MOUNT GREYLOCK AS A COSMOGENIC NUCLIDE DIPSTICK TO DETERMINE THE TIMING AND RATE OF SOUTHEASTERN LAURENTIDE ICE SHEET THINNING

By

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INTRODUCTION

This field trip to Mt. Greylock serves several purposes: 1) discuss and review the glacial history of Mt. Greylock and the surrounding region, 2) describe the motivation behind an NSF-funded project to understand the timing and rate of ice thinning in New England and adjacent areas, 3) explain the ‘dipstick approach’, the primary method being used to reconstruct ice thinning, and demonstrate the field work involved, 4) examine field sampling sites from this project and other glacial features on Mt. Greylock, and 5) discuss implications of initial results for paleoclimate and paleo-sea level change during the last deglaciation. Those interested can participate in a walk along the Appalachian Trail of approximately 2.7 miles (although trip participants are encouraged to lengthen their walk by ~2 miles and return back to the Greylock summit via the AT if desired).

Before beginning this field trip, it is important to reflect on why the fields of glacial geomorphology and ice-sheet reconstructions are areas of interest. Understanding ice sheets and their interactions with other components of the climate is essential to climate change mitigation and adaption discussions, but present projections of future ice sheet stability contain substantial uncertainties (e.g., Stocker et al., 2014; DeConto and Pollard, 2016, Hansen et al., 2016). These large uncertainties are due primarily to a limited understanding and poor model representation of ice sheet decay mechanics, particularly marginal processes such as ice-stream flow, ice-shelf buttressing, and subglacial meltwater drainage (Kleman and Applegate, 2014; Stokes, 2017). Modern observations of ice sheets in Greenland and Antarctica capture only a brief window in their history, and important basal processes are nearly impossible to observe. Reconstructing and studying the behavior of ice sheets during past periods of climate change offers a potential solution to observational and computational limitations.

The largest Northern Hemisphere ice sheet at the Last Glacial Maximum (LGM), the Laurentide Ice Sheet (LIS), is one of the most-investigated paleo-ice sheets, as its size and mixture of terrestrial and marine margins provide a unique opportunity to study the long-term behavior of ice margins analogous to modern-day locations in Greenland and Antarctica (Kleman and Applegate, 2014; Margold et al., 2015; Stokes et al., 2016; Stokes, 2017). Additionally, the near-complete retreat of such a large ice mass in only ~15-kyr illustrates the sensitivity of ice sheets to changing climatic conditions (Abe-Ouchi et al., 2013; Ullman et al., 2015).

The nature of interactions between the LIS and the rest of the climate system during the last deglaciation (~21-11 ka) are the source of much debate in the paleoclimate community. The last deglaciation was marked by a series of abrupt climate change events that are particularly apparent in proxy records from the North Atlantic and Greenland (e.g. Clark et al., 1999; Thornalley et al., 2010). A mechanism for similar abrupt events observed further back in the paleoclimate record (Dansgaard et al., 1993; Johnsen et al., 2001) was originally proposed by Broecker et al. (1985). They argued that changes in the strength of the Atlantic Meridional Overturning Circulation (AMOC) caused the North Atlantic to rapidly shift between two quasi-stable modes of climate. When the AMOC was strong, North Atlantic average annual temperatures were high (interstadial), and when the AMOC was weak, annual North Atlantic temperatures dropped (stadial). Clark et al. (1996) noted that a similar mechanism may have caused the abrupt warming and cooling events observed during the last deglaciation, and eventually hypothesized that the direction and volume of the LIS meltwater flux may have had a direct influence on the strength of the AMOC (Clark et al., 2001).

THE PROBLEM AND OUR APPROACH

While proxy evidence has emerged to support the notion of an AMOC with rapid fluctuations in strength during the last deglaciation (e.g., McManus et al., 2004), constraining the volume and source of the LIS meltwater flux through time has proved difficult. While the areal retreat history of the LIS has been refined by decades of research and numerous lines of evidence (e.g., Balco et al., 2002; Dyke, 2004; Ridge et al., 2012), the LIS thinning history is almost wholly unconstrained. Initial attempts to produce data on the thinning history of the LIS relied on ^{14}C dating of basal organic material in high- and low-elevation bogs (e.g., Spear, 1989; Spear et al., 1994). While these projects produced robust age/elevation data points, a highly-variable lag-time between deglaciation and deposition of organic matter in the bogs (Davis and Davis, 1980) lent considerable uncertainty to the data. Geophysical models of ice thickness based on inverted isostatic rebound patterns disagree considerably (e.g., Clark and Tarasov, 2014; Peltier et al., 2015), and numerical model simulations of ice thickness based on paleoclimate evolution contain substantial amounts of uncertainty (e.g., Gregoire et al., 2012; Abe-Ouchi et al., 2015).

The lack of suitable data constraints on the LIS thinning history has propagated uncertainty to other areas of paleoclimate. Notably, the volume of the LIS during its decay has remained uncertain, making it difficult to quantify the contribution of the LIS to global sea-level rise (Carlson and Clark, 2012) and discreet meltwater pulse events observed in deglacial sea-level reconstructions (Clark et al., 2002; Clark et al., 2004; Deschamps et al., 2012; Liu et al., 2016). Without constraints on the timing of LIS thinning, the relationships between the variable deglacial AMOC, the volatile North Atlantic climate, and the LIS remain poorly understood. Furthermore, the height of the LIS is an essential boundary condition for deglacial paleoclimate models (e.g., Barron and Pollard, 2002; Ullman et al., 2014). The height of an ice sheet affects atmospheric circulation, with higher ice sheets acting as larger impediments to upper-atmosphere flow (Bromwich et al., 2004; Abe-Ouchi et al., 2007; Langen and Vinther, 2009). Finally, accurate ice-sheet reconstructions are important for ice-sheet models, which are used for future stability projections and depend on data constraints to inform the internal mechanisms within the models (Applegate et al., 2012; Kleman and Applegate, 2014; Stokes et al., 2015).

For this project, we proposed the implementation of a relatively novel method to constrain part of the LIS thinning history: the ‘dipstick’ approach. In this approach, the ‘dipstick’ is a series of cosmogenic nuclide exposure ages produced at a range of elevations in an area of significant topography (Fig. 1; see commentary by Bierman, 2007). The theory behind the approach is that as the LIS thinned, it exposed more and more topography, and by taking exposure ages from a range of elevations in one location, the lowering ice surface can essentially be tracked through time. This method has been used to constrain thinning histories of the Scandinavian (Goehring et al., 2008); Antarctic (Stone et al., 2003; Ackert et al., 2007; Mackintosh et al., 2011; Johnson et al., 2014), and Greenland (Corbett et al., 2011) Ice Sheets, but has seen limited use with the LIS. Despite its tremendous size and influence on the deglacial paleoclimate, only three dipsticks have been developed for the LIS (Davis et al., 2015; Koester et al., 2017, Davis et al., 2017). This is primarily a consequence of the fact that the LIS predominately covered flat terrain over the interior regions of the northern United States and Canada, limiting the areas where dipsticks can be produced. While the eastern Canadian Arctic provides sufficient topographic relief for dipstick construction, its polar climate was likely conducive to non-erosive, cold-based ice during the last glacial and deglacial periods (Sugden, 1978; Briner et al., 2003), reducing the effectiveness of the dipstick approach (see Methods section for more detail). Therefore, New England and southern Quebec are the only regions with the necessary topography and paleoclimate for successful LIS dipstick construction.

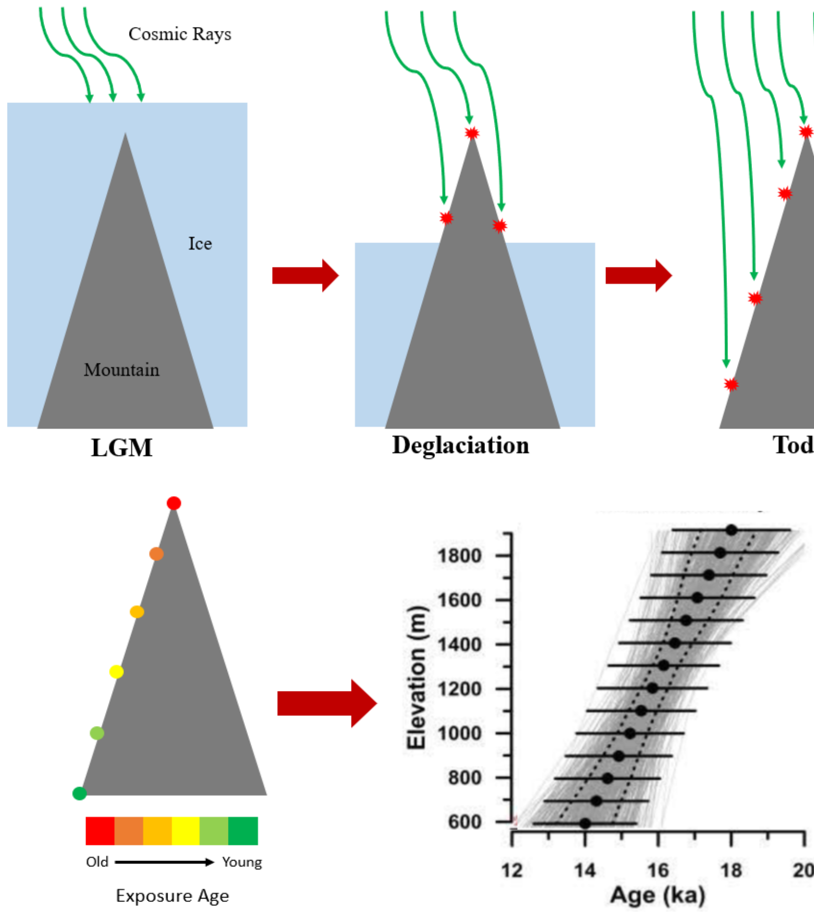

Figure 1: A simplified depiction of the dipstick method. Green lines in top panel are cosmic rays, while red starbursts represent the formation of terrestrial *in-situ* cosmogenic nuclides (TCNs). Top-left panel: at the LGM, a mountain is covered by an ice sheet, blocking cosmic rays from interacting with surficial minerals. Top-middle panel: as the ice sheet begins to thin, cosmic rays interact with minerals near the top of the mountain, forming TCNs at higher elevations only. Top-right panel: if the mountain is presently unglaciated, TCNs form at all elevations. Bottom-left panel: typical expected exposure age profile from a mountain in this scenario, due to early exposure of higher elevations. Bottom-right panel: an example of a 'dipstick' plot using synthetic data. The plot shows exposure age on the x-axis vs. elevation on the y-axis. Monte-carlo simulations (grey lines) using the analytical uncertainty on each point (error bars) give a population of possible thinning timing and rates.

GLOBAL CLIMATE AND SEA LEVEL DURING THE LAST DEGLACIATION

Last Glacial Maximum to The Oldest Dryas (~21 – 19 ka)

Summer insolation at high northern latitudes (boreal summer insolation) rose following the last eccentricity and precession minimums around 24 ka (Carlson and Winsor, 2012), slowly initiating the retreat of Northern Hemisphere ice sheets. A gradual Northern Hemisphere temperature increase that started around 21 ka has been linked to the shift in insolation forcing, with the temporal lag attributed to inertia within the climate system (He et al., 2013). The gradual Northern Hemisphere warming continued from 21-19 ka, a period characterized by a strong AMOC (McManus et al., 2004; Shakun et al., 2012) and the initial retreat of the southern margins of most Northern Hemisphere ice-sheets (Clark et al., 2009). It is estimated that global average sea level rose about 10 m from 21-19 ka, sourced primarily from Northern Hemisphere ice sheets (Fig. 2; Lambeck et al., 2014). The freshwater flux from this initial retreat may have disturbed the density-driven AMOC, which began reducing in strength around 19 ka (Fig. 2; McManus et al., 2004).

The Oldest Dryas (19-14.6 ka)

The AMOC reduction had profound effects on global climate, tipping the bi-polar see-saw and initiating the Oldest Dryas stadial period in the Northern Hemisphere (Liu et al., 2009; Denton et al., 2010; Clark et al., 2012). This stadial period is believed to have been characterized by extreme seasonality, with winter surface air temperatures estimated ~32˚C colder than today, but summer surface air temperatures only ~8˚C colder (Broecker, 2006; Buizert et al., 2014). This pattern is hypothesized to be a consequence of expanded winter sea-ice in the North Atlantic preventing ocean-atmosphere interactions, while continually rising boreal summer insolation maintained mild summer temperatures (Broecker, 2006; Denton et al., 2010). The reduced northward heat transport caused warming in the Southern Hemisphere that ultimately led to a reduction in sea ice and increased upwelling in the Southern Ocean (Denton et al., 2010). The upwelling resulted in a release of sequestered CO_2 into the atmosphere starting around 17.5 ka (Shakun et al., 2012; Clark et al., 2012). The subsequent rise of atmospheric CO_2 concentration led to greenhouse gas warming, emphasizing interstadial conditions in the Southern Hemisphere and gradually reducing the extreme seasonality dominating the Northern Hemisphere (Denton et al., 2010).

Global average sea level rose slowly in the beginning of the Oldest Dryas (from 19-16.5 ka; Carlson and Clark, 2012), likely due to reduced Northern Hemisphere ice sheet retreat in stadial conditions and a lag in Southern Hemisphere warming. From 16.5 – 14.6 ka, sea-level rise accelerated, increasing ~25 m over this span (Carlson and Clark, 2012; Lambeck et al., 2014). However, the source(s) of this large sea level rise are poorly understood, leading Carlson and Clark (2012) to conclude that, "*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 into the ocean, need to be better determined.*"

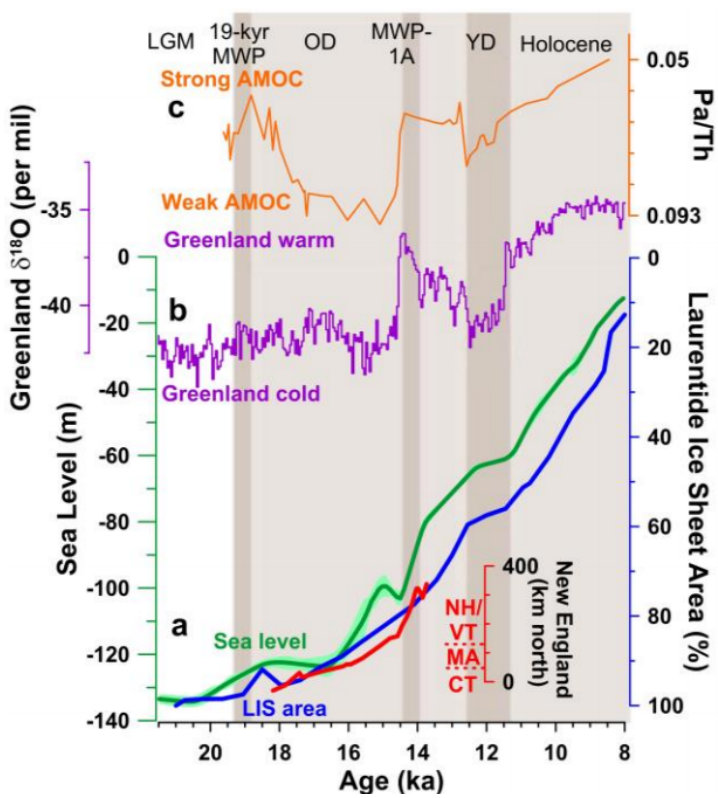

Figure 1. Deglacial ice melt, climate, and ocean circulation (figure from Davis et al., 2017). (**a**) Global sea level (green; from Lambeck et al., 2014), LIS extent (blue; from Dyke, 2004), and northward LIS retreat in central New England based on varves (red; Ridge et al., 2012). (**b**) Greenland $\delta^{18}O$, 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).

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

Around 14.6 ka, proxy records indicate that the AMOC rapidly strengthened (McManus et al., 2004), closely followed by an abrupt Greenland warming (Fig. 2; NGRIP members, 2004). Around the time of the AMOC recovery, one of the most remarkable and poorly understood events of the last deglaciation occurred. Over a period of approximately 300 years, global average sea level rose between ~9-15 m, suggesting a rate of sea-level rise exceeding 40 mm/yr (Deschamps et al., 2012; Lambeck et al., 2014; Liu et al., 2016). Called Meltwater Pulse 1a (MWP-1A), there is no clear consensus the source(s) and cause(s) of this event. Davis et al. (2017) summarized the ongoing attempts: *"While MWP-1A was first assumed to have originated exclusively from the LIS (Fairbanks, 1989; Peltier, 2005), 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. Therefore, any major LIS sea-level contributions could only have come from rapid ice sheet thinning… An accounting of the sources of sea-level rise during this singular event (MWP-1A) is thus far from complete."*

The Allerød interstadial, Younger Dryas, and Holocene (14.3-11.7 ka)

By MWP-1A, the southeastern LIS margin had already retreated north of Mount Greylock. Therefore, subsequent deglacial climate events, although highly interesting and worth further reading for interested trip participants, will only be discussed briefly here. Following MWP-1A, sea-level rise slowed to near the deglacial average rate for the next ~1.5-kyr (Lambeck et al., 2014). The AMOC remained strong during this period (McManus et al., 2004), sustaining interstadial conditions in the Northern Hemisphere that led to high ablation rates and an increased LIS meltwater flux (Ridge et al., 2012). Around 12.9 ka, the AMOC abruptly weakened (Fig. 2; McManus et al., 2004), leading to a brief North Atlantic stadial period known as the Younger Dryas. The cause of this AMOC reduction was initially hypothesized to be a routing of LIS meltwater to the newly exposed St. Lawrence river outlet (Broecker et al., 1989), however, more recent arguments in favor of a northern outlet through the Mackenzie River (Tarasov and Peltier, 2005; Condron and Winsor, 2012) now have robust proxy support (Keigwin et al., 2018). The AMOC recovered around 11.7 ka (Fig. 2; McManus et al., 2004), at the onset of the Holocene.

SOUTHEASTERN LAURENTIDE DEGLACIATION

The mountains of the northeastern United States and southern Quebec were overridden by the southeastern Laurentide at the LGM (Fig. 3 inset). Numerous data constraints on the retreat history of this ice margin, including minimum-limiting organic ^{14}C dates (e.g., Dyke, 2004), ^{14}C-anchored varve series (e.g., Ridge et al., 2012), and ^{10}Be exposure dates on moraines (e.g., Balco et al., 2002; Balco and Schaefer, 2006; Bromley et al., 2015) make it an ideal case study for understanding ice margin-ocean-climate interactions (Fig. 3). However, an almost complete lack of thinning information prevents detailed investigations.

Ice Extent

The southeastern margin of the LIS was at its maximum extent by at least 24 ka, as indicated by ^{10}Be ages of glacially-deposited boulders on the terminal moraines of Martha's Vineyard (29.2 ± 1.7 ka; Balco et al., 2002) and northern New Jersey (25.2 ± 1.3 ka; Corbett et al., 2017). An initial pullback, likely due to rising boreal summer insolation, placed the ice margin north of Buzzard's Bay by 21.0 ± 1.0 ka (Balco et al., 2002). The North American Varve Chronology indicates that southeastern LIS retreat continued steadily (~50 m/yr) during the Oldest Dryas, despite stadial conditions in the adjacent North Atlantic (Ridge et al., 2012). This pattern might be a consequence of the extreme seasonality that characterized North Atlantic climate during this period. As summer surface air temperatures are the primary control on ice sheet volume, due to their control on ablation (Abe-Ouchi, 2013), the mild summers of the Oldest Dryas may have sustained retreat of the southeastern LIS margin. Several moraine sequences were deposited throughout New England near the transition into the Bølling and MWP-1A (Fig. 3; Balco et al., 2009; Koester et al., 2017). After deposition of these moraines, the southeastern LIS margin retreated at a far higher pace (~300 m/yr), ultimately retreating north of New England by 13 ka (Ridge et al., 2012).

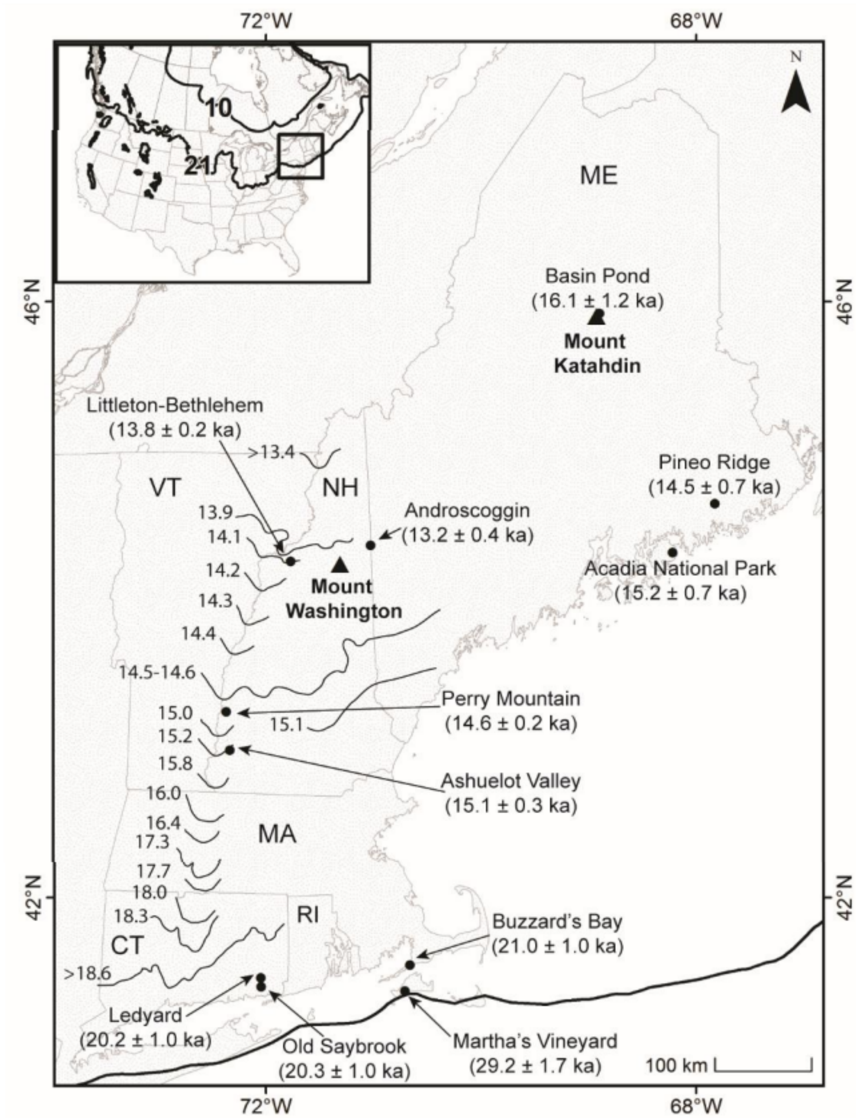

Figure 3. Laurentide Ice Sheet lateral extent through time (figure and caption from Davis et al., 2017). Isochrones show the North American Varve Chronology of deglaciation (from Fig. 11 in Ridge et al., 2012). The black dots are moraines dated using ^{10}Be and calibrated with the northeastern North 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 Thickness

While the retreat chronology of the LIS has been clearly defined in most areas using a variety of methods, data on its thinning history is sparse. Initial attempts to reconstruct a southeastern LIS thinning history produced ^{14}C ages from lake and bog basal sediments in the White Mountains, NH, and Mt. Katahdin, ME, meant to determine minimum-limiting ages for the lowering of the ice surface in the region (Table 1; Davis and Davis, 1980; Spear, 1989; Spear et al., 1994; Rogers, 2003). However, the results contain so much scatter and uncertainty that they are of limited use in determining the timing of ice thinning. It is similarly an open question whether ice sheet drawdown was very rapid (centuries) or much more gradual (millennia).

Site	Elev (m)	^{14}C Age	cal yr BP	Source
Moosilauke				
Deer Lake bog	1325	13,000 ± 400	14,195-16,820	Spear (1989)
Mirror Lake	213	13,800 ± 560	15,720-17,415	Davis and Davis (1980)
Franconia Notch				
Lonesome Lake	831	10,535 ± 495	11,065-13,355	Spear et al. (1994)
Profile Lake	593	10,660 ± 40	12,772-12,885	Rogers (2003)
Mt. Washington				
Lakes of Clouds	1538	11,530 ± 165	10,790-13,355	Spear (1989)
Lost Pond	625	12,870 ± 370	14,580-15,760	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 minimum0limiting, 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.

Preliminary cosmogenic exposure ages from our project have yielded more conclusive results, with exposure-age dipsticks providing evidence of rapid ice thinning in central Maine (~16-15 ka; Fig. 5; Davis et al., 2015), the Presidential Range of New Hampshire (starting ~17 ka; Davis et al., 2017), and on the Maine coast (15.2 ± 0.7 ka; Fig. 5; Koester et al., 2017). It should be noted that the dipsticks from central and coastal Maine feature broad 90% confidence intervals, preventing the exclusion of more complex thinning histories with varying thinning rates. Furthermore, nuclide concentrations from the highest summits in this region are higher than expected, likely due to the incomplete removal of cosmogenic nuclides from prior episodes of exposure by non-erosive, cold-based ice (Bierman et al., 2015; Koester et al., 2017). These high-elevation exposure ages, therefore, cannot be meaningfully interpreted because their cosmogenic nuclide concentrations do not reflect a single period of exposure. The dipsticks reported by Davis et al. (2015. 2017) and Koester et al. (2017) are useful, but to extrapolate the timing and pattern of ice thinning observed at these locations to other areas of the LIS would, by necessity, involve significant uncertainty. Accurately reconstructing the thinning behavior of the entire southeastern LIS therefore requires more data covering a large geographic area.

MOUNT GREYLOCK AND THE SURROUNDING REGION

Mount Greylock is geologically part of the Taconic Mountains but is located near the Berkshire Mountains of northwestern Massachusetts (Fig. 1). It is the tallest mountain in the state, with a summit elevation of 1064 m a.s.l. and a vertical relief of 751 m. Mt. Greylock is composed primarily of metamorphized sedimentary rock, including marble, which is overlain by schist and phyllite of the Taconic allochton (Bierman and Dethier, 1986). Mount Greylock is the product of thrust faulting during the Taconic orogeny, which thrust the Ordovician phyllite and schist over the dolomitic marble (Ratcliffe et al., 1993).

Bedrock is exposed across much of the Mount Greylock uplands, which is otherwise covered by till and colluvial deposits. Glacial striations, chattermarks, and grooves recorded in upland bedrock indicate that the ice-flow was predominately directed southeast, while flow indicators at lower elevations are generally in-line with the valley orientation. This pattern suggests that Laurentide ice became restricted to the valleys at some point in the last deglaciation (Fig. 4a; Bierman and Dethier, 1986). The retreating LIS dammed northward flowing water in the surrounding valleys, forming Glacial Lake Bascom (Fig. 4c), into which extensive deposits of fluvial and subaqueous ice-contact sediment were deposited (Bierman and Dethier, 1986). Lake Bascom reached an elevation of 306-317 m a.s.l. before the Laurentide retreated past a 273 m a.s.l. spillway located at Potter Hill, New York (Fig. 4d).

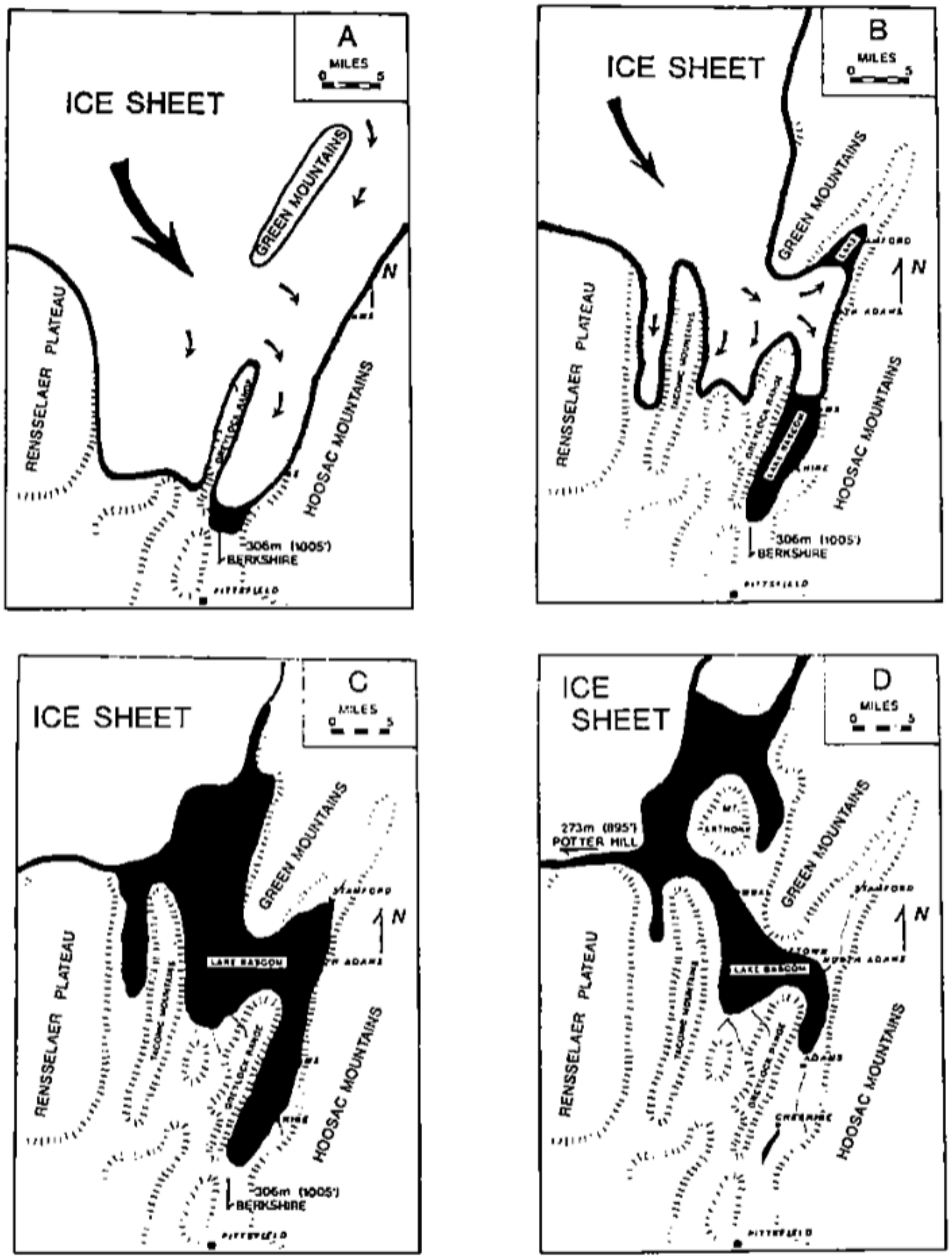

Figure 4. Simplified depiction of Laurentide ice retreat around Mount Greylock and formation of Glacial Lake Bascom (figure from Bierman and Dethier, 1986). Note the topographic control of ice in the valleys in (a) and (b).

RESEARCH MOTIVATIONS.

1) Did the southeastern LIS thinning rate vary during its deglaciation?

The last deglaciation encompassed several stadial and interstadial periods, but currently available thinning histories cannot discern if regional climate influenced the thinning rate of the southeastern LIS. Ice extent constraints indicate significant differences in margin retreat rates during stadial vs. interstadial periods (Dyke, 2004; Ridge et al., 2012); are these differences mirrored seen in thinning rates? Of particular interest is the behavior of the southeastern LIS around the time of MWP-1A.

2) Constrain Individual Ice Sheet Volume Histories

Uncertainties in volume estimates of the LIS through the last deglaciation are problematic because they prevent further investigations into the mechanisms and influence of freshwater forcing on the AMOC, LIS contribution to sea level rise, and the source(s) of discreet meltwater pulse events. Constraining the timing and rate of LIS thinning, even for just the southeastern sector, could provide valuable insight into these systems.

3) Improving Confidence and Accuracy in Deglacial Paleoclimate Models

The height of the LIS is an essential boundary condition for deglacial paleoclimate models (e.g. Barron and Pollard, 2002; Ullman et al., 2014). Ullman et al. (2014) showed that a 20% difference in LIS elevation at the LGM, the difference between published reconstructions by Peltier (2004) and Toscano et al. (2011), resulted in substantial differences in model simulations of the deglacial paleoclimate. Glacial orography has an outsized effect on climate for several reasons. The height of an ice sheet has a direct effect on surface air temperatures through vertical lapse rate, but it also affects atmospheric circulation, with higher ice sheets acting as larger impediments to upper-atmosphere flow (Bromwich et al., 2004; Abe-Ouchi et al., 2007; Langen and Vinther, 2009). The effect of this flow impediment is a re-organization of the atmospheric jet, which has downstream effects on midlatitude storm tracks and wintertime precipitation across much of the NH (Ullman et al., 2014). LIS thickness is also believed to affect the strength of the AMOC, both through its effect on wind stress in the Atlantic (Arzel et al., 2008) and its meltwater flux volume and direction (Clark et al., 2001; Ullman et al., 2014). Present uncertainties about the thickness of the LIS therefore introduce significant uncertainty into paleoclimate models of the last deglacial period.

4) Improved Understanding of Ice Margin Dynamics

Finally, an empirically-based three-dimensional reconstruction of the southeastern LIS would prove useful to ice-sheet modelers. Models used to recreate the decay of the LIS and other LGM ice sheets share many similarities with models designed to predict the future stability of the Greenland and Antarctic ice sheets. However, uncertainties in projection models have led to considerable disagreement in predictions of future sea-level rise (e.g. IPCC, 2013; DeConto and Pollard, 2016; Hansen et al., 2016). A three-dimensional southeastern LIS reconstruction would make it possible to gauge the accuracy of ice-sheet models that are used to simulate past, present, and future ice marginal areas.

Recent geomorphic evidence suggests that the southeastern LIS drained into the Atlantic Ocean through several ice streams, some of which occupied present-day New England and southern Quebec (Margold et al., 2015; Stokes, 2017). Additionally, the ice in this region was almost certainly buttressed by ice shelves that extended into the Atlantic (Stokes, 2017). This makes the southeastern LIS retreat through New England a useful case study for the behavior of complex marginal ice systems such as those in southern Greenland.

RESEARCH STRATEGY

Sampling

For this project, our nuclide of choice is ^{10}Be. As one of the most utilized cosmogenic nuclides in earth surface studies, ^{10}Be has a low analytical uncertainty, offering precise concentration measurements (Gosse and Phillips, 2001). Additionally, a regional calibration data set has been constructed in the northeastern United States by reconciling ^{10}Be concentration measurements with independent age constraints (Balco et al., 2009), mainly the ^{14}C-anchored North American Varve Chronology (Ridge et al., 2012). The ^{10}Be calibration data set has led to more accurate exposure ages in this region (e.g., Bromley et al., 2015; Koester et al., 2017; Davis et al., 2017).

^{10}Be forms primarily in quartz (Gosse and Phillips, 2001), so samples for this project are collected from the upper 2-4 cm of quartz-bearing, glacially-transported boulders or glacially-eroded bedrock. Our attempt to constrain southeastern Laurentide thinning takes advantage of the local topography of the northeastern United States and southern Quebec, with samples being collected from 12 mountains with the greatest vertical relief in their regions (Fig. 4). Sample locations include the highest points in the Green Mountains (Jay Peak, 1177 m; Mt. Mansfield, 1330 m; Killington Mt., 1289 m), the White Mountains (Mt. Washington, 1917 m; Mt. Lafayette, 1600 m), central Maine (Mt. Katahdin, 1606 m; Mt. Bigelow, 1247 m), coastal Maine (Cadillac Mt., 466 m), southern New Hampshire (Mt. Monadnock, 965 m), western Massachusetts (Mount Greylock, 1064 m), eastern Massachusetts (Wachusett, 611 m), southern New York (Catskill Mountains, up to 1277 m), and southern Quebec (Mt. Jacques-Cartier, 1268 m).

Figure 5. Figure from Davis et al. (2017): The locations of 12 ice-sheet mountain dipsticks (red dots) to constrain the thinning history of the southeastern LIS. Note, Mt. Kearsarge has been replaced as a dipstick location by Mt. Monadnock, in southwestern New Hampshire, and Mt. Bigelow, in northwestern Maine, has been added recently. 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.

Nine samples were collected from Mt. Greylock (Fig. 6). MG-01 and -02 were sampled from boulders near the summit composed of the Ordovician phyllite characteristic of the upper reaches of Greylock. MG-03, -04, and -06 were sampled from bedrock quartz veins at 1060, 900, and 780 m a.s.l., respectively. MG-05 and -07 were sampled from boulders at 770 and 900 m a.s.l., respectively, and also seem to be composed of local phyllite. MG-08 and -09 were sampled from large quartzite boulders located on a hillside to the north, across the valley from Mount Greylock, at around 400 m a.s.l.

Cosmogenic Nuclides

Terrestrial *in-situ* cosmogenic nuclides are unique isotopes created during nuclear reactions between terrestrial atoms and high energy cosmic rays. They have been used to analyze a variety of surface processes since the development of accelerator mass spectrometry made their detection and measurement possible (Gosse and Phillips, 2001). Davis and Schaeffer (1955) originally noted that rocks continually exposed to cosmic rays become progressively more irradiated at a rate dependent on the cosmic ray flux at that location. They realized that by measuring the present concentration of a (cosmogenic) nuclide in a rock, estimating the nuclide production rate, and accounting for atomic decay if the nuclide is radioactive, the exposure age of that rock could be calculated. Subsequent research revealed that the production rate of *in-situ* cosmogenic nuclides can be affected by a multitude of parameters, including latitude, elevation, and depth below the surface (Lal, 1991). In typical granites and related rocks, the cosmic ray flux attenuates within approximately 2 m of the surface (Gosse and Phillips, 2001), and it is this property that makes cosmogenic nuclides so appealing for glacial chronologies. Warm-based ice, which can move dynamically on a lubricating layer of meltwater at its base, regular erodes more than 2 m of surficial material when advancing, exposing minerals at the surface that were previously shielded from cosmic radiation (Balco,

Figure 6. Map of Mount Greylock, surrounding valleys, and sample selection sites. Note: Samples MG-08 and -09 are not pictured, they were sampled on the hill located to the north of Mount Greylock. Stars are locations of field trip stops. Dotted line along Greylock ridge is the optional ~2-mile walk along the Appalachian Trail.

2011). These surficial minerals only begin accumulating cosmogenic nuclides when the overlying ice retreats and exposes them to cosmic rays. Consequently, surface exposure ages from glacially-polished bedrock or glacially-transported boulders generally correspond to the date of ice retreat from that location.

There are, however, cases in which the cosmogenic nuclide concentration in a rock sample does not correspond to a single exposure. If a sample was covered by non-erosive, or cold-based, ice that was frozen to the bed, nuclides from previous exposures may survive glaciation and lead to larger measured concentrations than would be expected due to a single exposure. Conversely, if a sample was covered by till, significant snow accumulation, or relict ice-fields after the retreat of continental ice, there will be a lower nuclide concentration than expected due to partial shielding of the sample. These cases must be considered when analyzing nuclide concentrations.

Statistical Evaluation of Thinning Time and Rate

We evaluate the most likely timing and rates of ice thinning at each dipstick location by performing Monte Carlo simulations through the population of exposure age/elevation data points. In each simulation, data points assume a fixed age at random within the bounds of their analytical uncertainty (assuming a Gausian uncertainty distribution), and then a linear regression is run through the resulting absolute age/elevation data points. At least 1000 of these simulations are run, with a filter installed to prevent negative sloping regression lines (indicating thickening ice with time). From these simulations, a population of possible thinning rates and times is generated (Fig. 5; Johnson et al., 2014).

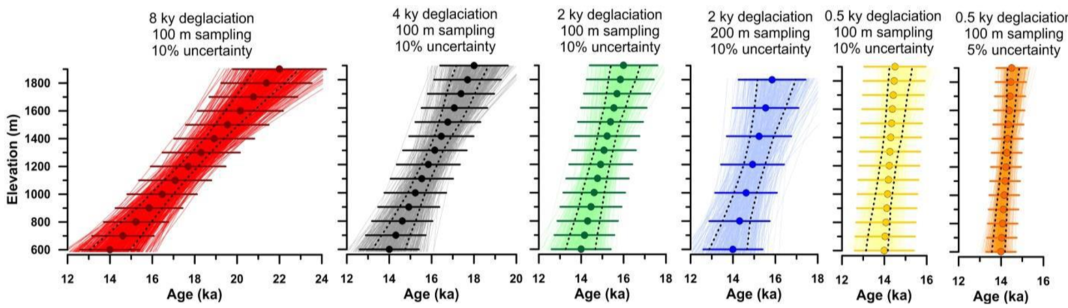

Figure 7. Synthetic dipsticks testing sensitivity to deglaciation duration, sampling density, and ^{10}Be geologic uncertainty typical for cosmogenic datasets. Thin colored lines show 500 Monte Carlo-generated regressions, dotted lines give 68% confidence intervals.

A New England area grand synthesis

The ultimate goal of this project is to generate a three-dimensional depiction of statistically-plausible reconstructions for the retreat of the southeastern LIS. One of our project collaborators recently summarized this goal, *"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 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."* (Davis et al., 2017).

INITIAL RESULTS

Of the 84 samples collected for the New England grand synthesis, 42 have been processed and had ^{10}Be concentrations measured. Of interest to this field trip are exposure ages from samples collected at Mount Greylock (Fig. 6; n = 4) and Peekamoose Mt., in the nearby Catskill mountains of southern NY (Fig. 7; n = 5).

Four exposure ages have been calculated on Mt. Greylock. One of the samples, MG-01, contains more than double the internal uncertainty of any other Greylock sample. This is likely due to the abundance of K^+ and Mg^{2+} observed in the sample, which was present even after Be extraction and may have caused slight interferences in AMS ratio measurements. Including 1σ analytical uncertainty, all ages from Mt. Greylock fall between approximately 13-15.5 ka. Monte Carlo dipstick simulations (Fig. 6a) give a 90% confidence interval for simulated thinning rates at Mt. Greylock of 0.06 – 0.41 m yr^{-1} (Fig. 6b), while highest simulated exposure (1060 m asl) occurred at 14.88 ± 1.16 ka (Fig. 6c)

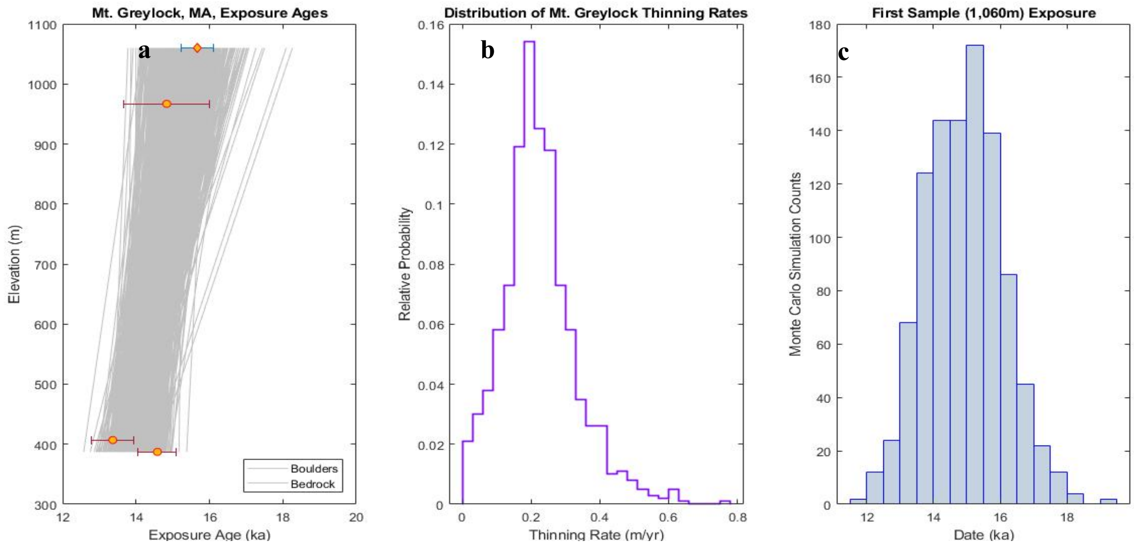

Figure 8. Results of Monte Carlo dipstick simulations using exposure ages from Mt. Greylock, MA. (a) Exposure ages with 1σ measurement uncertainties as well as individual Monte Carlo dipstick simulations (gray lines). (b) Probability distribution of simulated thinning rates. (c) Distribution of exposure ages for the highest elevation captured by the Mt. Greylock dipstick

The five processed exposure ages from Peekamoose Mt. are split into two groups that do not overlap within 1σ analytical uncertainty. Higher elevation ages range from 19.8-18.6 ka, while lower elevation samples have ages between 17.2-18.5 ka. Monte Carlo dipstick simulations (Fig. 7a) give a 90% confidence interval (5th – 95th percentiles) for the LIS thinning rate in the Catskills of 0.09 – 0.22 m yr^{-1}, with a right-skewed distribution that allows for the possibility of much higher thinning rates (Fig. 7b). In addition, the highest elevation simulated, 1170 m asl, was first exposed at 19.46 ± 0.41 ka (Fig. 7c)

The exposure ages from Mount Greylock and Peekamoose Mt. agree within 1σ external uncertainty with the suggested ice retreat in the Connecticut River Valley given by the North American Varve Chronology (Fig. 4; Ridge et al., 2012).

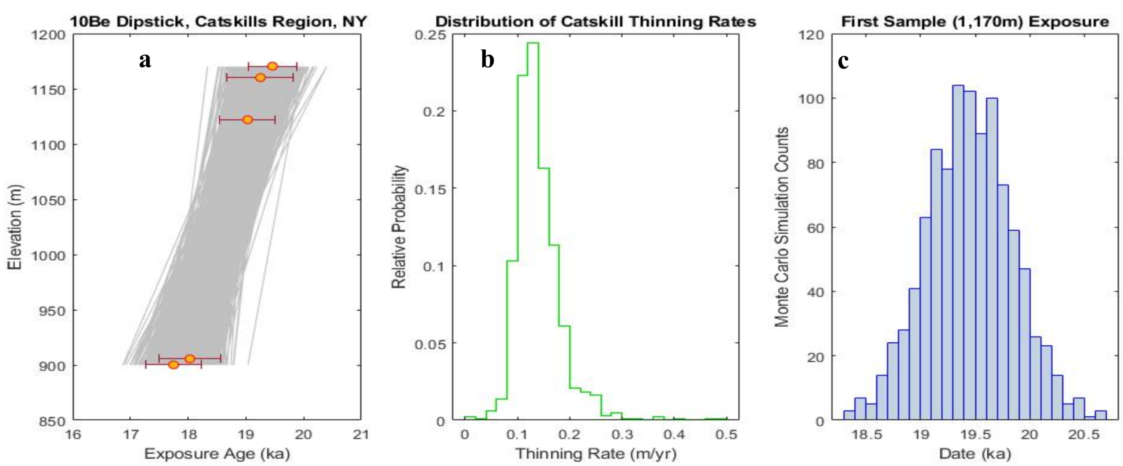

Figure 9. Results of Monte Carlo dipstick simulations using exposure ages from the Catskill Mountains, NY. (a) Exposure ages with 1σ measurement uncertainties as well as individual Monte Carlo dipstick simulations (gray lines). (b) Probability distribution of simulated thinning rates. (c) Distribution of exposure ages for the highest elevation captured by the Catskills dipstick.

DISCUSSION

Together, the exposure ages from Peekamoose Mt. and Mount Greylock provide evidence for southeastern LIS thinning during the Oldest Dryas. This evidence supports the hypothesis of Denton et al. (2005, 2010) that the cold average annual temperatures recorded in Greenland ice cores during the Oldest Dryas were a consequence of extreme seasonality in the North Atlantic, caused primarily by a substantial increase in sea-ice extent during winter. In fact, Buizert et al. (2014) estimated that North Atlantic summer temperatures during the Oldest Dryas may have only been about 8˚C colder than today. Our thinning evidence suggests that summer temperatures in the North Atlantic during the Oldest Dryas were high enough to sustain ablation of the southeastern LIS margin. Additionally, the LIS thinning captured by the Peekamoose Mt. dipstick entirely pre-dates Heinrich Event I (~16 ka; Shaw et al., 2006), indicating that the thinning was not due to drawdown resulting from this ice streaming event. A sustained southeastern LIS thinning during the Oldest Dryas supports the argument of Clark et al. (2001) that the direction and strength of the LIS meltwater flux modulated the strength of the AMOC during the last deglaciation. Southeastern LIS thinning would have had a suppressing effect on the AMOC, and proxy data indicates a suppressed AMOC for over 4-kyr during the Oldest Dryas (McManus et al., 2004).

Despite their locations within 1˚ latitude, samples from Peekamoose Mt. and Mount Greylock appear to capture two different stages of LIS thinning. While Monte Carlo simulations suggest that the lowest sample on Peekamoose Mt. (~900 m a.s.l.) may have been exposed around the same time as the highest Mount Greylock sample (~1050 m a.s.l.), >90% of the simulations have the entirety of the Peekamoose Mt. dipstick exposed at least a thousand years earlier than the highest Mount Greylock sample. This pattern agrees with other New England deglacial chronologies, which suggest that ice retreated slowly during the Oldest Dryas, taking about 2-kyr to retreat northward through Massachusetts (Fig. 4; Ridge et al., 2012). Samples from lower elevations on Peekamoose Mt. might reveal a more extended thinning history, but it is entirely plausible that the mountain was exposed down to 900 m a.s.l. before the summit of Mount Greylock (~1050 m a.s.l.) was exposed from underneath the LIS. The proximity of these two sites combined with the apparent lag of exposure could indicate that the southeastern LIS margin was steeply sloped during its retreat.

With our presently available data, definitive conclusions on the LIS thinning rate at Peekamoose Mt. and Mount Greylock cannot be made, but general statements are offered here. All but two of the samples processed from Mount Greylock have overlapping exposure ages (within 1σ analytical uncertainty), preventing the exclusion of extremely rapid thinning (>1 m/yr). Evidence of rapid LIS thinning has been noted by our project collaborators in coastal Maine (Koester et al., 2017), central Maine (Davis et al., 2015), and the Presidential Range of New Hampshire (Davis et al., 2017). The exposure ages from Peekamoose Mt. suggest a slower thinning rate (<1 m/yr), but again, more data is needed to make conclusive statements.

ROAD AND TRAIL LOGS

Time, Place, Logistics

START TIME AND LOCATION: Trip begins on Sunday, October 14th, at 9:00 AM outside the Mount Greylock State Reservation Visitor Center located at 30 Rockwell Rd, Lanesborough, MA 01237. The drive to the Visitor Center is about 2 hours from Lake George, NY, and parking and restroom facilities are available.

DESCRIPTION: This field trip will begin at the Mount Greylock Visitor Center with a brief introduction to the geologic history of the area, focused primarily on surficial geomorphology and the influence of continental glaciation on Mount Greylock and adjacent valleys. While at the Visitor Center we will try to consolidate vehicles. We will then drive up Rockwell Rd towards the summit, making a detour down Sperry Road to Stony Ledge, a lookout point that offers magnificent views of Mount Greylock. This is one of the few close-up vantage points to view “The Hopper”, the most southerly glacial cirque in New England. More discussion about the influence of glaciation on Mount Greylock will occur here, and field trip participants are encouraged to (carefully) explore Stony Ledge for glacial striations in the exposed bedrock. Following the stop at Stony Ledge, we will return to Rockwell Rd and continue up to the summit. At the summit, trip participants can take advantage of the Bascom Lodge, which has restroom facilities, food and drinks for purchase, a clean water spigot, and information about the surrounding area. At the summit, we will go over a more in-depth review of local and regional geomorphology, with an emphasis on the Laurentide Ice Sheet and its behavior as it overrode and then retreated past Mount Greylock. We will also briefly discuss the motivations and approach of an NSF-funded research project to constrain the timing and rate of ice thinning in the northeastern United States and point out one of the sample locations on the summit. Weather permitting, interested trip participants are invited to then take a ~2 mile walk north on the historic Appalachian Trail to Mt. Williams, a part of the Greylock ridgeline. Along the way, we will point out other samples for the ice-thinning project and the lead author (a hiking fanatic) will discuss some of the history of the Appalachian Trail and early exploration in the area. For those not participating in the Appalachian Trail walk, the field trip will end by approximately 1 pm. For those participating in the walk, the trip should end by around 3 pm.

Road Mileage

0.0 Begin at Mount Greylock State Reservation Visitor Center (30 Rockwell Rd, Lanesborough, MA)

5.5 Take left turn off Rockwell Rd onto Sperry Rd

7.1 **STOP 1.** Park at Stoney Ledge. Sperry Rd makes a loop here, please find parking off the road. Stoney Ledge offers fantastic views of Mount Greylock and The Hopper, a glacial cirque on Greylock's west flank. Exposed bedrock offers the potential to investigate glacial striations!

8.7 After driving back down Sperry Rd, take a left turn back on to Rockwell Rd, heading towards the summit

10.2 Rockwell Rd. joins North Adams Rd, continue right towards summit

11.0 **STOP 2.** Arrive at Mount Greylock summit. Park in designated parking lot and meet outside Bascom Lodge. Bascom Lodge provides restrooms, food for purchase, and free clean drinking water. An in-depth discussion of regional glacial geomorphology, the chronology of the Laurentide Ice Sheet in the area, and the motivations and preliminary results of an NSF-funded ice sheet reconstruction project will take place here. For those not participating in the walk on the Appalachian Trail, the summit marks the end of the trip. Auto roads depart the summit to the north and south (we will drive up the southern road), please drive carefully!

Trail Mileage for walk along Appalachian Trail

0.0 Start walk at Mount Greylock summit.

0.25 **STOP 1.** Briefly examine two glacially-transported boulders deposited on the Greylock ridgeline. These boulders were sampled for the NSF dipstick project.

2.2 **STOP 2.** Mt. Williams summit. This is not truly a separate mountain, as it is part of the Greylock ridgeline, but it is a local high-point that offers fantastic views of the surrounding area. Two samples were collected at this location for the NSF dipstick project, one bedrock and one boulder. From here, trip participants can choose to either return via the AT to the Greylock summit or continue down the AT for another half mile, eventually meeting the northern Greylock auto road (Notch Rd). The trip leader will arrange to have a van meet at Notch Rd and transport participants back up to the summit, although participants are welcome to arrange their own transport if desired.

End of trip; thank you

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