, R., 1985. Esker characteristics in terms of glacier physics, Katahdin esker system, Maine. Geol. Soc. Am. Bull. 96, 639–646.
Smith, G.W., Hunter, L.E., 1989. Late Wisconsinan deglaciation of coastal Maine. In: Tucker, R.D., Marvinney, R.G. (Eds.), Studies in Maine Geology - Quaternary Geology, vol. 6, pp. 13–32.
Stone, G.H., 1899. The glacial gravels of Maine. U.S. Geol. Surv. Monogr. 34, 499.
Stone, J.O., 2000. Air pressure and cosmogenic isotope production. J. Geophys. Res. B 105, 753–759.
Stuiver, M., Borns Jr., H.W., 1975. Late Quaternary marine invasion in Maine: its chronology and associated crustal movement. Geol. Soc. Am. Bull. 86, 99–104.
Thompson, W.B., Borns Jr., H.W., 1985. Surficial geological map of Maine. Maine Geol. Surv. 1, 500,000 scale map.
Thompson, W.B., Crossen, K.J., Borns Jr., H.W., Andersen, B.G., 1989. Glacimarine deltas of Maine and their relation to late Pleistocene-Holocene crustal movements. In: Anderson, W.A., Borns Jr., H.W. (Eds.), Neotectonics of Maine, Maine Geological Survey Bulletin, vol. 40, pp. 43–67.
Thompson, W., Fowler, B., Dorion, C., 1999. Deglaciation of the northwestern White Mountains, New Hampshire. Geogr. Physique Quat. 53, 59–77.

Thompson, W.B., Griggs, C.B., Miller, N.G., Nelson, R.E., Weddle, T.K., Kilian, T.M., 2011. Associated terrestrial and marine fossils in the late-glacial Presumpscot Formation, southern Maine, USA, and the marine reservoir effect on radiocarbon ages. Quat. Res. 75, 552–565.
Thompson, W.B., Dorion, C.C., Ridge, J.C., Balco, G., Fowler, B.K., Svendsen, K.M., 2017. Deglaciation and late-glacial climate change in the White Mountains, New Hampshire, U.S.A. Quat. Res. 87, 96–120.
Toucanne, S., Soulet, G., Freslon, N., Silva Jacinto, R., Dennielou, B., Zaragosi, S., Eynaud, F., Bourillet, J.-F., Bayon, G., 2015. Millennial-scale fluctuations of the European ice sheet at the end of the last glacial, and their potential impact on climate. Quat. Sci. Rev. 123, 113–133.
USGS, 2013. USGS NED n45w068 1/3 Arc-second 2013 1x1° ArcGrid. http://ned.usgs.gov/.
Voisin, D.T., 1998. Late Quaternary Post-glacial History of Kaler's Pond, Waldoboro, Maine. B.S. Honors Thesis. Bates College, Lewiston, Maine, 85pp.