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NEIGC 2017 Day Three: October 1

Oct 1st, 2017

C1: Applying the Cosmogenic Nuclide Dipstick Model for Deglaciation of Mt. Washington

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Davis, P.T., Koester, A.J., Shakun, J.D., Bierman, P.R., and Corbett, L.B., 2017, Applying the Cosmogenic Nuclide Dipstick Model for Deglaciation of Mt. Washington in Johnson, B. and Eusden, J.D., ed., Guidebook for Field Trips in Western Maine and Northern New Hampshire: New England Intercollegiate Geological Conference, Bates College, p. 247-272. https://doi.org/10.26780/2017.001.0014

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APPLYING THE COSMOGENIC NUCLIDE DIPSTICK MODEL FOR DEGLACIATION OF MT. WASHINGTON

by
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INTRODUCTION

The purpose of this field trip includes the following: 1) describe the motivation and methods being used for a NSF-funded project to understand deglaciation of the mountains in northern New England and adjacent areas, 2) review past glacial history studies in the Presidential Range, 3) examine field sampling sites and initial results from Mt. Washington, one of the key mountainous areas in our study, and 4) discuss implications for global sea-level change during the last deglaciation in New England.

The late Pleistocene ice sheets were important agents of land surface and climate change during the Ice Ages of the last 2.8 million years. Research over the past several decades has generated ever more precise reconstructions of ice sheet extent histories based on organic ^{14}C (e.g., Dyke, 2004), cosmogenic nuclides (e.g., Balco et al., 2002; Carlson et al., 2007; Rinterknecht et al., 2006), and varve dating (e.g., Ridge et al., 2012). In contrast, the thickness evolution of the ice sheets has been far more difficult to constrain, largely because the ice sheets generally covered flat regions. In the rare instances where ice sheets covered mountainous areas such as northern New England, the lag time was highly variable between deglaciation and the deposition of organic matter for ^{14}C dating at higher elevations (Davis and Davis, 1980). While a robust relationship between ice sheet area and volume exists for glaciers in equilibrium (Paterson, 1994), this need not be the case for a deglaciating ice sheet due to, for instance, changes in basal temperature and subglacial meltwater, ice streaming, ice-ocean interactions, interruption of ice flow in regions of complex topography, and elevation-temperature and precipitation feedbacks (Fyke et al., 2014).

Accurate ice sheet thickness reconstructions are important for several reasons. (1) Ice sheet orography is a critical boundary condition for modeling paleoclimate during the last deglaciation (e.g., Liu et al., 2009). For instance, the height of ice sheets has a direct effect on surface temperature through lapse rate cooling, but also has downstream effects related to atmospheric planetary waves that control heat flow and storm tracks, and the strength of Atlantic Meridional Overturning Circulation (AMOC) (Ullman et al., 2014). (2) Ice sheet thickness must be known in order to quantify ice volume changes, and thus the contribution of individual ice sheets to global sea-level rise and the attendant freshwater forcing to the ocean (Carlson and Clark, 2012). (3) The timing and rate of ice sheet thinning sheds light on how ice sheets responded to the overall global warming and abrupt climate changes of the last deglaciation (Gregoire et al., 2012). (4) Ice thickness reconstructions can help validate numerical ice sheet models, which are important for understanding the processes of deglaciation as well as improving projections of ice sheet responses to future global warming (Stokes et al., 2015).

The Greenland ice core record reveals abrupt Northern Hemisphere warming and cooling events during the last deglaciation (Clark et al., 1999; Andersen et al., 2004) that have been linked to variations in the strength of the AMOC and its associated northward heat transport (Clark et al., 1996; McManus et al., 2004). Heinrich Stadial 1, a cooling event between ~19 and 14.6 ka, is thought to have occurred due to freshwater forcing from the Northern Hemisphere ice sheets weakening the AMOC. Recovery of the AMOC at ~14.6 ka then produced an abrupt warming into the Bølling-Allerød interstadial (Liu et al., 2009). Meltwater Pulse 1a (MWP-1A), a sea-level rise event of 14-18 m in 350 years (Deschamps et al., 2012; Carlson et al., 2012), occurred synchronously with the Bølling interstadial. However, it is unclear from which ice sheet MWP-1A was sourced (Liu et al., 2015), though two scenarios have been proposed. The first is the 'Northern scenario' where melting Northern Hemisphere ice sheets caused MWP-1A, leading to a weakening of the AMOC and Older Dryas cooling after the Bølling warming (Fairbanks, 1989; Manabe and Stouffer, 1995; Peltier, 2005; Peltier and Fairbanks, 2006). The second is the 'Southern scenario', which suggests the melting of the Antarctic Ice Sheet caused MWP-1A, triggering a reactivation of the AMOC and the Bølling warming (Clark et al., 1996; Weaver et al., 2003; Bassett et al., 2005).

Modern studies of the Greenland Ice Sheet are reporting increasingly negative surface mass balance trends (Veliconga et al., 2014; McMillan et al., 2016), possibly due to accelerated thinning from surface water drainage to the bed (Zwally, 2002). Contributions from the Greenland Ice Sheet to global mean sea level rise is larger than Antarctica and has increased from 0.09 mm yr^{-1} over 1992-2002 to 0.59 mm yr^{-1} over 2002-2011 (Vaughan et al., 2013), and has more recently reached 0.74 ± 0.14 mm yr^{-1} (McMillan et al., 2016). Paleo-constraints on inland ice sheets can provide valuable information about ice sheet thinning dynamics during periods of abrupt climate change, which can be used to improve models predicting future ice sheet decay (Hansen et al., 2015; Shakun et al., 2015; Winkelmann et al., 2015).

THE PROBLEM AND OUR APPROACH

Sea-level reconstructions reveal global ice volume variations through time, but provide little information on how that volume was partitioned among ice sheets (Clark et al., 2009). Although geophysical models attempt to invert isostatic rebound patterns into ice sheet thickness reconstructions, they exhibit considerable disagreement (Clark and Tarasov, 2014; Peltier et al., 2015). Similarly, numerical models attempt to simulate past ice sheet evolution in response to climate change (e.g., Abe-Ouchi et al., 2013; Gregoire et al., 2012), but they are subject to a wide variety of uncertainties such as the climate forcing and ice-sheet dynamics (Stokes et al., 2015). All such models ultimately require ground-truthing – but stronger geologic constraints are needed to advance this data-model dialogue, as highlighted by a recent community-wide workshop (Whitehouse and Tarasov, 2014).

A potential, albeit geographically-limited, solution to the problem of reconstructing ice sheet thickness through time is cosmogenic exposure dating along vertical transects, also known as ice sheet "dipsticks." This technique measures the build-up of cosmogenic nuclides in a series of glacial boulders and/or outcrops down a mountainside to determine when each was exposed to cosmic radiation as the ice sheet surface lowered during deglaciation (see commentary by Bierman, 2007). Glacial dipsticks have been instrumental in constraining the thinning history of ice sheets in Scandinavia (Brook et al., 1996; Goehring et al., 2008), Antarctica (Stone et al., 2003; Ackert et al., 2007; Mackintosh et al., 2007, 2011; Johnson et al., 2014), and Greenland (Corbett et al., 2011; Nelson et al., 2014).

Strikingly, other than our own recent study at Acadia National Park in Maine (Koester et al., 2017), no major glacial dipsticks have been measured for the Laurentide Ice Sheet (LIS), which was the largest body of ice at the Last Glacial Maximum (LGM), accounting for ~65-90 of the 130 m LGM sea level lowstand (Clark and Mix, 2002). There are only two regions where substantial topographic relief (>1000 m) was uncovered by LIS retreat during the last deglaciation – the mountains of New England and southern Quebec (Fig. 1), and much more remote parts of the eastern Canadian Arctic, including Baffin Island and northern Labrador. Samples from either of these regions could directly constrain the thinning history of the large, but now vanished LIS, although, for reasons related to ice sheet basal thermal conditions in the Canadian Arctic (Marsella et al., 2000; Corbett et al., 2016; Margreth et al. 2016), the mountains of New England are much more likely to provide accurate deglacial ages.

Figure 1. The locations of 12 ice-sheet mountain dipsticks (red dots) to constrain the thinning history of the southeastern LIS. As of July 2017, we have sampled along vertical transects from all sites except for the Catskills and Mt. Kearsarge. The smaller map shows the LIS outline at the LGM (yellow), and highlights the study region (white box). Note that this region contains the only large mountains underlying interior portions of the LIS, and thus provides a unique opportunity to reconstruct the vertical collapse of the ice sheet.

GLOBAL CLIMATE AND SEA LEVEL DURING THE LAST DEGLACIATION

The last deglaciation provides an outstanding opportunity to understand the complex interplay between ice sheets, ocean circulation, and climate. We provide below a summary of the last deglaciation to highlight the relevance of these questions and to flag uncertainties that our proposed research could help address.

The Oldest Dryas (19-14.6 ka)

The last deglaciation commenced with an abrupt 5-10 m sea-level rise over a few centuries at ~19 ka (Carlson and Clark, 2012), which has been associated with initial pullback of Northern Hemisphere ice sheets due to summer insolation forcing (Fig. 2a) (Clark et al., 2009), although emerging marine data suggest that the Antarctic Ice Sheet also began retreating at this time for unknown reasons (Weber et al., 2011, 2014). The resulting freshwater forcing to the North Atlantic may have weakened the AMOC causing a bipolar seesaw climate response with hyper-cold conditions centered around the North Atlantic and warming in the Southern Hemisphere (Fig. 2b,c) (He et al., 2013). Cold stadial conditions in the Northern Hemisphere and a sluggish AMOC persisted for the next four millennia of the Oldest Dryas interval (McManus et al., 2004; Shakun et al., 2012), but the Northern Hemisphere ice sheets continued retreating and the LIS underwent a major iceberg discharge episode during Heinrich event 1 at ~16 ka (Hemming, 2004). Significant ice loss during the cold Oldest Dryas may seem somewhat surprising. A possible explanation is that the Oldest Dryas was characterized by extreme seasonality, with cooling predominantly during winter as the weakened AMOC promoted sea-ice expansion while summers continued warming due to rising insolation and atmospheric CO_2 (Denton et al., 2005). Coral-based estimates of global sea-level rise during the Oldest Dryas range widely from 8 to 21 m, and the contribution of individual ice sheets is especially uncertain (Carlson and Clark, 2012). Of particular interest, sea-level rise seems to have outpaced LIS area retreat (Fig. 2a). Marshall et al. (2002) highlight a similar disparity, noting that, "*the isostatic record demands substantial ice thinning subsequent to LGM, at a time (14-20 ka) when there is no strong signal of ice sheet retreat (Dyke and Prest, 1987). Model results suggest that this is possible via the increasing role of fast basal flow in this period, as* *more of the Laurentide ice sheet becomes warm-based and basal melt water accumulates at the bed. This essentially argues for a transition from thicker, largely cold-based ice sheets at the LGM to a thin and mobile, more West Antarctic Ice Sheet-like ice sheet through the deglaciation.*" As a review on deglacial sea level by Carlson and Clark (2012) recently concluded though, "*the volume contributions of individual ice sheets to sea level change between 19.5 ka and 14.6 ka, which are required to specify freshwater fluxes and their entry points to the ocean, need to be better determined*." Determination of the volume contribution for the southeastern part of the LIS is a major goal of this project.

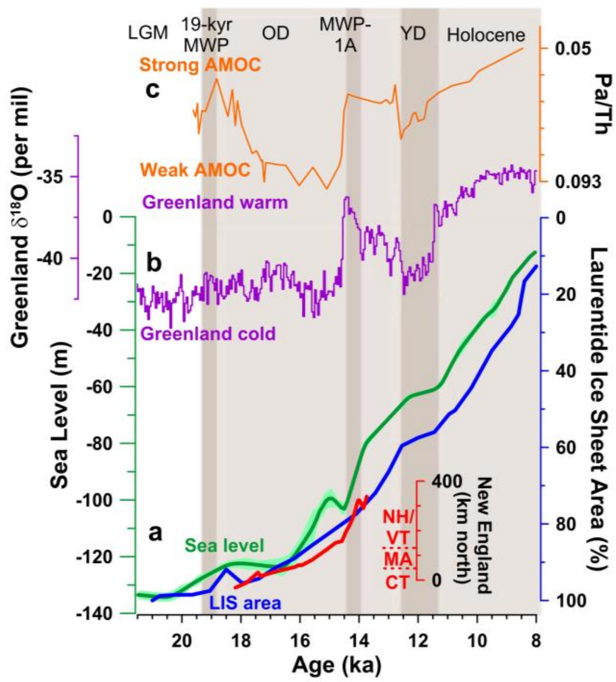

Figure 2. Deglacial ice melt, climate, and ocean circulation. (**a**) Global sea level (green) (Lambeck et al., 2014) LIS areal extent (blue) (Dyke, 2004), and northward LIS retreat in central New England based on varves (red) (Ridge et al., 2012). (**b**) Greenland $\delta^{18}O$, a proxy for temperature (NGRIP members, 2004). (**c**) Protactinium/thorium ratios in a North Atlantic sediment core, a proxy for AMOC strength (McManus et al., 2004). Note the differences between LIS extent and global sea level, the increased rate of LIS retreat in New England coincident with MWP-1A, and the general associations between changes in ice-sheet retreat/sea-level rise, ocean circulation strength, and temperature.

Meltwater Pulse 1a and Bølling warming (14.6-14.3 ka)

An abrupt warming of the Northern Hemisphere occurred at the onset of the Bølling interstadial at 14.6 ka as the AMOC resumed (Liu et al., 2009), and coincided with the largest jump in deglacial sea level –MWP-1A (Fig. 2). This 14-18-m sea-level rise occurred in no more than 350 years, implying rates of sea-level rise in excess of 40 mm/yr (Deschamps et al., 2012), or more than an order of magnitude faster than sea-level rise today (3 mm/yr; (Church and White, 2011). While MWP-1A was first assumed to have originated exclusively from the LIS (Fairbanks, 1989; Peltier, 1994), given its large size, sea-level fingerprinting and Southern Ocean marine evidence suggest a significant though uncertain Antarctic contribution (Weaver et al., 2003; Deschamps et al., 2012; Weber et al., 2014). Planktonic $\delta^{18}O$ runoff records from the Gulf of Mexico (Wickert et al., 2013), the Arctic (Carlson, 2009), and the Labrador Sea (Obbink et al., 2010) detect only minor contributions from various sectors of the LIS to MWP-1A. Furthermore, LIS areal retreat was no greater during MWP-1A than before or after the event (Fig. 2a). Therefore, any major LIS sea-level contributions could only have come from rapid ice sheet thinning. Just to provide a sense of scale, if MWP-1A were sourced evenly from across the entire LIS, it would lower the ice sheet surface by

~600 m, a thinning easily detectable using the dipstick method we are currently employing. Sourcing MWP-1A from only a part of the ice sheet would obviously increase this surface lowering estimate further. Gregoire et al. (2012) simulate a 9-m sea-level rise in 500 years in a numerical ice sheet model as the LIS and Cordilleran Ice Sheet separated due to saddle collapse and suggest that this process may account for MWP-1A, though ^{14}C ages suggest that these ice sheets actually separated well before MWP-1A (Clague and James, 2002; Dyke, 2004). The Eurasian Ice Sheet complex was much smaller than the LIS, and therefore a less likely candidate to explain MWP-1A. Adding to this puzzle, recent glacioisostatic modeling suggests that Antarctica only contained ~8 m sea-level equivalent of additional ice at the LGM (Whitehouse et al., 2012), limiting its potential contribution to MWP-1A, and available Antarctic dipsticks indicate only modest thinning (Mackintosh et al., 2007, 2011). An accounting of the sources of sea-level rise during this singular event (MWP-1A) is thus far from complete (Fig. 3).

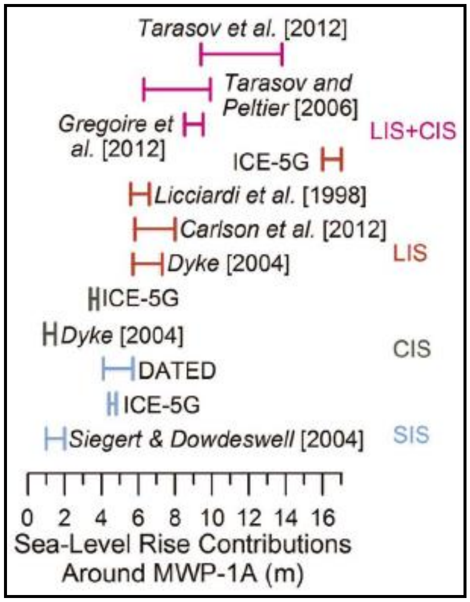

Figure 3. Estimates of ice-sheet contributions to MWP-1A. LIS = Laurentide Ice Sheet; CIS = Cordilleran Ice Sheet; SIS = Scandinavian Ice Sheet. (from Carlson and Clark, 2012).

The Allerød, Younger Dryas, and Holocene (14.3-6.5 ka)

Sea-level rise returned to pre-MWP-1A rates after 14.3 ka and sea level increased ~7-10 m during the Allerød period over the next millennium (Fig. 2a) (Edwards et al., 1993; Bard et al., 1996; Peltier and Fairbanks, 2006), likely dominated by Northern Hemisphere sources (Carlson and Clark, 2012). The abrupt Younger Dryas cold event in the Northern Hemisphere commenced 12.9 ka as the AMOC weakened again, perhaps due to southern LIS retreat into Canada and routing of freshwater runoff from the Mississippi to the St. Lawrence drainage (Broecker et al., 1989; Broecker, 2006). LIS retreat and sea-level rise slowed during the Younger Dryas, before again picking up pace as the AMOC resumed at the onset of the Holocene at 11.6 ka (Bard et al., 2010) and summer insolation reached a maximum. The Eurasian Ice Sheet disappeared by 10 ka (Boulton et al., 2001), and the remaining Holocene sea-level rise came from LIS retreat in Canada (Dyke, 2004; Carlson et al., 2007) and the Antarctic Ice Sheet.

The central message that emerges from this summary is that while the broad pattern of climate change, ocean circulation, and sea-level rise during the last deglaciation are reasonably well constrained, the contributions of individual ice sheets to these processes, which is critical to understanding the internal dynamics of the climate system, are not. In particular, well-documented ice margin retreat histories are not complemented by similarly strong vertical thinning constraints, and thus ice volume uncertainties remain substantial.

SOUTHEASTERN LAURENTIDE DEGLACIATION

With the global summary above as context, we detail here the deglaciation of the southeastern LIS, the focus of our current research. Central New England has one of the best-constrained ice margin histories in the world (Fig. 4), owing to considerable ^{14}C dating of lakes and bogs, and in particular, extensive varve sequences, which have been tied to the ^{14}C timescale and the Greenland ice core record (Ridge et al., 2004, 2012). Furthermore, ^{10}Be ages from the Connecticut and Champlain Valley lowlands have also been linked to these other chronometers, resulting in a precise regional production rate calibration (Balco et al., 2009). This well-dated margin record makes the region especially conducive to constraining ice volume changes and understanding ice sheet behavior, if, and only if, the thinning history can be well determined.

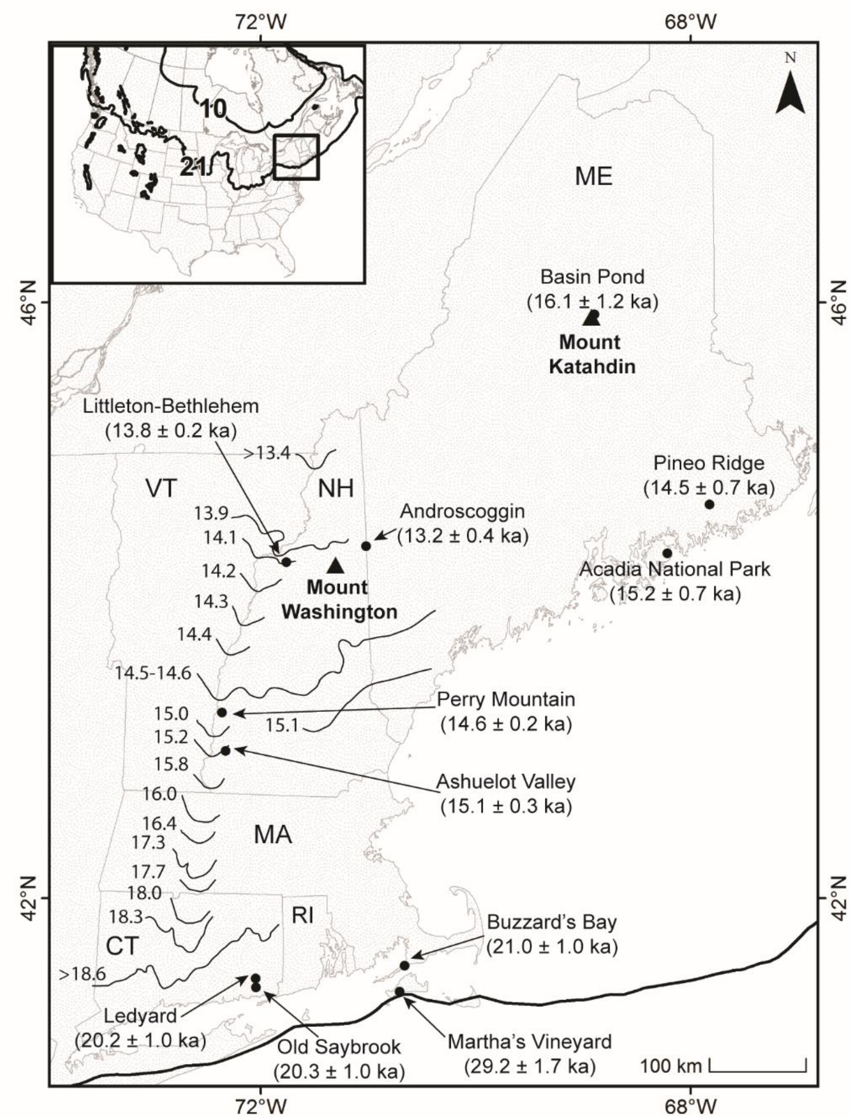

Figure 4. Laurentide Ice Sheet lateral extent through time. Isochrones show the North American Varve Chronology of deglaciation (from Fig. 11 in Ridge et al. (2012)). The black dots are moraines dated using ^{10}Be and calibrated with the northeastern North American production rate. (Martha's Vineyard and Buzzard's Bay – Balco et al., 2002; Old Saybrook and Ledyard – Balco and Schaefer, 2006; Ashuelot Valley, Perry Mountain and Littleton-Bethlehem – Balco et al., 2009; Androscoggin – Bromley et al., 2015; Basin Pond – Davis et al., 2015; Pineo Ridge and Acadia National Park – Koester et al., 2017). Inset figure shows the extent of the LIS at 21 ka and 10 ka (Dyke, 2003).

Ice Extent

The LGM and subsequent deglaciation chronology throughout the New England area has been well researched and includes minimum-limiting radiocarbon ages (e.g. Dyke et al., 2004), glacial varves (Ridge, 2004; Ridge et al., 2012), and cosmogenic nuclide dating (Balco et al., 2002; Balco and Schaefer, 2006; Bierman et al., 2015; Bromley et al., 2015; Davis et al., 2015; Corbett et al., 2017; Koester et al., 2017; Hall et al., 2017; Fig. 4). The LIS reached its maximum extent (Fig. 4) before ~25 ka as indicated by recalculated ^{10}Be exposure ages on terminal moraines from Martha's Vineyard, MA (27.5 ± 1.6 ka, Balco et al., 2002) and northern New Jersey (25.2 ± 1.3 ka, Corbett et al., 2017). After the LIS began pulling back from its terminal moraines, northward retreat was gradual through several millennia at ~50 m/year (Ridge et al., 2012), then increased after ~20 ka likely due to increased Northern Hemisphere insolation (Clark et al., 2009). There is little indication of Heinrich event 1 at the southern margin, except perhaps indirectly, with the modest Chicopee readvance in Massachusetts at 17.3 ka, possibly occurring in response to North Atlantic cooling (Ridge et al., 2012). Coastal Maine, mostly below the marine limit, rapidly deglaciated 15-16 ka (Davis et al., 2015; Koester et al., 2017; Hall et al., 2017), and the North Charlestown moraines were deposited across central New England just prior to the Bølling warming. Thereafter, and synchronous with MWP-1A, the rate of retreat increased dramatically to ~300 m/year (Fig. 2a, 4) (Ridge et al., 2012). Except for the Littleton-Bethlehem Readvance north of the White Mountains of New Hampshire at 13-14 ka (Balco et al., 2009; Ridge et al., 2012) (Fig. 2b), this rapid retreat continued until the LIS margin exited New England into southern Quebec during the late Allerød (Fig. 4) (Dyke, 2004).

The North American Varve chronology from the Connecticut River valley provides insight into the retreat of the LIS from Massachusetts to Vermont and indicates the retreat rate began between 50 and 100 m/yr (Ridge, 2004; Ridge et al., 2012). Thereafter, the retreat rate increased to ~300 m/yr during the Bølling Interstadial and passed Mt. Washington around ~14.2 ka (Ridge, 2004; Ridge et al., 2012; Fig. 4). Glacial varves were also thicker during the Bølling, implying more intense summer melt. In addition, a model of the LIS at the end of the Oldest Dryas (~15 ka) and during the Bølling-Allerød (~14.2 ka) found the surface balance increased during the Bølling (Carlson et al., 2012), supporting the varve thickness record. As the ice sheet retreated further inland the Pineo Ridge moraine complex in coastal Maine was abandoned around the Bølling Interstadial (14.5 ± 0.7 ka; Koester et al., 2017) followed by the Littleton-Bethlehem moraine, just north of the Presidential Range in northern New Hampshire (13.8 ± 0.7 ka, n = 4; Balco et al., 2009; Thompson et al., 2017), and the Androscoggin moraine in northeastern New Hampshire and western Maine (13.2 ± 0.7 ka, n = 7; Bromley et al., 2015) before retreating further north into Canada.

Ice Thickness

In contrast to the dozens of ^{10}Be ages, hundreds of organic ^{14}C ages, and thousands of varve counts constraining ice retreat at lower elevations in New England, the ice thickness history is largely uncertain. Existing ^{14}C ages at higher elevations are scant, they come almost exclusively from lake and bog basal sediments and are thus only minimum-limiting ages, and they are very noisy, spanning several millennia and showing no coherent trends with elevation (Table 1). It is thus not possible to say whether ice sheet thinning occurred predominantly during the cold Oldest Dryas or the warm Bølling/MWP-1A interval or the still-warm but slower-sea-level-rise Allerød interval. It is similarly an open question whether ice sheet drawdown was very rapid (centuries) or much more gradual (millennia).

Two recent studies from our team present initial cosmogenic ages on a small set of samples from Katahdin, ME, and Mt. Washington, NH, and raise intriguing questions in these regards. Samples from the Katahdin highlands have statistically indistinguishable deglaciation ages (n=6, 15.3±2.1 ka, 1σ) from boulders on the Basins Pond moraine halfway up the mountain (n=5, 16.1±1.2 ka, 1σ), as well as a lone boulder in the nearby lowlands (14.5±0.8 ka, 1σ) (Davis et al., 2015) (Fig. 5). These ages thus imply rapid ice surface lowering at ~16-15 ka, though more gradual thinning cannot be excluded due to the small sample number and age scatter, which may reflect measurement imprecision as AMS was done in the 1990s. In addition, ^{10}Be and ^{26}Al nuclide concentrations in several summit bedrock samples from both mountains are 2-10 times higher than would be expected due to ~15 kyr of postglacial exposure (and confirmed in corresponding *in situ* ^{14}C ages) (Fig. 5), suggesting that the summits were covered by non-erosive, cold-based ice at the LGM (Bierman et al., 2015). Also important, nuclides inherited from prior periods of exposure are found only in summit samples, while lower on the mountains and throughout the rest of New England, such inheritance is rare in boulders and bedrock. The lack of nuclides inherited from prior periods of exposure indicates that in New England ice was largely erosive (warm-based) and that the cosmogenic clock was reset during or after the LGM, as our initial project results at Acadia National Park in Maine demonstrate (Koester et al., 2017), although the relief there is only about 300 meters, substantially less than many other mountains in our study.

Site	Material	Elev (m)	^{14}C age	cal yr BP	Reference
Moosilauke					
Deer Lake bog	bulk conv.	1325	13,000±400	14,195-16,820	Spear (1989)
Mirror Lake	bulk/macros conv.	213	13,800±560	15,720-17,415	Davis and Davis (1980)
Franconia Notch					
Lonesome Lake	bulk conv.	831	10,535±495	11,065-13,355	Spear et al. (1994)
Profile Lake	wood conv.	593	10,660±40	12,772-12,885	Rogers (2003)
Mt. Washington					
Lakes of Clouds	bulk conv.	1538	11,530±165	13,200-13,500	Spear (1989)
Lost Pond	bulk conv.	625	12,870±370	14,700-16,000	Spear et al. (1994)

Table 1. Low versus high-elevation ^{14}C ages from the White Mountains, NH. Note that the age pairs tend to be fairly similar, suggesting that ice-sheet drawdown may have been rapid. On the other hand, these ages are only minimum-limiting, and the close correspondence in ages may reflect the timing of revegetation and the first occurrence of datable organic material. Indeed, this complication may explain why higher-elevation ages tend to be younger than lower-elevation ages, opposite the pattern expected from top-down deglaciation.

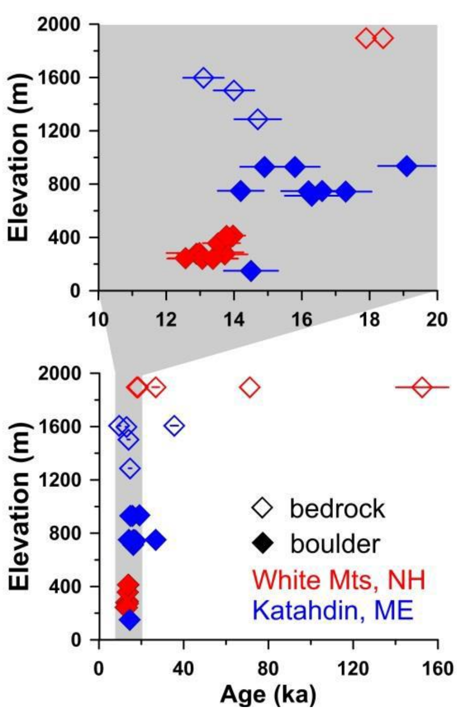

Figure 5. ^{10}Be ages from the Katahdin, ME, and Mt. Washington, NH, areas. Panel on top shows zoom-in of panel on bottom. Ages are from Balco et al. (2009), Davis et al. (2015), Bromley et al. (2015), and Bierman et al. (2015), and were determined using the CRONUS calculator using the northeast North America production rate (Balco et al., 2009). A few bedrock samples from the summits have older exposure ages than expected, indicating nuclides inherited from prior periods of exposure. The data are too sparse and do not include boulder ages higher on the mountains to provide robust constraints on thinning, but they highlight the potential of such work.

NORTHERN NEW ENGLAND SUMMITS

The Presidential Range is located in the White Mountains of New Hampshire and consists of 13 peaks ranging from 1,235 m (Mt. Jackson) to 1,917 m (Mt. Washington). The large cirque basins throughout the Presidential Range (i.e. Huntington Ravine, Tuckerman Ravine, and the Great Gulf on Mt. Washington) have been a topic of debate since they were first studied by J.W. Goldthwait (1913, 1916). J.W. Goldthwait concluded that the cirques were carved by alpine glaciers before the most recent continental glaciation and cited evidence that included till on the cirque floors from northern provenances, absence of end moraines on cirque floors, and asymmetric cirque cross-valley profiles. Despite the strong evidence, Antevs (1932) and Johnson (1917, 1933) opposed J.W. Goldthwait's conclusions about the sequence of cirque and ice sheet glaciation. R.P. Goldthwait (1970) later provided evidence from pebble lithologies in till on the uplands and in the north-facing cirques and concluded that the till in the Presidential Range was deposited by continental ice, supporting the idea that the cirques were carved before continental glaciation. Further, unlike Wagner (1970) in northern Vermont, Waitt and Davis (1988) and Davis (1999) found no evidence of cirque glaciation following continental ice overriding all mountainous areas of northern New England. However, Fowler et al. (2012) suggest that deposits at the mouth and along the sidewalls of the Great Gulf may provide evidence for alpine (or continental ice) in that cirque post-dating continental ice overriding the Presidential Range.

Of particular interest to our current work are the relatively level high-elevation areas known as "lawns" (ex. Bigelow Lawn, the Alpine Garden, and Monticello Lawn), which together make up a topographic feature called the Presidential Upland (Fig. 6) that represents an old Tertiary erosion surface formed during a prolonged lull in tectonic uplift (Goldthwait, 1940), whereas Thompson (1960a, 1960b, 1961) believed that the features were the result of freeze-thaw processes during the Quaternary. Eusden and Fowler (2013) sided with R.P. Goldthwait.

Figure 6. View looking north at Mt. Washington's summit cone from rôche moutonnée near the junction of the Camel Trail and Davis Path on the Bigelow Lawn, part of the old Tertiary erosion surface known as the Presidential Upland. Note the patterned ground with stone polygons in the foreground.

There is clear evidence from glacial erratics (Fig. 7), glacially molded surfaces, and the thinness of soils on summits that the LIS overrode New England mountains (Tarr, 1900; J.W. Goldthwait, 1916; R.P. Goldthwait, 1940; Davis, 1976, 1989; Fowler et al., 2013), as opposed to the mountains being centers of radial outflow of ice during the late Wisconsinan (Flint et al., 1942: Flint, 1951). However, the timing of initial LIS advance into New England during the last glaciation is poorly constrained. The glacial stratigraphic record of most of New England includes an upper till and a lower till. The upper till is part of a widespread drift sheet in New England and is interpreted to be late Wisconsinan based on minimum-limiting radiocarbon ages. The more weathered lower till deposits underlying late Wisconsinan (Marine Isotope Stage (MIS) 2) deposits is less common in New England, but occurrences have been described from the Boston Harbor drumlins, Massachusetts (Kaye, 1961), Nash Stream, New Hampshire (Koteff and Pessl, 1985), and New Sharon, Maine (Caldwell, 1986; Weddle, 1989), but its age remains unknown. The lower till's age was originally assigned to the early Wisconsinan (MIS 4) (Borns and Calkin, 1977; Stone and Borns, 1986; Vincent and Prest, 1987), but others have suggested the lower till is more likely Illinoian (MIS 6) or older due to radiometric dating and amino acid racemization age estimates on detrital coral from the upper part of the Sankaty Head on Nantucket Island, MA (Oldale, 1982; Oldale and Colman, 1992). In addition, ice volume estimates from oxygen isotope and sea level records indicate ice was less extensive during the early Wisconsinan than the late Wisconsinan, possibly suggesting that the LIS did not extend as far south as New England then (Oldale and Colman, 1992; Lambeck et al., 2014).

Figure 7. View looking south at Mt. Washington from about 30 m below summit of Mt. Jefferson with two large, sub-rounded, granitic erratics in foreground (rucksack and trekking pole for scale). Angular and sub-angular blocks are frost-riven from the schist bedrock, common on all of the summit cones of the northern Presidential Range. Great Gulf cirque headwall and Mt. Clay on north side and Mt. Monroe to southwest of Mt. Washington.

Deglacial constraints on LIS retreat at higher elevations in New England include minimum-limiting basal radiocarbon ages from alpine lakes (Table 1) and a few cosmogenic nuclide exposure ages on Katahdin and Mt. Washington. For instance, a basal bulk radiocarbon age from the lower Lakes of the Clouds (1,534 m), just below the summit cone of Mt. Washington, dates to 13.35 ± 0.2 cal. ka (Table 1; 11,530 ± 165 ^{14}C yrs BP; I-10684; Spear, 1989, Cwynar et al., 2001), although this age probably substantially post-dates deglaciation since the ice margin had already retreated to the Canadian border by this time (Ridge et al., 2012). Radiocarbon ages from alpine lakes on Katahdin are also several thousand years younger than continental ice retreat, perhaps due to the lag in vegetation colonization following deglaciation (Davis and Davis, 1980). Davis et al. (2015) measured surface exposure ages at multiple sites from the top to base of Katahdin and concluded that the LIS thinned rapidly between 16 and 15 ka in central Maine. Cosmogenic nuclide exposure ages from the summits of Katahdin and Mt. Washington were 2-10 times higher than expected indicating that cold-based ice likely covered the summits of New England mountains (Bierman et al., 2015).

MOTIVATING QUESTIONS

The background given in the previous two sections highlights several long-standing problems related to ice sheet deglaciation, sea-level rise, climate, and ocean circulation to which we can contribute greatly through the dipstick approach. Our research specifically addresses the following five questions concerning the southeastern LIS.

1) Did the LIS thin during the Oldest Dryas cold interval, or specifically during Heinrich event 1?

Such thinning might suggest that Oldest Dryas cooling was mostly during winter and related to sea-ice expansion (Denton et al., 2005), or that Heinrich 1 dynamical discharge caused significant ice-sheet drawdown over the southeastern LIS (Shaw et al., 2006).

2) Did the LIS thin synchronously with MWP-1A and Bølling warming?

The southern LIS would have presumably been one of the most vulnerable ice masses in the world to ablate during the Bølling given its southerly location and likely sensitivity to climate change in the nearby North Atlantic. The Midwest ice lobes, however, were thin due to underlying deformable beds (Clark, 1992) and contributed little to sea-level rise based on Gulf of Mexico runoff records (Wickert et al., 2013). The North American varve record indicates that ice margin retreat there increased dramatically during MWP-1A, perhaps implying a significant contribution, though this margin retreat was no faster than during the following thousand years after the sea-level event ended (Fig. 2a, 4). There are currently no ice sheet dipsticks anywhere in the world that date pronounced thinning to the MWP-1A interval, although scatter in vertical data from Norway permit this possibility (Goehring et al., 2008).

3) How fast did the ice surface lower?

Rapid drawdown might point to active ice dynamics, such as meltwater delivery to the bed and increased sliding, and would provide constraints on how fast a vulnerable subpolar ice sheet can collapse in a warming climate, perhaps broadly analogous to the southern Greenland Ice Sheet today. Gradual thinning would imply weaker climate forcing and/or less dynamic discharge.

4) How did southeastern LIS melt relate to changes in the AMOC?

In addition to the general correspondence between New England LIS margin retreat and North Atlantic temperature evolution (Fig. 2a,b), New England varve thickness records also suggest a <u>direct correlation</u> between summer melt of the LIS and Greenland temperature (Ridge et al., 2012). On the other hand, a $\delta^{18}O_{sw}$ record from the nearby Laurentian Fan suggests that southeastern LIS meltwater production was <u>inversely related</u> to North Atlantic temperature, being higher during the Oldest Dryas and three cold intervals within the Bølling/Allerød than when the North Atlantic was warmer (Obbink et al., 2010). Likewise, mass-balance modeling suggests that the impact of Bølling warming on the southeastern LIS may have been offset by an associated increase in precipitation (Carlson et al., 2012). These conflicting views highlight the causal uncertainty in how southeastern LIS evolution and North Atlantic climate are linked – for instance, to what extent does LIS melt cause North Atlantic cooling through freshwater forcing of the AMOC versus North Atlantic warming causes LIS melt?

5) How well do ice sheet models simulate LIS deglaciation?

Ultimately, one of the central goals of Earth science is to accurately model the Earth system, and this enterprise hinges crucially on ground truthing models (Stokes et al., 2015). While several models of ice-sheet retreat exist, based on either isostatic inversion techniques or glaciological modeling, the models are still poorly constrained by geologic data and exhibit substantial differences, such as estimates of the LIS contribution to MWP-1A (Fig. 3). These model discrepancies are apparent for the southeastern LIS, with pronounced differences in ice thickness at the LGM, the ice-sheet profile, and the timing and rate of thinning through the deglaciation (Fig. 8). Our data provide a novel test for these models and serve as a target for future modeling efforts.

RESEARCH STRATEGY

Sampling

We constrain the timing and rate of southeastern LIS thinning by measuring cosmogenic nuclide dipsticks on a dozen of the highest peaks in New England and southern Quebec (Fig. 1). We target boulders as much as possible, given that they are less likely to contain inheritance from prior periods of exposure (Hallet and Putkonen, 1994; Putkonen and Swanson, 2003; Putkonen and O'Neal, 2006; Balco, 2011; Heyman et al., 2011); bedrock is only sampled when boulders are not available because nuclides created during prior periods of exposure can be preserved in bedrock beneath cold-based ice (e.g., Bierman et al., 1999, 2015; Colgan et al., 2002; Briner et al., 2003; Goehring et al., 2008). Our goal is production of a 3D-model of deglaciation of the mountains of the northern New England area to assess ice volume changes over time.

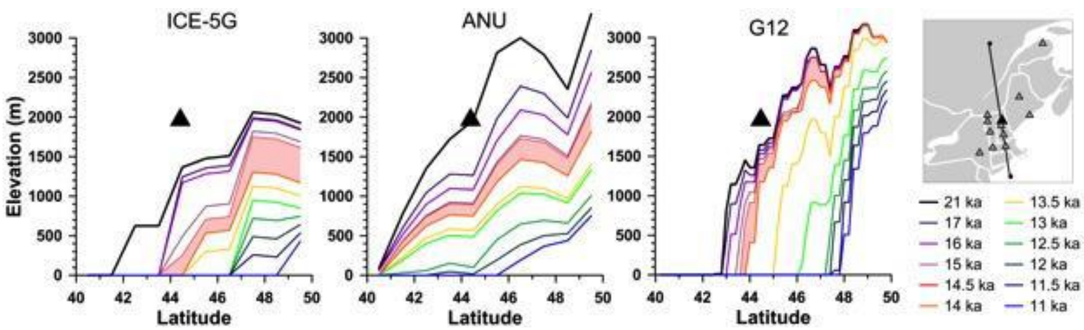

Figure 8. Ice sheet models of deglaciation. Southeastern LIS profiles along 71°W (the longitude of Mt. Washington, shown as the black line in the inset map) during various time steps of the last deglaciation (see legend for ages) based on the ICE-5G model (Peltier, 2004), a model from the Australian National University (ANU; Lambeck et al., 2002), and a dynamical model of the North American ice sheet system (G12; Gregoire et al., 2012). The summit location of Mt. Washington is represented by the solid black triangle. Simulated ice thinning during the Bølling warm interval (14.5-14.0 ka) is highlighted in pink on each panel. Note the dramatic differences in ice sheet thicknesses at 21 ka, ice sheet profiles, and timing and rates of thinning. Our dipsticks are well-positioned to distinguish between these, and other, models of deglaciation.

Sample sites (Fig. 1) include the highest points along the Green Mountains of Vermont (Jay, 1177 m; Mansfield, 1330m; Killington, 1289 m); high peaks of New Hampshire's White Mountains (Washington, 1917 m; Lafayette, 1600 m; Kearsarge, 895 m; (we note that Joe Licciardi and one of his graduate students at UNH exposure dated the highest mountain in southern New Hampshire, Mt. Monadnock (Hodgdon, 2016)); the highest peaks in Maine (Katahdin, 1606 m; recently added Mt. Bigelow, 1247 m; and Cadillac, 466 m, on the coast); western (Greylock, 1064 m) and eastern (Wachusett, 611 m) Massachusetts high points; the Catskill Mountains of New York (up to 1277 m); and the tallest mountain in southeastern Canada (Jacques-Cartier, 1268 m). These mountains were chosen because they: (i) provide the maximum relief available and so were exposed to much of the ice sheet thickness, (ii) span 7° of latitude and 9° of longitude, (iii) are composed mostly of quartz-bearing rocks (granite, schist, quartzite) ideal for cosmogenic exposure dating, (iv) will be relatively inexpensive to study since they are easily accessible and close to our home institutions, and (v) the six northern mountains were all within the LIS margin at the time of MWP-1A (Fig. 4).

Nuclides

In situ ^{10}Be and ^{14}C are primarily formed by cosmic ray spallation of oxygen in quartz-bearing rock and soil surfaces where nuclide concentrations build up over time (Gosse and Phillips, 2001). Cosmic ray flux attenuates within a few meters of surface, but the highest concentration is at the surface due to neutron attenuation with depth (Gosse and Phillips, 2001). In glacierized areas, erosive, warm-based ice typically erodes many meters into the underlying bedrock to remove nuclides from prior periods of exposure. The concentration of cosmogenic nuclides can be converted into exposure ages with a production rate to measure when ice retreated and exposed the area. However, non-erosive, cold-based ice that is frozen to the bed can leave behind nuclides inherited from previous exposure periods leading to older than expected ages (Gosse and Phillips, 2001). On the other hand, post-glacial cover by snow or soil can shield the surface from cosmic rays and lead to an artificially young exposure age (Schildgen et al., 2005). A regional ^{10}Be production rate has been calibrated for northeast North American from independently dated moraines within New England reducing our uncertainty on exposure ages (Balco et al., 2009).

Based on our initial cosmogenic exposure ages on bedrock from the summits of Katahdin and Mt. Washington (Fig. 5; Bierman et al., 2015; Davis et al., 2015), we should not have been surprised that a few ^{10}Be ages from higher summits are older than expected considering the existing varve and radiocarbon age control (Fig. 4, Table 1). Similarly, existing data sets (Balco et al., 2002; Balco and Schaefer, 2006; Davis et al., 2015) suggest that a few boulders carry inherited nuclides. For samples with higher than expected ages, we measure *in situ* ^{14}C (Bierman et al., 2015), which removes the confounding variable of inheritance from exposure prior to the LGM (with its short half-life, most ^{14}C produced during prior interglacials decays away during 20-30 ky of burial by ice (Lifton et al., 2001).

On the basis of existing paired ^{26}Al/^{10}Be data for Katahdin and Mt. Washington (e.g., Davis et al., 2015; Bierman et al., 2015), which provide statistically similar exposure ages for both nuclides, we believe that measuring ^{26}Al in dipstick samples likely does not provide useful additional information about sample history. Although ^{26}Al measurements can be useful in other situations, measuring ^{26}Al is redundant in New England samples because during a 100 ky glacial cycle, total exposure is several times longer than burial (~80 ky of exposure, 20 ky of burial); therefore, interglacial exposure quickly raises ^{26}Al/^{10}Be ratios to production values following each short period of burial by ice. As a result, ^{26}Al analysis mirrors ^{10}Be ages, thus funds are better spent measuring *in situ* ^{14}C or more samples for ^{10}Be.

Thinning rates

We calculate thinning rates following Johnson et al.'s (2014) approach, who measured Holocene cosmogenic dipsticks at Pine Island Glacier, Antarctica. This approach involves fitting error-weighted least-squares regressions through the dipstick profiles. Uncertainties are quantified through Monte Carlo simulations, in which the cosmogenic ages are allowed to randomly vary within their Gaussian uncertainties and the regression is recalculated (Fig. 9). Negative or zero slopes are rejected as physically untenable.

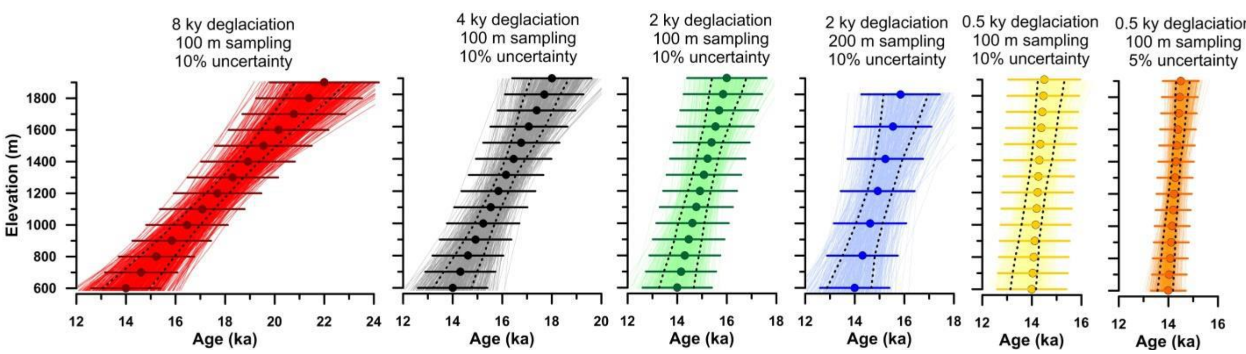

Figure 9. Synthetic dipsticks for Mt. Washington, NH, testing sensitivity to deglaciation duration, sampling density, and ^{10}Be geologic uncertainty typical for cosmogenic datasets. The thin colored lines behind each dipstick show 500 Monte Carlo-generated regressions in which ^{10}Be ages were allowed to randomly vary within their Gaussian uncertainties, and dotted lines give 68% confidence interval.

A New England area grand synthesis

The data we generate will be incorporated into a comprehensive database detailing all extant chronological data from the region – radiocarbon ages (e.g., Dyke, 2004), cosmogenic exposure ages (e.g., Balco et al., 2002; Balco and Schaefer, 2006; Balco et al., 2009; Bromley et al., 2015; Davis et al., 2015; Bierman et al., 2015; Koester et al., 2017; Hall et al., 2017), and varve constraints (Ridge et al., 2012). We envision this reconstruction representing the culmination of decades of glacial geologic work in this data-rich region, with ice volume calculations now possible using the vertical constraints that our cosmogenic exposure age data will provide. This reconstruction will then be compared to offshore marine records, climate records (Fig. 2b), and models (Fig. 8) to understand the nature of southeastern LIS deglaciation, and infer its possible causes and consequences. Our reconstruction will allow us to estimate New England ice volume losses, which together with a recent reconstruction of the south-central LIS's contribution to deglacial sea-level rise based on ice-sheet models and Gulf of Mexico runoff records (5.5 ± 2.1 m; Wickert et al., 2013), will better constrain the sea-level contribution history of the entire southern part of the LIS.

INITIAL RESULTS

Twenty ^{10}Be ages from boulders and bedrock along a vertical transect on the east side of Mt. Washington, the highest peak in New England (1917 m), constrain the timing and rate of LIS thinning during the last deglaciation (Fig. 10). Also, six new *in situ* ^{14}C ages for bedrock and boulders from the upper reaches of the mountain provide additional age constraints on deglaciation. With our data, we also seek to better explain the distinctive topography of the Presidential Range in New Hampshire, specifically the relatively level landscape at higher elevations versus deeply incised mountainsides below.

Our ^{10}Be exposure ages range from 12.5 ± 0.6 to 81.6 ± 4.5 ka (Fig. 10), and show a strong ordering with elevation; they are similar and agree with the ~14 ka timing of regional deglaciation (except for MW-13 (34.3 ±. 0.6 ka)) up to ~1,600 m asl, but then curve to increasingly older ages toward the summit, reaching values that are 3-6 times higher than the regional deglaciation age (Fig.11). These anomalously old ages suggest that there was minimal glacial erosion higher on the mountain, consistent with a transition from warm- to cold-based ice at about 5300 ft (1,600 m) elevation, which accords with the gradual topography of the Presidential "lawns" above this elevation.

The ^{10}Be ages between 1520 and 730 m a.s.l. are indistinguishable from one another at 1σ, and have a mean exposure age of 15.1 ± 0.8 ka (n = 7; 1SD). The two ^{10}Be exposure ages on boulders from Pinkham Notch at ~670 m a.s.l. (13.0 ± 0.4, 12.7 ± 0.2 ka) are substantially younger than the 15.35 ± 0.6 cal ^{14}C age (12,870 ± 370 ^{14}C yrs BP, Spear et el., 1994) from near-basal organic sediments in nearby Lost Pond (Table 1), slightly younger than the well-dated Androscoggin moraine north of the Presidential Range (13.2 ± 0.4 ka; Bromley et al., 2015), and much younger than suggested by ice retreat in the Connecticut Valley varve chronology to the west (14.1 ka; Ridge et al., 2012). Thus, we consider the two exposure ages on boulders downslope from Square Ledge in Pinkham Notch to be outliers, perhaps due to the boulders falling into place following deglaciation; therefore, we consider the mean Androscoggin moraine exposure age to be a better constraint for the base of our glacial dipstick.

DISCUSSION

Comparison of our exposure dating results from Mt. Washington suggest that continental ice in the area lowered rapidly during the Bølling-Allerød, which accords well with an increase in ice margin retreat rates in the Connecticut River valley to the west based on the North American varve chronology and with the NGRIP δ^{18}O ice core record from Greenland (Fig. 12). Our Mt. Washington glacial dipstick with its 1100 meters of relief provides a better opportunity to assess ice surface lowering than did our 300-meters of elevation range at Acadia National Park in Maine (Koester et al., 2017).

Although none will have quite as much relief, we are developing several other glacial dipsticks from mountains in the New England area, including the Chic Choc Mountains of Quebec (Fig. 1), for comparison with our Mt. Washington and Acadia National Park studies.

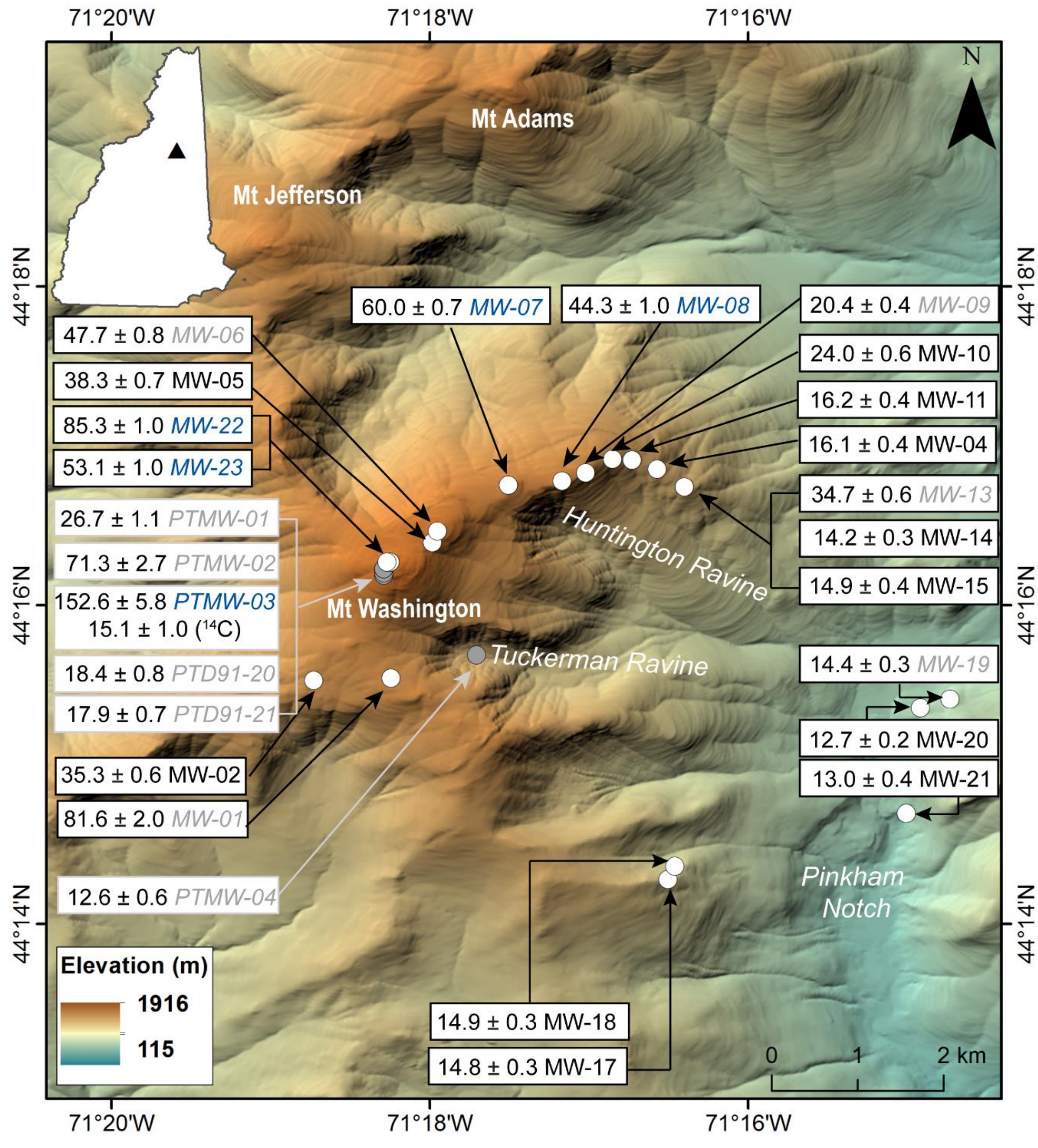

Figure 10. A LiDAR digital elevation model of Mt. Washington showing ^{10}Be ages (white circles; black outlined boxes) with 1σ internal uncertainties (in ka). Bierman et al. (2015) uncertainty-weighted ^{10}Be-^{26}Al ages are shown in boxes without outlines. Bedrock ages are grey italicized, boulder ages are black, and frost-riven block ages are blue italicized. One *in situ* ^{14}C age for a summit frost-riven bedrock sample is shown, but six new *in situ* ^{14}C ages along the elevational transect are not. The inset shows the location of Mt. Washington in New Hampshire.

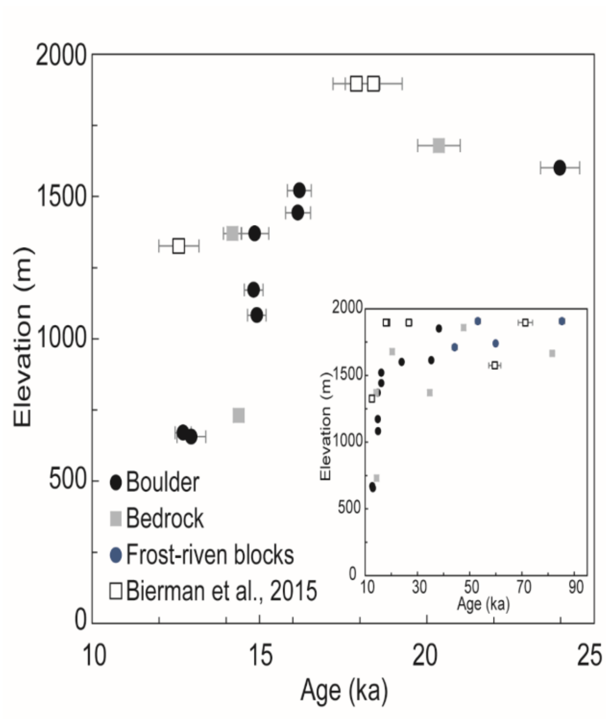

Figure 11. ^{10}Be ages of frost-riven blocks (blue circles), boulders (black circles), and bedrock (grey squares) taken from Mt. Washington, with horizontal bars showing internal error. The inset diagram shows all of the data, including samples with inherited ^{10}Be nuclides.

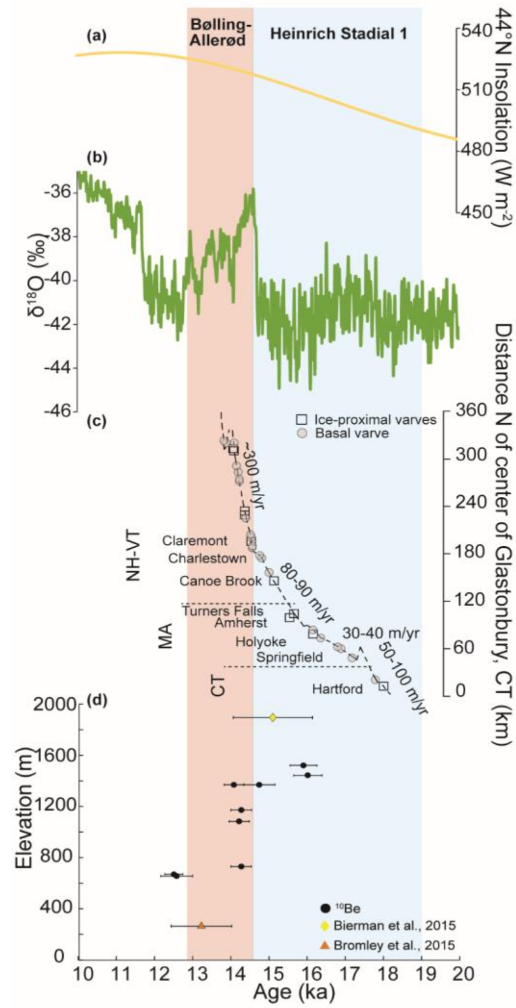

Figure 12. Paleoclimate records from the Northern Hemisphere compared to our ^{10}Be ages from Mt. Washington. (a) 44° N June insolation curve (Laskar et al., 2004), (b) Greenland ice core δ^{18}O (NGRIP dating group, 2006), a proxy for North Atlantic temperature, (c) The NEVC modified from Ridge et al., 2012 (d) ^{10}Be dipstick ages from Mt. Washington with ^{14}C summit age (Bierman et al., 2015), and Androscoggin moraine (Bromley et al., 2015).

We are also sampling bedrock from 20 additional summits in the White Mountains as a check for ^{10}Be inheritance to determine the elevational and spatial distribution of non-erosive, cold-based continental ice. Although ^{10}Be inheritance is common in polar landscapes (Bierman et al., 1999, 2014, 2016; Davis et al., 1999, 2006, Marsella et al., 2000; Briner et al., 2006; Miller et al., 2006; Corbett et al., 2013, 2016; Margreth et al., 2016), our study is one of the first to suggest that non-erosive, cold-based ice sheets are a factor to be considered in temperate mountainous regions. As suggested in Bierman et al. (2015), variable glacial erosion rates between summits and valleys may play a strong role in development and maintenance of northern Appalachian topography through the Quaternary.

Bierman et al. (2015) also compared exposure ages from the summits of Katahdin and Mt. Washington to a global sea-level of ice volume record (Lambeck et al., 2014), as shown here in reduced form (Fig. 13). Our 20 new ^{10}Be exposure ages also include several from the summit area of Mt. Washington that require multiple exposure periods prior to the LGM. However, recent work suggests that the St. Lawrence Lowland was free of Laurentide ice before about 31 ka (Parent and Dubé-Loubert, 2017), which along with the post-glacial incursion of the Champlain Sea as early as 14 ka (Lamothe, 1989; Parent and Occhietti, 1989, 1999), leaves less time for decay of *in situ* ^{14}C in rock on the Mt. Washington summit areas created prior to overrunning by continental ice. Perhaps a local Appalachian ice sheet or ice cap covered the mountainous areas of northern New England during parts of the late Wisconsinan, an idea invoked long ago by Flint (1951). However, the moraine record in the lowlands adjacent to the Presidential Range to the north and the North American varve record from glacial Lake Hitchcock to the west suggest continental ice recession toward the north during deglaciation (Thompson et al., 1999, 2017; Ridge; 2004; Ridge et al., 2012; Bromley et al., 2015).

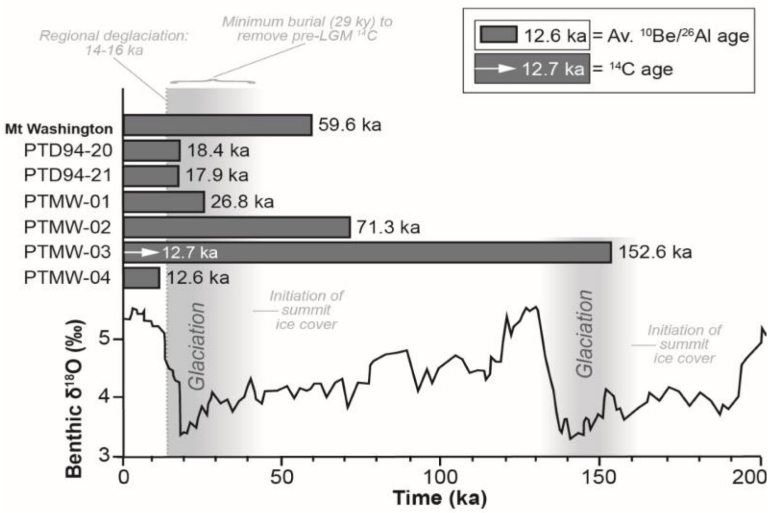

Figure 13. Schematic history of exposure samples from Mt. Washington reported in Bierman et al. (2015). Benthic ^{18}O record proxy for global ice volume (Lambeck et al., 2014). Dark bars are uncertainty-weighted average (^{10}Be, ^{26}Al) exposure ages. Twenty new ^{10}Be exposure ages reported in this chapter show a similar distribution (see Fig. 11). White arrow: one *in-situ* ^{14}C exposure age; six others are being analyzed. Gray shaded area represents five half-lives of ^{14}C (~29 k.y.) required to decay ^{14}C created prior to overrunning by Laurentide Ice Sheet. Regional deglacial age (16–14 ka) is shown by dotted line. LGM: Last Glacial Maximum.

ACKNOWLEDGMENTS

We thank the Mount Washington Auto Road and the Mount Washington Observatory, who have graciously agreed to assist us with this field trip. We also thank Brian Higginbotham, who assisted in the field, along with Chris Halsted and Anthony Vickers, who are now involved with the project as graduate students at Boston College.

ROAD AND TRAIL LOGS

Time, Place, Logistics

START TIME AND LOCATION: Trip begins on Sunday, October 1st, 7:30 AM, in the gravel parking area on the west side of NH Rte 16 to the immediate south of the Auto Road entrance (322363.00 m E, 4906302.00 m N). The base of the Auto Road is about 30 miles west of Bethel, Maine, and takes about 40 minutes to drive. From Bethel follow U.S. Rte 2 west to Gorham, NH, and then take NH Rte 16 south to the Auto Road entrance.

DESCRIPTION: This field trip will include a drive up Mount Washington's Auto Road to the summit for examination of sample sites for cosmogenic nuclide exposure dating that are part of a NSF-funded research project to construct glacial dipsticks for the deglaciation history of northeastern United States and surrounding areas. Depending on the weather, those interested are invited to hike down the mountain about 7 km (about 4.5 miles) and 1200 vertical meters (about 4000 vertical feet) via the Nelson Crag Trail to visit additional sampling sites to those on the summit and along the Auto Road. The Nelson Crag Trail is steep and difficult, especially if wet, so those wishing to participate in this part of the field trip must be prepared for adversity. There also will be an opportunity for a guided tour of Mount Washington Observatory on the summit, where hot lunches and drinks also may be purchased. We will drive as a caravan to the summit, making two or three stops on the way up to observe sampling sites and the alpine landscape, including the Great Gulf cirque and the northern peaks of the Presidential Range. For those not hiking down, the field trip will end by 2 pm, allowing additional time for the drive home. For those hiking down, the field trip should end by about 5 pm.

Warning: Due to the fragile nature of the alpine ecosystem, please always walk on trails or rocks. Expect the possibility of extremely cold and unpredictable weather. Be prepared with proper clothing and good hiking boots for very rocky, uneven terrain, winter–like conditions, and perhaps extremely high winds. There are no bathroom facilities for those electing to hike down the Nelson Crag Trail. Vehicles must be consolidated.

Road Mileage

0.0 Begin at Mount Washington Auto Road gate, opposite NH Rte 16 from Glen House.

4.1 **STOP 1.** Park in small lot on the east side of the road just above the switchback and 4000 ft elevation post, and carefully walk across road for views of the Great Gulf cirque; the northern Presidential peaks, right to left, Madison, Adams, Jefferson; and the hanging cirques Madison, Gulf, Jefferson Ravine, and Sphinx Basin. Cosmogenic exposure age samples were not collected here.

5.3 **STOP 2.** Park in the vicinity of Cragway Spring at about 4800 ft elevation on the outside of the right hairpin turn and just above on the west side of the road. Be really careful crossing the road here! Both bedrock and boulders were collected near here for cosmogenic exposure dating. The bedrock exhibits beautiful glacial polish quartz veins and blebs, but unfortunately our one bedrock sample for exposure dating consisted of mostly feldspar rather than quartz. Those hiking down the Nelson Crag Trail in the afternoon will have the opportunity to examine boulders sampled downslope to the east from here.

6.5 **STOP 3.** Park in the large lot on the northwest side of the road at about 5700 ft elevation in an area known as the "Cow Pasture." This elevation lies above the "lawns" in the Presidential Range, which are believed to be part of a pre-Quaternary surface known as the Presidential Upland, formed during prolonged periods of fluvial erosion beginning about 60 million years ago. The landscape above this elevation, including the cone of Mt. Washington (known at "The Rock Pile"), was not eroded by streams during the Tertiary or ice sheets during the Quaternary to the same degree as the Presidential Upland. A short 0.1-mile walk east to the junction of the Nelson Crag and Huntington Ravine Trails will allow examination of some large frost-riven blocks that were sampled for cosmogenic exposure dating.

8.0 **STOP 4.** Park in the lower summit parking lot at about 6200 ft elevation. A short walk up wood steps brings one to the summit complex, including the Mount Washington State Park building, which provides rest rooms, a cafeteria, and an optional tour of the Mount Washington Observatory, which has maintained continuous weather records on the summit since 1932. The summit sign at 6288 ft (1917 m) elevation marks the most prominent peak east of the Mississippi River. A few meters northwest from the summit sign, just north of the Tip Top House (a historic hotel, now museum, built in 1853 and the oldest surviving building on the summit), lie several frost-riven blocks of bedrock that we sampled for surface exposure dating. A short 0.1-mile walk southwest from the summit sign to "Goofer Point," where several bedrock samples were collected for surface exposure dating, provides a fine view of Lakes of the Clouds, the adjacent A.M.C. hut with the same name, and Mt. Monroe behind, with its striking stoss-lee glacial erosional topography (gentle side to N20W, steep side facing S20E).

For those not hiking down the Nelson Crag Trail, the summit marks the end of trip. Please drive down the Auto Road carefully!

Trail Mileage for hike down Nelson Crag Trail

0.0 Trail log from the lower summit parking lot for hike down Nelson Crag Trail.

1.0 **STOP 1.** Examine boulders sampled for exposure age along the Nelson Crag Trail where it lies about 0.1 mile east of the 6.5-mile mark at the "Cow Pasture" on the Auto Road, same area as Stop 3 on the drive up.

1.8 **STOP 2.** Examine boulders sampled for exposure age along Nelson Crag Trail near the bump known as Nelson Crag.

2.9 **STOP 3.** Examine boulders and bedrock sampled for exposure age along Nelson Crag Trail just below Cragway turn on Auto Road, same area as Stop 2 on the drive up.

4.6 **STOP 4.** Continue down Nelson Crag Trail to intersection with Old Jackson Road (trail) near 2-mile mark on Auto Road and Lowe's Bald Spot. Unfortunately, we were not able to collect exposure age samples near here, but collected samples from similar elevations from Square Ledge and the Glen Boulder Trail in Pinkham Notch.

6.5 **STOP 5.** Follow Old Jackson Road south to A.M.C. Pinkham Notch Camp where vehicle shuttles will return hikers to the base of the Auto Road, about 3 miles north on NH Rte. 16.

End of trip; thank you

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