



Cosmogenic exposure age evidence for rapid Laurentide deglaciation of the Katahdin area, west-central Maine, USA, 16 to 15 ka



P. Thompson Davis^a, Paul R. Bierman^b, Lee B. Corbett^{b,*}, Robert C. Finkel^{c,d}

^a Department of Natural and Applied Sciences, Bentley University, Waltham, MA 02454-4705, USA

^b Geology Department and Rubenstein School of the Environment and Natural Resources, University of Vermont, Burlington, VT 05405-1758, USA

^c Department of Earth and Planetary Sciences, University of California, Berkeley, CA 95064, USA

^d Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory, Livermore, CA 94550, USA

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ABSTRACT

Katahdin, the highest peak in Maine and part of the second highest mountain range in New England, provides an opportunity to assess the timing and style of continental ice sheet surface lowering during deglaciation. We collected 14 samples from boulders on the adjacent Basin Ponds moraine, from bedrock and boulders on the upper part of the mountain, and from boulders in the surrounding area to estimate the age at which they were exposed by deglaciation of the Laurentide Ice Sheet. Measurements of *in situ* produced ¹⁰Be, which are consistent with measurements of ²⁶Al, indicate that the Katahdin edifice became exposed from under ice by 15.3 ± 2.1 ka ($n = 6$), an age indistinguishable from the adjacent Basin Ponds moraine (16.1 ± 1.2 ka, $n = 5$). A boulder in the lowlands several km south of the moraine dates to 14.5 ± 0.8 ka, and a boulder deposited at Pineo Ridge, about 170 km SE of Katahdin, dates to 17.5 ± 1.1 ka. These data show that samples collected over an elevation range of 1.6 km and a distance of >170 km all have exposure ages that are indistinguishable within uncertainties. Together these data suggest that the Laurentide Ice Sheet surface dropped rapidly and the ice sheet margin retreated quickly across Maine between about 16 and 15 ka, perhaps influenced by calving of the marine-based ice sheet in the St. Lawrence Lowlands to the north and the Penobscot basin to the south.

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1. Introduction

Until recently, the deglacial chronology of the Laurentide Ice Sheet in New England was primarily constrained by minimum-limiting ¹⁴C ages on organic material deposited in ponds and bogs as well as shells in marine sediments deposited following deglaciation (e.g., Davis and Jacobson, 1985; Thompson et al., 1996, 1999; Dorion et al., 2001). However, an unknown lag time between deglaciation and the deposition of the first datable organic material (Davis and Davis, 1980), the uncertainty of reservoir corrections for marine samples (Kaplan, 1999; Thompson et al., 2011), and the paucity of ¹⁴C datable samples (Balco and Schaefer, 2006) means that in many places the chronologic framework is insufficient to test competing hypotheses for the timing and style of deglaciation. Flint (1929), for example, proposed that large parts of the Laurentide Ice Sheet melted in place, an idea that was adopted by Goldthwait (1938, 1970), and Goldthwait and Mickelson (1982).

Then, Antevs (1939) and Lougee (1940) countered that the ice sheet retreated with an active ice margin, a concept later adopted by Koteff and Pessl (1981). Recent work by Ridge et al. (1999, 2012) and Ridge (2004) used ¹⁴C dating to produce a numerical chronology for the glacial Lake Hitchcock varve record of Antevs (1922), which has been used in conjunction with other data to draw glacial retreat isochrones across western New England (Ridge et al., 2012).

Since the first cosmogenic dating of moraine deposits in 1990 (³⁶Cl, Phillips et al., 1990), cosmic-ray produced isotopes (e.g., ³He, ¹⁰Be, and ²⁶Al) have been used extensively to estimate exposure ages of glacially-related deposits around the world (e.g., Gosse et al., 1995; Bierman et al., 1999; Marsella et al., 2000; Gosse and Phillips, 2001; Briner et al., 2005; Davis et al., 2006; Schaefer et al., 2006, 2009; Kelly et al., 2008; Ivy-Ochs et al., 2009; Owen, 2009). In New England, however, the application of cosmogenic nuclide exposure dating has been limited. Clark et al. (1995) reported ¹⁰Be concentrations of samples collected just inside the Laurentide margin in New Jersey and used these data along with independent ¹⁴C age control to constrain nuclide production rates since 21 ka. Balco et al. (2002) dated coastal moraines in

* Corresponding author. Tel.: +1 802 380 2344.

E-mail address: Ashley.Corbett@uvm.edu (L.B. Corbett).

Massachusetts and determined that the Laurentide Ice Sheet reached its maximum extent there at about 23 ka (recalculated to 27 ka in reference to modern AMS standards). Balco and Schaefer (2006) used ^{10}Be to date boulders on moraines in southern Connecticut with high precision and tied those ages to the New England varve chronology (Ridge et al., 1999). Then, Balco et al. (2009) used well-dated sites in New England and the Canadian Arctic to calibrate ^{10}Be production rates for northeastern North America during deglaciation.

Here, we present measurements of *in situ*-produced cosmogenic ^{10}Be and ^{26}Al for 14 samples collected on Katahdin, from the lowland south of Katahdin, and from a ^{14}C -dated moraine-marine delta complex at Pineo Ridge about 170 km to the southeast, close to the present-day Maine coast (Fig. 1). We consider our data in light of the existing ages generated using other chronometers (Kaplan, 1999, 2007; Dorion et al., 2001; Borns et al., 2004) and the surface processes that can affect cosmogenic exposure ages (Davis et al., 1999; Colgan et al., 2002; Heyman et al., 2011). We use the cosmogenic nuclide data to test several long-standing hypotheses including: continental ice covered summit areas during the late Wisconsinan; continental ice surfaces lowered and cirque glaciers did not reform during deglaciation; and the Basin Ponds moraine and other moraines downslope were formed by a stillstand or re-advance of continental ice in the lowland surrounding Katahdin, and not by cirque glaciers.

2. Background, study site, and previous work

Katahdin (meaning “greatest mountain” in Penobscot) is the highest peak in Maine (1605 m), with a local relief of about 1450 m,

only surpassed in height in the northeastern United States by the Presidential Range in New Hampshire (Figs. 1 and 2). The mountain is composed of a large Devonian pluton (Katahdin granite) that intrudes lower and middle Paleozoic sedimentary and volcanic rocks, which underlie the surrounding lowlands (Caldwell, 1972; Hon, 1980; Rankin and Caldwell, 2010). Most workers agree that Katahdin was covered by ice at some time in the Pleistocene. Erratics found by Tarr (1900) and Antevs (1932) near the summit of Katahdin, and by Caldwell (1972) on other mountains in the Katahdin region, support this view. Non-weathered erratic cobbles and weakly developed soil profiles on the summit areas, as well as modeled ice profiles, suggested to Davis (1976, 1989) that the summit areas of Katahdin were glaciated during the late Wisconsinan. A general model for deglaciation in northern New England calls for thinning of the Laurentide Ice Sheet that exposed the higher mountains as nunataks (Borns, 1985). This concept is incorporated in a numerical model for the deglaciation of northern New England and adjacent maritime Canada by Hughes et al. (1985).

Katahdin is unusual in New England because it has several distinct cirques. The three largest cirques lie on the east side of Katahdin and have headwall heights that range between about 345 and 720 m. Although the three great east-side cirques have flat to concave floors and steep headwalls composed largely of bedrock (Figs. 1 and 2), postglacial rockfall and avalanche debris mask the lower slopes of the cirque headwalls and sidewalls. These cirques are remarkably steep, especially when compared with other cirque-like features in northeastern United States, believed by some (Wagner, 1970; Craft, 1979; Bradley, 1981; Fowler, 2010), but not others (Borns and Calkin, 1977; Gerath and Fowler, 1982; Fowler,

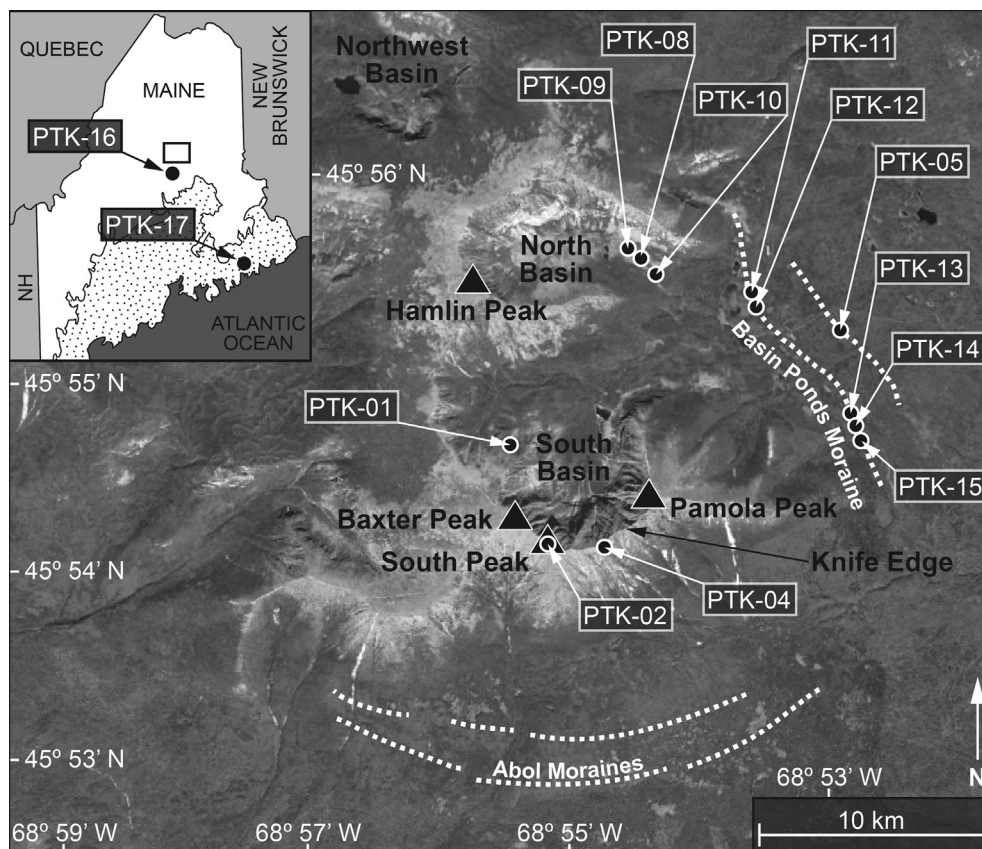


Fig. 1. Map of the study area. Main panel shows Google Earth satellite imagery of the Katahdin area, with relevant features labeled. Black and white circles denote the location of cosmogenic samples and white dashed lines show the location of moraines described in the text. Inset map shows the location of Katahdin in Maine, as well as two additional cosmogenic samples. Stippled pattern shows the region of post-glacial marine submergence.

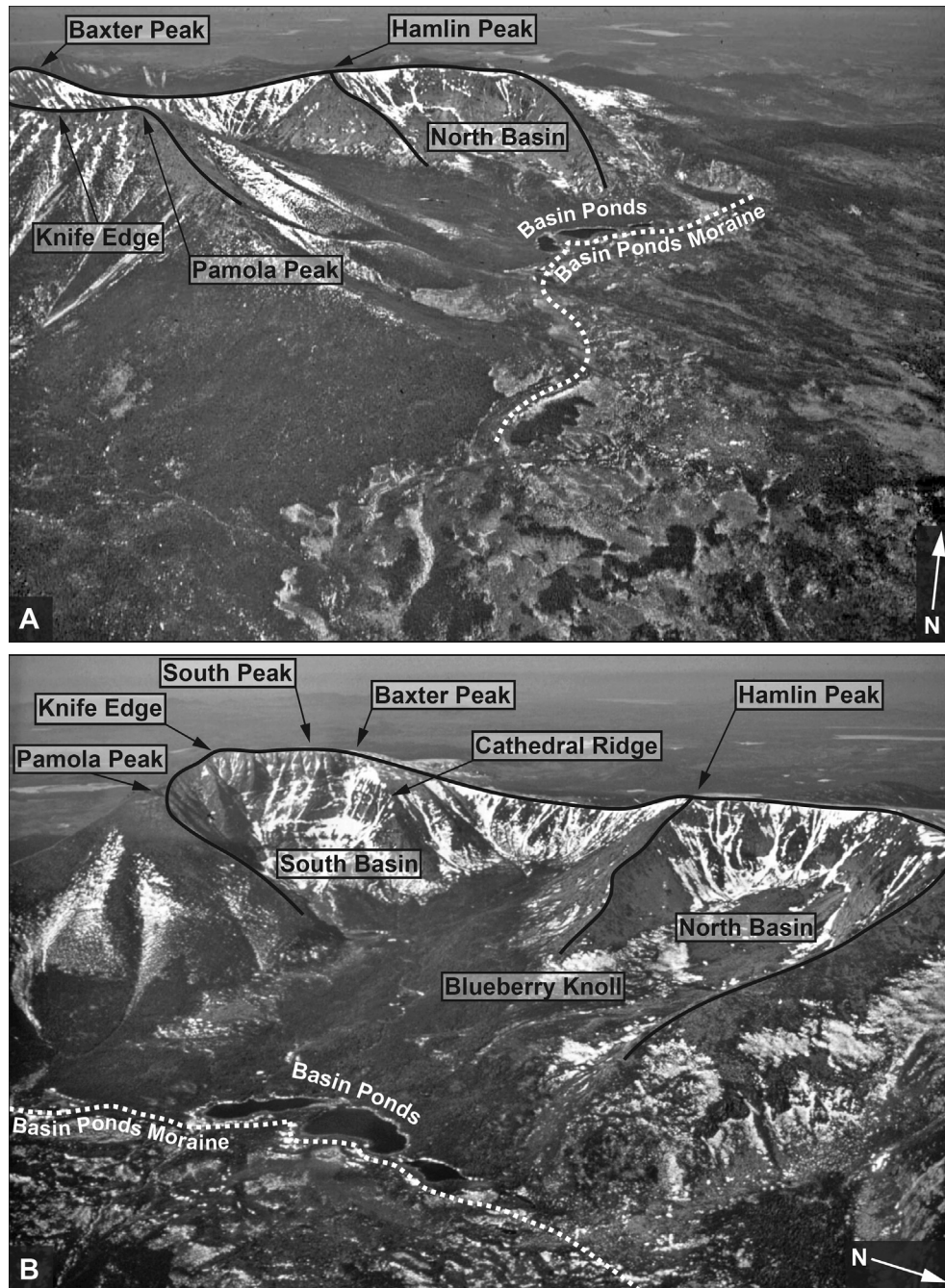


Fig. 2. Oblique air photographs of Katahdin. A. Aerial view looking northwest at the Katahdin massif, with the Basin Ponds moraine in the foreground. B. Aerial view looking southwest at the east-facing cirques. Relevant features mentioned in the text are labeled and ridge crests have been outlined in black. Photos by P.T. Davis.

1984; Waitt and Davis, 1988; Loso et al., 1998; Davis, 1999), to have been occupied by cirque glaciers in the late Wisconsinan. On Katahdin, Caldwell (1959, 1966, 1972, 1980, 1998) long held that alpine glaciation occurred both before and following deglaciation of the last continental ice sheet, whereas Davis (1976, 1989, 1999) and Davis and Davis (1980) countered that there is no indisputable evidence for cirque glaciers postdating continental ice recession. Davis (1976, 1999) maintains that looped recessional moraines typical of cirque glaciers do not occur on the floors of any of the cirques on Katahdin. Moreover, on the floor of North Basin (Fig. 1), Davis (1976, 1999) has identified roches moutonnées with steep sides facing obliquely up-cirque, along with the highest percentages of erratic pebbles in any cirques on the mountain, suggesting

that the last glacial erosion and deposition was that of continental ice flowing from the northwest onto the mountain.

The greatest controversy about the glacial history of Katahdin concerns the moraines found on the mountain near the mouths of the three, large, east-facing cirques. Tarr (1900), Antevs (1932), and Caldwell (1959, 1966, 1972, 1980, 1998) believed that the large prominent Basin Ponds moraine (Figs. 1, 2 and 3E) was a medial moraine formed between combined alpine glaciers from the three cirques and the still-active ice tongue of a continental ice sheet to the east. Thus, Caldwell (1959, 1966, 1972, 1980, 1998) believed that alpine glaciers were not only contemporaneous with ice sheet glaciation at the Basin Ponds moraine, but also post-dated ice sheet glaciation of the cirques. Davis (1976, 1989, 1999) countered that



Fig. 3. Photographs of cosmogenic sample sites. A. View looking north along Knife Edge at Baxter Peak with sample site for PTK-04 in foreground; B. Close up view of polished bedrock at PTK-04 (pocket knife for scale); C. Glacially molded bedrock on Cathedral Ridge at PTK-01, with Knife Edge in background; D. View looking west into North Basin (PTK-08 and PTK-09) from Blueberry Knoll (PTK-10); E. Sampling PTK-12 on Basin Ponds moraine; F. View looking north of bog behind recessional moraine with PTK-05; G. Pockwockamus Rock where PTK-16 was sampled from top surface; H. Boulder on Pineo Ridge moraine where PTK-17 was collected from top surface (photos by P.T. Davis; see online version of this article to view this figure in color).

alpine glaciers did not exist following ice sheet glaciation, did not map any moraines within the three large east-facing cirques, and interpreted the Basin Ponds moraine to be a lateral moraine built entirely by a mass of continental ice east of Katahdin rather than a medial moraine built by cirque glaciers. Davis (1976, 1989) also mapped two large terrace-like features that span most of the south flank of Katahdin, found the features to be composed of till with striated and faceted erratic clasts, and thus interpreted the ridges as lateral moraines (the Abol moraines; Fig. 1), marking a marginal position of continental ice to the south of the mountain, as cirques do not exist on the south side.

Efforts to provide a chronology for the cirques and moraines on Katahdin by ^{14}C dating of basal sediments from bogs and ponds have not been successful. Davis (1976, 1999) and Davis and Davis (1980) interpreted a basal age of 3050 ± 90 ^{14}C yr BP (I-7347; 2.99–3.45 cal ka, using CALIB 7.0, Reimer et al., 2014) in a sediment core from Chimney Pond in South Basin (Fig. 1) as thousands of years more recent than deglaciation of South Basin cirque. Likewise, basal ages from Lower Basin Pond (5665 ± 110 ^{14}C yr BP; I-7348; 6.28–6.70 cal ka) and a bog behind a recessional moraine (7070 ± 90 ^{14}C yr BP; SI-1049; 7.69–8.04 cal ka) downslope of the Basin Ponds moraine are thousands of years too young, probably

explained by long lag times required for the bouldery depressions to retain sediments (Davis and Davis, 1980). Basal sediments from Lower Togue Pond, about 10 km south of the Basin Ponds moraine, provided an age of $11,630 \pm 260$ ^{14}C yr BP (SI-2992; 12.93–14.04 cal ka), which is likely also a minimum limit, as are nearly all bog- and pond-bottom ^{14}C ages (Davis and Davis, 1980).

3. Study design and methods

The quartz-rich granite bedrock and boulders on Katahdin are well suited for cosmogenic exposure dating. Samples were collected with a hammer and chisel from flat-lying top surfaces of glacial boulders or bedrock exposures that appeared to be glacially molded, with negligible surface weathering, erosion, postglacial sediment cover, or topographic shielding. In the field, we measured latitude, longitude, and the thickness of each sample. We estimated sample site elevation from topographic maps.

We collected 13 samples from the Katahdin region and one sample from Pineo Ridge, about 170 km to the southeast (Fig. 1). Six samples were collected from the edifice of Katahdin: one each from bedrock outcrops on South Peak (PTK-02), the Knife Edge (PTK-04; Fig. 3A and B), and Cathedral Ridge (PTK-01; Fig. 3C) high on the mountain, two from boulders in North Basin (PTK-08 and PTK-09; Fig. 3D), and one from a boulder on Blueberry Knoll (PTK-10; Fig. 3D) at the mouth of North Basin cirque (Fig. 1). Five boulder samples (PTK-11 to PTK-15) were collected from the Basin Ponds moraine (Figs. 1, 2 and 3E). We sampled one boulder from a moraine outboard of the Basin Ponds moraine (PTK-05; Fig. 3F) and another from Pockwockamus Rock (PTK-16; Fig. 3G), about 14 km away in the lowland. The Pineo Ridge sample (PTK-17; Fig. 3H) was from a boulder lying on a moraine adjacent to the top surface of the glaciomarine delta.

Samples were prepared at University of Vermont between 1997 and 2000. Quartz was isolated with a series of dilute acid etches (Kohl and Nishiizumi, 1992) and dissolved in concentrated HF. Be and Al were isolated using pH-specific precipitation followed by cation exchange chromatography (Bierman and Caffee, 2002). Samples were prepared in four batches consisting of six samples and two process blanks in each batch. About 250 μg of ^9Be was added to each sample (1000 ppm SPEX Be standard) as carrier. Because samples contained substantial native ^{27}Al , no Al carrier was added. We measured the $^{10}\text{Be}/^9\text{Be}$ and $^{26}\text{Al}/^{27}\text{Al}$ ratios using accelerator mass spectrometry (AMS) at Lawrence Livermore National Laboratory. Measured $^{10}\text{Be}/^9\text{Be}$ ratios ranged from 2.11×10^{-13} to 5.75×10^{-13} and measured $^{26}\text{Al}/^{27}\text{Al}$ ratios ranged from 4.34×10^{-14} to 3.56×10^{-13} . ^{10}Be precisions averaged $2.5 \pm 0.5\%$ (1SD) and ^{26}Al precisions averaged $5.2 \pm 1.3\%$ (1SD). Be ratios were normalized to standards LLNL1000 or LLNL3000, with assumed ratios of 1000×10^{-15} and 3000×10^{-15} , respectively. Al ratios were normalized to standard KNSTD9919, with an assumed ratio of 9919×10^{-15} (see Supplementary data Table S1). Concentrations and $^{26}\text{Al}/^{10}\text{Be}$ ratios reported in Table 1 reflect normalization to standard values at the time of measurement; however, $^{26}\text{Al}/^{10}\text{Be}$ ratios have been normalized to the current accepted standard values in Table 2 (Nishiizumi et al., 2007; Balco et al., 2008).

To correct for backgrounds, we used the median process blank ratios of $2.44 \pm 0.93 \times 10^{-14}$ for $^{10}\text{Be}/^9\text{Be}$ ($n = 9$) and $2.38 \pm 1.31 \times 10^{-15}$ for $^{26}\text{Al}/^{27}\text{Al}$ ($n = 7$; median $\pm$ 1SD around the mean, see Supplementary data Table S2). Blanks used for background correction include those in batches of samples from New England glacial features all processed during the same timeframe and contained the same amount of ^9Be as the samples and ~ 2000 μg of ^{27}Al . We subtracted the median background ratio from each sample ratio and propagated the uncertainty (the SD of

Table 1
Sampling and cosmogenic isotopic data for Katahdin samples.

Sample name	Location	Sample type	Latitude ($^{\circ}\text{N}$)	Longitude ($^{\circ}\text{W}$)	Elevation (m a.s.l.)	^{10}Be conc. ($\times 10^5$ atoms g^{-1}) ^a	^{10}Be unc. ($\times 10^3$ atoms g^{-1}) ^a	^{26}Al conc. ($\times 10^5$ atoms g^{-1}) ^a	^{26}Al unc. ($\times 10^4$ atoms g^{-1}) ^a	$^{26}\text{Al}/^{10}\text{Be}$ ratio ^b	$^{26}\text{Al}/^{10}\text{Be}$ ratio unc.
PTK-01	Cathedral Ridge	Bedrock	45.91028	68.92191	1287	2.02	6.44	11.88	7.54	5.89	0.42
PTK-02	South Peak	Bedrock	45.90292	68.91852	1598	2.22	9.92	14.17	5.55	6.39	0.38
PTK-04	Knife Edge	Bedrock	45.90227	68.91327	1503	2.23	7.83	14.23	5.63	6.40	0.34
PTK-05	BP recess, moraine	Boulder	45.92154	68.88766	713	1.43	5.54	8.75	5.23	6.13	0.44
PTK-08	North Basin floor	Boulder	45.92911	68.90772	930	1.49	5.76	9.86	4.83	6.62	0.41
PTK-09	North Basin floor	Boulder	45.93019	68.90895	936	1.79	6.36	12.05	5.41	6.74	0.39
PTK-10	Blueberry Knoll	Boulder	45.92781	68.90494	930	1.40	6.11	9.28	4.95	6.64	0.46
PTK-11	BP moraine	Boulder	45.92381	68.89444	748	1.26	5.37	9.14	5.46	7.27	0.53
PTK-12	BP moraine	Boulder	45.92381	68.89414	749	1.33	5.72	9.22	4.23	6.91	0.43
PTK-13	BP moraine	Boulder	45.91180	68.87747	744	1.56	5.58	8.73	3.63	5.61	0.31
PTK-14	BP moraine	Boulder	45.91201	68.87839	750	2.28	6.31	14.89	5.98	6.52	0.32
PTK-15	BP moraine	Boulder	45.91180	68.87870	750	1.29	6.52	7.22	3.24	5.58	0.38
PTK-16	Pockwockamus Rock	Boulder	45.75000	68.87500	150	0.79	5.30	4.40	2.34	5.56	0.47
PTK-17	Pineo Ridge	Boulder	44.67335	67.82246	60	0.79	5.47	5.63	4.94	7.09	0.79

^a Be analyses were normalized to standard LLNL1000 or LLNL3000; Al analyses were normalized to standard KNSTD9919 (see data repository).

^b Measured $^{26}\text{Al}/^{10}\text{Be}$ ratio, dependent on standards used for normalization.

Table 2
Cosmogenic nuclide exposure age data for Katahdin samples.

Sample name	Location	Sample type	¹⁰ Be exposure age (ka) ^a	¹⁰ Be Internal unc. (ka)	¹⁰ Be external unc. (ka)	²⁶ Al exposure age (ka) ^a	²⁶ Al Internal unc. (ka)	²⁶ Al external unc. (ka)	²⁶ Al/ ¹⁰ Be ratio ^b	²⁶ Al/ ¹⁰ Be ratio unc.	Uncertainty-weighted average ¹⁰ Be and ²⁶ Al age	Uncertainty-weighted average ¹⁰ Be and ²⁶ Al age unc.
PTK-01	Cathedral Ridge	Bedrock	14.7	0.5	0.9	14.8	0.9	1.2	6.79	0.48	14.7	0.69
PTK-02	South Peak	Bedrock	12.6	0.6	0.8	13.7	0.5	0.9	7.36	0.44	13.1	0.60
PTK-04	Knife Edge	Bedrock	13.5	0.5	0.8	14.7	0.6	0.9	7.38	0.39	14.0	0.60
PTK-05	BP recess. moraine	Boulder	16.5	0.6	1.0	15.9	1.0	1.3	6.52	0.49	16.3	0.80
PTK-08	North Basin floor	Boulder	15.5	0.6	1.0	16.3	0.8	1.1	7.09	0.44	15.8	0.73
PTK-09	North Basin floor	Boulder	18.6	0.7	1.1	19.8	0.9	1.3	7.22	0.41	19.1	0.85
PTK-10	Blueberry Knoll	Boulder	14.6	0.6	1.0	15.4	0.8	1.1	7.10	0.49	14.9	0.72
PTK-11	BP moraine	Boulder	15.4	0.7	1.0	17.8	1.1	1.4	7.78	0.57	16.2	0.81
PTK-12	BP moraine	Boulder	16.0	0.7	1.0	17.5	0.8	1.2	7.39	0.47	16.6	0.77
PTK-13	BP moraine	Boulder	17.8	0.6	1.1	16.9	0.7	1.1	6.42	0.35	17.3	0.76
PTK-14	BP moraine	Boulder	25.6	0.7	1.4	28.6	1.2	1.8	7.53	0.37	26.8	1.12
PTK-15	BP moraine	Boulder	14.6	0.7	1.0	13.8	0.6	0.9	6.42	0.43	14.2	0.68
PTK-16	Pockwockamus Rock	Boulder	15.1	1.0	1.2	14.1	0.8	1.0	6.38	0.56	14.5	0.80
PTK-17	Pineo Ridge	Boulder	16.4	1.1	1.4	19.8	1.8	2.0	8.16	0.91	17.5	1.14

^a Age calculations were done in CRONUS (version 2.1, constants version 2.2.1, Balco et al., 2008) with the northeastern North American production rate (Balco et al., 2009) assuming zero erosion, no inheritance, and a density of 2.7 g cm⁻³.

^b Ratios from CRONUS (version 2.1, constants version 2.2.1, Balco et al., 2008) corrected for currently accepted standard values per Nishiizumi et al. (2007).

the blanks) in quadrature. We chose to subtract the median blank value because it best reflects the most likely blank.

Exposure ages were calculated using the CRONUS Earth online calculator (Balco et al., 2008), main calculator version 2.1 and constants version 2.2.1, scaling for the standards used for normalization of the measured isotopic ratios and the recalibration of those standards (Nishiizumi et al., 2007). We used the regionally-calibrated northeastern North American sea-level production rates of 3.93 ± 0.19 atoms g⁻¹ yr⁻¹ for Be and 26.5 ± 1.3 atoms g⁻¹ yr⁻¹ for Al as described in Balco et al. (2009). Calculations in CRONUS used the Lal (1991)/Stone (2000) constant production rate model and scaling scheme and assumed no erosion, shielding, or inherited nuclides, and a density of 2.7 g cm⁻³.

We report ¹⁰Be, ²⁶Al, and average exposure ages inversely weighted by the uncertainty of both the ¹⁰Be and ²⁶Al ages. We calculate uncertainty-weighted average ages and their uncertainty using external age uncertainties from CRONUS and the following equations:

$$\text{Weighted average age} = \frac{\frac{^{10}\text{Be age}}{(^{10}\text{Be age unc})^2} + \frac{^{26}\text{Al age}}{(^{26}\text{Al age unc})^2}}{\frac{1}{(^{10}\text{Be age unc})^2} + \frac{1}{(^{26}\text{Al age unc})^2}}$$

Unc. of weighted average age

$$= \sqrt{\frac{1}{(^{10}\text{Be age unc})^2} + \frac{1}{(^{26}\text{Al age unc})^2}}$$

4. Results

We measured late Pleistocene exposure ages (12.6–28.6 ka; ¹⁰Be and ²⁶Al; n = 28) for all 14 samples in this study (Table 2 and Fig. 4). In the text below, we report the ¹⁰Be and ²⁶Al uncertainty-weighted average age because the average ²⁶Al/¹⁰Be ratio (7.11 ± 0.54 , 1SD) is indistinguishable from the production ratio of 6.75 (Balco et al., 2008) and because ²⁶Al and ¹⁰Be ages are well correlated with an intercept of -1.81 ± 2.01 ky (inseparable from 0) and a slope consistent with unity (1.16 ± 0.12 , 1SD). Uncertainty-weighted average ¹⁰Be and ²⁶Al ages range from 13.1 to 26.8 ka (Table 2).

Exposure ages from polished or molded bedrock samples high on the Katahdin uplands (PTK-02 and -04, South Peak and the Knife

Edge, 1600 and 1500 m asl, Fig. 3A and B) are 13.1 and 14.0 ka, respectively. Glacially molded bedrock on Cathedral Ridge (PTK-01; Fig. 3C), about 1300 m asl, has an exposure age of 14.7 ka. Boulders at the base of the edifice (~930 m asl) in North Basin (PTK-08 and -09; Fig. 3D) and at Blueberry Knoll (PTK-10; Fig. 3D) have exposure ages of 15.8, 19.1, and 14.9 ka. Boulders on the Basin Ponds moraine (PTK-11 to -15) have ages ranging from 14.2 to 26.8 ka. One boulder (PTK-05) from a moraine downslope from the Basin Ponds moraine has an exposure age of 16.3 ka. The Pockwockamus (PTK-16; Fig. 3G) and Pineo Ridge (PTK-17; Fig. 3H) boulders have ages of 14.5 and 17.5 ka, respectively.

If we consider the six samples from the edifice of Katahdin as a population (PTK-01, PTK-02, PTK-04, PTK-08, PTK-09, PTK-10), the mean age is 15.3 ± 2.1 ka (1SD, n = 6) with a median age of 14.8 ka. These ages are statistically similar to those of the Basin Ponds moraine samples, which average 16.1 ± 1.2 ka (1SD) with a median of 16.3 ka (n = 5; PTK-05 and PTK-11 through PTK-15, excluding PTK-14 as an outlier because it likely contains nuclides inherited from a prior period of exposure). Both of the above ranges overlap within 1 SD uncertainty with the ages of Pockwockamus Rock (PTK-16; 14.5 ± 0.8 ka) and the Pineo Ridge boulder (PTK-17; 17.5 ± 1.1 ka).

5. Discussion

Cosmogenic exposure ages suggest that continental ice covered Katahdin's uplands during the late Wisconsinan, refute the possibility that the Basin Ponds moraine was formed by cirque glaciers on Katahdin following recession of continental ice in the area, and suggest that cirque glaciers did not reform after continental ice ablated. We also show that more than 1000 m of continental ice surface lowering took place very rapidly, within the resolution of the exposure age data, and that ice likely retreated rapidly across southeastern Maine.

5.1. Vertical and horizontal distribution of exposure ages in high-relief terrain

Although samples from the Katahdin uplands all record post-LGM (Last Glacial Maximum) deglaciation ages, they have the youngest exposure ages of the dataset. These ages are all from bedrock, which more commonly carries inherited cosmogenic nuclides than boulder samples (Bierman et al., 1999; Briner et al.,

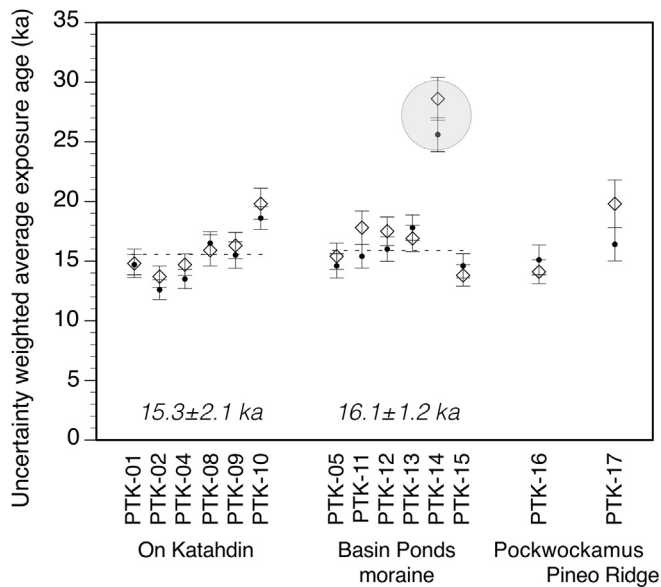


Fig. 4. Cosmogenic exposure ages (black dot for ^{10}Be , open diamond for ^{26}Al) for samples from Katahdin area as well as Pineo Ridge. Samples from the edifice are considered as one population and samples from the Basin Ponds and recessional moraines are considered as another population. Sample surrounded by gray circle is considered an outlier. Bars show external uncertainty of ages according to CRONUS calculator; dashed lines are means of uncertainty-weighted average ^{10}Be and ^{26}Al exposure ages that are listed with 1 SD uncertainty.

2003, 2006). The average exposure age for the three polished or molded bedrock samples on the Katahdin uplands (PTK-01, Cathedral Ridge, 1300 m asl, Fig. 3C; PTK-02, South Peak, 1600 m asl, Fig. 3A; PTK-04, the Knife Edge, 1500 m asl, Fig. 3B) is 13.9 ± 0.8 ka. This young average (in comparison to samples collected lower on the mountain) could suggest delayed exhumation from a thin till cover, burial by snow or ice, or limited erosion after initial exposure. No till is present in these locations today (Fig. 3A–C), nor does deep snow cover now persist in these ridge locations; however, the presence of glacial polish is consistent with thin till or snow/ice cover protecting outcrops from weathering for at least part of the late Pleistocene and/or Holocene.

North Basin cirque was free of ice before about 16 ka. The average exposure age of the three boulders in North Basin is 16.6 ± 2.2 ka (PTK-08, PTK-09, and PTK-10). The boulder sample from the top of Blueberry Knoll at the mouth of North Basin has an exposure age of 14.9 ± 0.7 ka (PTK-10, Table 2, Fig. 3D). Caldwell (1959, 1966, 1972, 1980, 1998) suggested that Blueberry Knoll (Fig. 2B) was an end moraine deposited by a cirque glacier, but Davis (1976, 1999) interpreted the shape of the knoll and seismic data to suggest a bedrock origin, an argument buttressed by the observation that soils here contain 15% erratic pebbles sourced from continental ice (Davis, 1976; Waitt and Davis, 1988). Between Blueberry Knoll and the head of North Basin is a ridge of bouldery material that Caldwell (1966, 1972, 1980, 1998) considered to be an end moraine, but Davis (1976, 1999) believed to be hummocky ground moraine. A boulder on the ridge (PTK-09; Fig. 3D) has an exposure age of 19.1 ± 0.9 ka (Table 2). A third boulder (PTK-08; Fig. 3D) from a hummock on the floor of North Basin about half way between Blueberry Knoll and the two small ponds (Figs. 1 and 3D) has an age of 15.8 ± 0.7 ka (Table 2). Taken together, these ages argue against the existence of post-LGM cirque glaciation of Katahdin.

Boulders on the Basin Ponds moraine and a recessional moraine downslope (Figs. 1 and 2) have exposure ages that average

16.1 ± 1.2 , indistinguishable from the 16.6 ± 2.2 ka average age for the three boulders in North Basin cirque and Blueberry Knoll. Because nearly all boulder-size material in the moraine is composed of Katahdin granite, Caldwell (1966, 1972, 1980, 1998) inferred that the Basin Ponds moraine was deposited by ice originating from the three large cirques to the west, merging with and perhaps overriding the continental ice here. However, up to 28% of the 2–5 cm (median axis) pebbles in the moraine are erratic suggesting that they originated from continental ice (Davis, 1976; Waitt and Davis, 1988). Moreover, the Basin Ponds moraine (Fig. 1) extends both north and south well beyond the mouths of the cirques, wraps tightly around the prominent bedrock ridge to the south, and does not drop in elevation north to south, but remains within the 2400- to 2500-foot contours (730–760 m asl). A till exposure in the recessional moraine downslope from the Basin Ponds moraine (Fig. 1) contains 40–60% erratic (non-granitic) pebbles (Davis, 1976; Waitt and Davis, 1988), suggesting that the recessional moraine was also formed by a shrinking mass of continental ice located in the lowlands east of Katahdin. Taken together with the Abol moraines on the south slope of Katahdin (Fig. 1), Davis (1976, 1999) inferred that the Basin Ponds moraine was formed by an advance of remnant continental ice in the lowlands to the south and east of the mountain. Our cosmogenic exposure ages suggest that the Basin Ponds and associated recessional moraine were deposited ~16.1 ka ($n = 5$), including the exposure age of 16.3 ± 0.8 ka for the boulder we sampled on the recessional moraine (PTK-05, Fig. 3F).

Our data suggest rapid continental ice surface lowering in the Katahdin area (Fig. 4) because there is no statistically significant difference in exposure age from top to bottom of the edifice, nor is there any difference in the average age of the edifice samples and those from the Basin Ponds moraine. In addition, a sample from the top of Pockwockamus Rock (~2.5 m high, PTK-16, Fig. 3G) about 0.3 km east of the south entrance road to Baxter State Park has an exposure age of 14.5 ± 0.8 ka (Table 2), indistinguishable from the average boulder age (16.1 ± 1.2) on the Basin Ponds and recessional moraines. These exposure ages are consistent with (but older than) minimum-limiting calibrated basal sediment ^{14}C ages from Lower Togue Pond, Upper South Branch Pond, and Matteseunk Lake in the Katahdin area (Table 3, Fig. 5).

5.2. Possible correlation of the Basin Ponds moraine and the Pineo Ridge moraine complex

Our data indicate that there is no difference in age, within the resolution of our exposure age data, between moraine samples from Katahdin and one sample from the Pineo Ridge moraine complex near the coast in southeastern Maine (Fig. 1). Both of these large moraines record a much more robust ice margin than do the lower relief morainal banks produced at the grounding line in coastal Maine (Smith, 1982; Hunter and Smith, 2001) or the small ribbed moraines found above the marine limit in central Maine (Caldwell et al., 1985). There are no prominent moraines between the Pineo Ridge moraine complex and the Basin Ponds moraine on Katahdin about 170 km to the northwest, but rather mostly kame and kettle topography along with ground moraine. Thus, the Basin Ponds moraine at about 750 m asl and the Pineo Ridge moraine complex at about 65 m asl may have been formed contemporaneously along an active ice margin with a surface slope of about 4 m/km, similar to surface slopes for parts of modern ice sheets. The ice surface profile shown in Fig. 5 is similar in form to one drawn by Shreve (1985) based on esker system paths in Maine. Our cosmogenic exposure ages are consistent with both of these two prominent moraines being deposited during a cold interval known as the Oldest Dryas in Europe (Alley et al., 1993; von Grafenstein et al., 1999).

Table 3
¹⁴C ages that provide constraint on deglaciation across southeastern Maine.

Site name	Latitude (°N)	Longitude (°W)	Material	Lab number	Uncalibrated ¹⁴ C age (yr BP)	Calibrated age (ka BP), considering 1000 yr marine residence time ^{a,b}	Calibrated age (ka BP), considering combined IntCal04/Marine04 ^b	Reference
Upper South Branch Pond	46.090	68.893	Terrestrial veg.	SI-4463	10,965 ± 230	12.87 (12.40–13.31)	12.87 (12.40–13.31)	Anderson et al., 1986
Lower Togue Pond	45.825	68.881	Terrestrial veg.	SI-2992	11,630 ± 260	13.48 (12.97–14.10)	13.48 (12.97–14.10)	Davis and Davis, 1980
Mattaseunk Lake	45.590	68.378	Marine shells	OS-1322	13,450 ± 75	14.59 (14.20–15.02)	15.62 (15.32–15.89)	Dorion et al., 2001
T2 R8 NWP	45.368	68.549	Marine shells	OS-3160	13,300 ± 65	14.26 (14.04–14.67)	15.40 (15.18–15.66)	Dorion et al., 2001
Dover-Foxcroft	45.204	69.181	Marine shells	OS-11022	13,550 ± 60	14.86 (14.43–15.14)	15.79 (15.55–16.03)	Dorion et al., 2001
Boyd Lake	45.170	68.924	Marine shells	AA-9293	13,075 ± 90	13.93 (13.74–14.16)	15.06 (14.61–15.37)	Dorion et al., 2001
Gould Pond	44.993	69.319	Marine shells	AA-7463	13,290 ± 85	14.27 (13.98–14.74)	15.39 (15.14–15.70)	Dorion et al., 2001
Lily Lake	44.828	67.102	Marine shells	OS-2151	13,350 ± 50	14.36 (14.11–14.70)	15.47 (15.26–15.70)	Kaplan, 1999
Carrying Place Bluff	44.813	66.979	Marine shells	OS-2075	13,800 ± 80	15.26 (15.02–15.59)	16.12 (15.82–16.38)	Dorion et al., 2001
Turner Brook	44.669	67.250	Marine shells	AA-7461	13,810 ± 90	15.28 (15.00–15.63)	16.13 (15.83–16.40)	Dorion et al., 2001
Sprague Neck	44.664	67.319	Marine shells	AA-7462	13,370 ± 90	14.44 (14.08–14.92)	15.50 (15.21–15.80)	Dorion et al., 2001
Dennison Point	44.642	67.242	Marine shells	OS-2154	14,000 ± 85	15.55 (15.26–15.84)	16.39 (16.11–16.70)	Kaplan, 1999
Long Pond	44.595	68.023	Marine shells	OS-3466	12,950 ± 120	13.80 (13.54–14.10)	14.78 (14.22–15.20)	Dorion et al., 2001
Toddy Pond	44.543	69.057	Marine shells	OS-2662	13,000 ± 60	13.86 (13.73–14.04)	14.95 (14.60–15.21)	Dorion et al., 2001
Sargent Mtn. Pond	44.334	68.270	Org. sediment	Beta-240351	13,260 ± 50	15.94 (15.75–16.13)	15.94 (15.75–16.13)	Norton et al., 2011

^a All ¹⁴C ages of marine shells corrected by subtraction of 1000 years for marine reservoir effect in Maine following Thompson et al. (2011).

^b Age estimates include the median intercept and the minimum and maximum ages in parentheses based on 2 standard deviations from minimum and maximum intercepts using CALIB 7.0 (Reimer et al., 2014).

5.3. Rapid rates of ice retreat

Our cosmogenic exposure ages also suggest that the Basin Ponds and Pineo Ridge moraines formed within a time of rapid retreat of continental ice during the deglaciation of Maine. A plot of exposure ages from the Katahdin uplands, the Basin Ponds and recessional moraines, the lowland southeast of the mountain, and the Pineo Ridge moraine in southeastern Maine shows no significant differences in the deglaciation age of these features (Fig. 4). We recognize that the 3.5-m high boulder we sampled on the Pineo Ridge moraine complex in southeastern Maine (PTK-17, Figs. 1 and 3H) has a ¹⁰Be exposure age of 16.4 ± 1.4 ka and a ²⁶Al exposure age of 19.8 ± 2.0 ka (Table 2), older than many ages we measured near and on Katahdin. But, we also note that both the ¹⁰Be and ²⁶Al ages for PTK-17 have significantly higher uncertainty than other data in this paper and that the 07KNSTD-normalized ²⁶Al/¹⁰Be ratio exceeds the nominal production ratio (²⁶Al/¹⁰Be = 6.75) by more than 1 SD (8.16 ± 0.91), consistent with the ²⁶Al age being an overestimate.

The timing of deglaciation in southeastern Maine, near where we sampled, has been constrained using ¹⁴C ages of marine shells deposited in post-glacial marine sediments and by dating organic material from bog and pond bottoms (e.g., Kaplan, 1999, 2007; Dorion et al., 2001; Borns et al., 2004; Dyke, 2004; Norton et al., 2011). We summarize relevant ¹⁴C ages in Table 3 considering that recent research, focused on a single site in Portland, Maine (Fig. 5; Thompson et al., 2011), suggests a higher (~1000 year) post-deglaciation marine reservoir correction for marine shells in the Penobscot Lowland than the global average of ~400 years used in CALIB 7.0 (Reimer et al., 2014) or the 600 years used by Borns et al. (2004).

Radiocarbon dating (Table 3) suggests that continental ice left the Maine coast between 15 and 16 ka, depending on which samples are deemed representative and which marine reservoir correction is adopted. For example, near-basal organic material in a core collected from Sargent Mountain Pond on Mt. Desert Island (Norton et al., 2011; Beta-240351, Table 3) gives a calibrated age of 15.75–16.13 ka. Several other ages of marine shells from near the coast are similarly old, with median calibrated ages ranging from 15.3 to 15.5 ka (Thompson et al., 2011, 1000-year marine reservoir correction) and 16.1 to 16.4 ka (Reimer et al., 2014; IntCal04/Marine04 marine reservoir correction). When all of the cosmogenic exposure ages and all of the calibrated radiocarbon ages from the coastal and interior lowland of Maine are considered as probability

density functions, the populations appear similar (Fig. 6). This similarity suggests that the deglaciation of the Katahdin edifice and the deposition of the Basin Ponds moraine occurred, within the resolution of our cosmogenic nuclide exposure ages (Table 2) and the ¹⁴C ages (Table 3), at the same time that the coast of Maine deglaciated. This conclusion is robust regardless of which reservoir correction is used.

Retreat isochrones constructed by Borns et al. (2004), who compiled 77 ¹⁴C ages, also suggest that the continental ice sheet grounding line retreated rapidly between Maine's coast and interior across the Penobscot Lowland (their Fig. 1; redrawn as our Fig. 5), which was submerged by rising post-glacial sea level. Their work suggests that it took between 1000 and 1200 ¹⁴C years to deglaciate the 170 km between the coast and Katahdin, implying an ice margin retreat rate of 140–170 m/¹⁴C yr.

This marine submergence was likely coincidental with a calving ice margin, as described for several other areas around the globe, including the St. Lawrence Lowland (Thomas, 1977; Occhietti et al., 2004; Richard and Occhietti, 2005), Maritime Canada and New England (Borns and Hughes, 1977; Hughes et al., 1985), the Columbia Glacier in southern Alaska (Brown et al., 1982), and Antarctica (Hughes, 2002). The extent of post-glacial marine submergence reaches within about 50 km of Katahdin (Figs. 1 and 5) and could explain the rapid retreat of the continental ice sheet grounding line and rapid thinning of the ice sheet in the Katahdin area. The last remaining continental ice in New England likely resided in the lowlands of northern Maine during the Younger Dryas chron (Borns et al., 2004), but not on Katahdin (Fig. 5).

Several other recent cosmogenic studies have documented rapid (instantaneous within the resolution of the ¹⁰Be and ²⁶Al chronometer) ice margin retreat at the last termination (Briner et al., 2007, 2013; Young et al., 2012, 2013; Gjermundsen et al., 2013; Kelley et al., 2013). For example, in eastern Baffin Island, Briner et al. (2009) found that a Laurentide outlet glacier retreated 110 km through Sam Ford Fjord at ~9.5 ka, with the most rapid retreat occurring in the deepest parts of the fjord. Across Baffin Bay, in Upernavik northwestern Greenland, Corbett et al. (2013) documented a marine-terminating outlet glacier that retreated 100 km in several centuries at about 11.3 ± 0.5 ka. At the same time (10.8 ± 0.3 ka), Helheim Glacier in eastern Greenland retreated through its 80-km fjord in less than a millennium (Hughes et al., 2012).

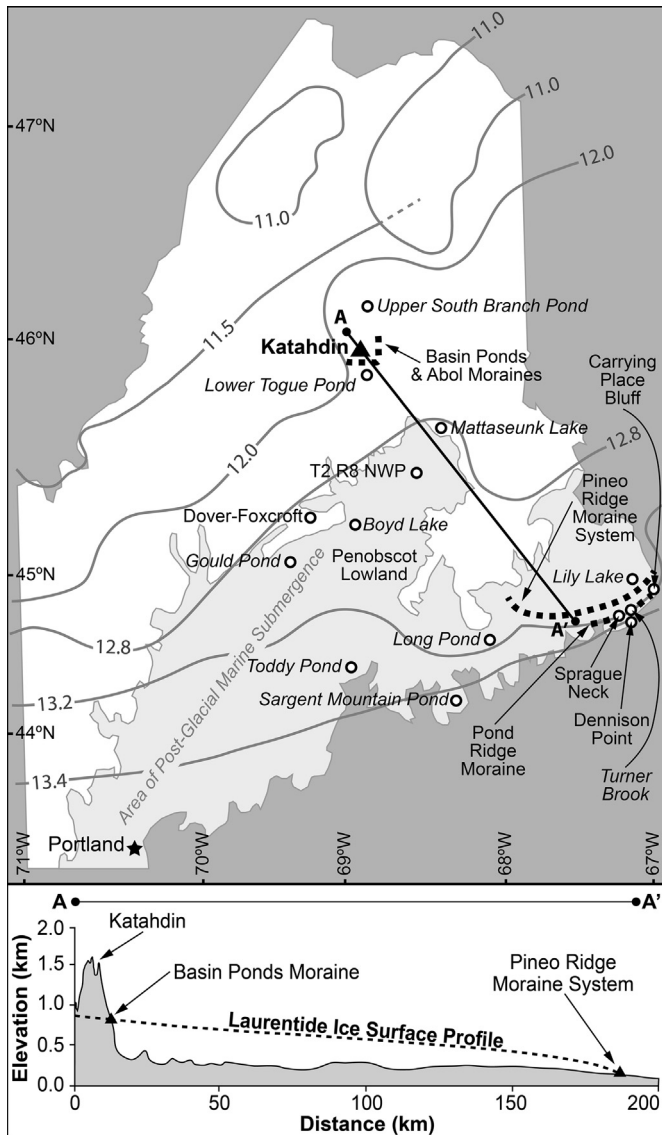


Fig. 5. Glacier margin ice retreat isochron map for Maine (revised from [Borns et al., 2004](#)). Isochrones are ^{14}C ages. Thick dashed lines denote the location of major moraines; black and white circles show the location of ^{14}C -dated sites in [Table 3](#) and discussed in the text. Gray area shows the region of post-glacial marine submergence. Ice surface profile has vertical exaggeration of $\sim 25\times$.

5.4. Comparison of glacial retreat in Maine with northern New Hampshire

The Littleton moraine complex on the northern side of the Presidential Range in northern New Hampshire marks a stillstand of the Laurentide Ice Sheet as its margin retreated to the northwest ([Thompson et al., 1996, 1999](#)). The cosmogenic exposure ages for the Basin Ponds moraine on Katahdin (16.1 ka) are about 2300 years older than the Littleton moraine, the ^{10}Be age of which is based on four exposure ages averaging 13.8 ± 0.3 ka ([Balco et al., 2009](#)) and constraints provided by the glacial Lake Hitchcock varve chronology in western New Hampshire ([Ridge et al., 1999, 2012](#)). The Littleton moraine complex is thought to be correlative with the short-lived European Older Dryas cool interval ([Thompson et al., 1996, 1999; Ridge, 2004; Balco et al., 2009; Ridge et al., 2012](#)), whereas the Katahdin and other cosmogenic exposure ages in Maine reported

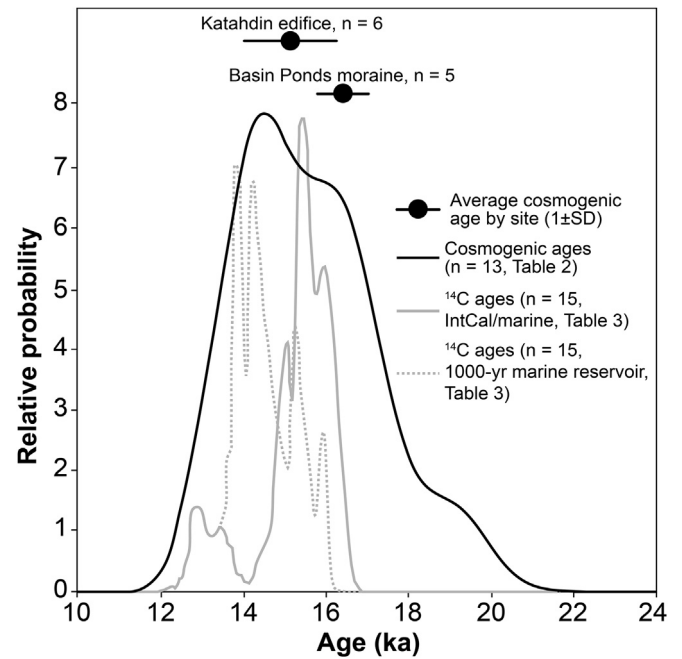


Fig. 6. Summed probability density plots for uncertainty-weighted, average ^{10}Be and ^{26}Al exposure ages as well as calibrated organic and reservoir-corrected marine shell ^{14}C ages. Black line shows cosmogenic exposure ages ($n = 13$, [Table 2](#), omitting sample PTK-14). Gray line shows ^{14}C ages ($n = 15$, [Table 3](#)) using CALIB 7.0 and the combined IntCal04/Marine04 dataset for calibration. Dashed gray line shows ^{14}C ages using a 1000-year marine reservoir correction ([Thompson et al., 2011](#)) and the CALIB 7.0 IntCal dataset for calibration ([Reimer et al., 2014](#)).

here appear to be correlative with the earlier and more prolonged Oldest Dryas cold interval ([Alley et al., 1993; von Grafenstein et al., 1999](#)). One possible explanation for the discrepancy in these ^{10}Be exposure histories from northern New England is that the continental ice sheet disintegrated in Maine earlier than it did in the areas to the west because of a greater influence of calving bay ice margins in the St. Lawrence valley ([Occhietti et al., 2004; Richard and Occhietti, 2005](#)) and the Penobscot Lowland ([Dorion et al., 2001; Borns et al., 2004, Fig. 5](#)). Another possible explanation is systemic cosmogenic nuclide inheritance in the boulder samples from the Katahdin area and Pineo Ridge, although the coherence of ages from the Basin Ponds moraine makes this explanation less likely.

6. Conclusions

Our analysis of cosmogenic exposure ages from the Katahdin area of Maine allow us to conclude that:

- 1.) The Katahdin area was deglaciated between 16 and 15 ka, several thousand years earlier than suggested by minimum-limiting ^{14}C ages from pond and bog bottoms.
- 2.) The Basin Ponds moraine was formed by a stillstand or re-advance of remnant lowland continental ice, and not by cirque glaciers.
- 3.) Sizeable cirque glaciers did not reform following continental ice sheet deglaciation, so cirques visible on Katahdin must have been primarily carved before the last overriding of the Laurentide Ice Sheet.
- 4.) Cosmogenic exposure ages from summit areas, the cirques, the Basin Ponds moraine, and the lowland surrounding Katahdin are indistinguishable, suggesting that the continental ice surface lowered rapidly during deglaciation.

- 5.) The exposure age for one boulder on the Pineo Ridge moraine complex in southeastern Maine is similar to but slightly older than exposure ages from Katahdin. Considered along with extensive radiocarbon data, the cosmogenic exposure age data suggest that the Basin Ponds and Pineo Ridge moraines formed contemporaneously along the ice margin during a time of generally rapid continental ice retreat across a 170-km transect from the coast to the interior, retreat that may have been facilitated by a marine calving bay ice margin in the Penobscot Lowland.

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Appendix A. Supplementary data

Supplementary data related to this article can be found at <http://dx.doi.org/10.1016/j.quascirev.2015.03.021>.

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