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Author(s)	Bromley, Gordon R.M.; Hall, Brenda L.; Thompson, Woodrow B.; Kaplan, Michael R.; Garcia, Juan Luis; Schaefer, Joerg M.
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Late glacial fluctuations of the Laurentide Ice Sheet in the White Mountains of Maine and New Hampshire, U.S.A.

Gordon R.M. Bromley[1*], Brenda L. Hall[1], Woodrow B. Thompson[2], Michael R. Kaplan[3], Juan L. Garcia[1,4], Joerg M. Schaefer[3]

[1]*School of Earth & Climate Sciences and the Climate Change Institute, Edward T. Bryand Global Sciences Center, University of Maine, Orono, ME 04469-5790, USA*

[2]*Maine Geological Survey, 93 State House Station, Augusta, ME 04333-0093*

[3]*Instituto de Geografia, Pontificia Universidad Catolica de Chile, Avenida Vicuna Mackenna 4860, Santiago, 782-0436, Chile*

[4]*Lamont-Doherty Earth Observatory, Geochemistry, Route 9W, Palisades, NY 10964, USA*

*Corresponding author. Tel: +1 207-581-2190; Fax: 207-581-1203;
E-mail: gordon.r.bromley1@maine.edu

Abstract. Prominent moraines deposited by the Laurentide Ice Sheet in northern New England document readvances, or stillstands, of the ice margin during overall deglaciation. However, until now, the paucity of direct chronologies over much of the region has precluded meaningful assessment of the mechanisms that drove these events, or of the complex relationships between ice-sheet dynamics and climate. As a step towards addressing this problem, we present a cosmogenic ^{10}Be surface-exposure chronology from the Androscoggin moraine complex, located in the White Mountains of western Maine and northern New Hampshire, as well as four recalculated ages from the nearby Littleton-Bethlehem moraine. Seven internally consistent ^{10}Be ages from the Androscoggin terminal moraines indicate that advance culminated ~13.2 ±

0.8 ka, in close agreement with the mean age of the neighboring Littleton-Bethlehem complex. Together, these two datasets indicate stabilization or advance of the ice-sheet margin in northern New England, at ~14-13 ka, during the Allerød/Greenland Interstadial I.

Introduction

During the last glacial maximum (LGM; ~26-19 ka), the Laurentide Ice Sheet (LIS) constituted the largest ice mass on Earth. With an area of ~13,000,000 km^2 and maximum surface elevations of as much as 3000 m (Denton and Hughes, 1981), the LIS exerted a considerable influence on global climate through albedo forcing (e.g., Broccoli and Manabe, 1987), displacement of atmospheric circulation (e.g., Williams *et al.*, 1974; Hostetler and Clark, 1997; Oviatt, 1997; Birkel, 2010), and, particularly along the ice-sheet's Atlantic margins, ocean-ice sheet interactions (e.g., Bond *et al.*, 1993; MacAyeal, 1993). Indeed, perturbations of the ice sheet's maritime sectors have been invoked widely as a key component – if not the cause – of Heinrich events (e.g., Hulbe *et al*., 2004), during which considerable volumes of ice apparently were disgorged into the North Atlantic Ocean with profound consequences for downwind ocean circulation (Broecker, 2003). Concurrently, the temperate nature and relatively low latitude of the ice sheet mean the LIS itself will have been highly sensitive to climate, particularly along southern and eastern margins where deglacial readvances have been correlated with perturbations of North Atlantic circulation (e.g., Lowell *et al.*, 1999; Dorion *et al.*, 2001; Borns *et al*., 2004; Kaplan, 2007). Consequently, accurate reconstruction of past ice-sheet behavior provides the means to assess both the role of the LIS in Late Quaternary climate variability and sensitivity of the ice sheet to key events, such as Heinrich stadials and late-glacial climate reversals.

Comprehensive geologic mapping of glacial deposits and features over the past several decades has revealed the overall pattern of ice-sheet behavior in New England during the LGM and subsequent deglaciation. At its maximum extent, the southeastern sector of the LIS terminated at the large moraine systems off the coast of southern New England – Long Island, Martha's Vineyard, and Nantucket Island (Fig. 1; Oldale, 1982; Oldale and O'Hara, 1984; Stone and Borns, 1986; Balco *et al*., 2002) – while farther east the ice margin lay far offshore at George's Bank in the Gulf of Maine (Fig. 1; Pratt and Schlee, 1969; Stone and Borns, 1986). Inland from these limits, minor moraine belts and landform assemblages in southern New England document the northward retreat of the ice-sheet margin through Connecticut, Massachusetts, and into New Hampshire (Fig. 1; Koteff and Pessl, 1981; Ridge, 2004; Balco and Schaefer, 2006; Balco *et al*., 2009). In Maine, swarms of minor moraines occur where deglaciation of the coastal lowland was accompanied by marine submergence. The distribution of glacial landforms in these areas indicates that ice-sheet retreat initially was gradual and oscillatory, with frequent stillstands and minor readvances resulting in the extensive (> 100 km in length) coastal moraine belt in Maine and New Brunswick (Borns, 1973, 1980; Borns and Hughes, 1977; Rampton *et al*., 1984; Thompson and Borns, 1985; Smith and Hunter, 1989; Kaplan, 1999; Borns *et al*., 2004). In a few places, the advances or stillstands were long enough to build extensive deposits, such as the Pond Ridge moraine and Pineo Ridge moraine complex (Borns *et al.*, 2004; Fig. 1). The scarcity of moraines and prevalence of eskers proximal to the coastal moraine belt indicates that subsequent deglaciation proceeded rapidly and without major interruption until the ice margin reached the vicinity of the Appalachian Mountains (Davis and Jacobson, 1985). There, several extensive moraine complexes record fluctuations of this sector of the LIS and are the focus of this study.

Compared to extensive mapping studies, the glacial chronology for this SE sector of the LIS is less complete and, in places, controversial. On the basis of limiting radiocarbon ages and existing ^{10}Be surface-exposure ages, we know that the ice sheet was most extensive during the LGM *sensu stricto* (Tucholke and Hollister, 1973; Oldale, 1982; Stone and Borns, 1986; Stone *et al.*, 1998, 2002; Balco *et al*., 2002, 2009), consistent with other sectors of the LIS (Dyke and Prest, 1987). Recalculated ^{10}Be ages from the Buzzards Bay moraine in Massachusetts (Fig. 1) suggest that the ice margin remained at or near its maximum extent until ~21 ka (Balco *et al*., 2009). However, estimates for the onset of deglaciation vary widely, with varve (e.g., Ridge, 2004) and surface-exposure (Balco and Schaefer, 2006; Balco *et al*., 2009) chronologies indicating a significantly earlier (by as much as ~5 ka) retreat of the LIS than that determined by recent terrestrial radiocarbon dating (e.g., Peteet *et al*., 2012). Nonetheless, radiocarbon ages of mollusk shells from raised marine deposits in eastern Maine indicate that the ice-sheet margin had retreated to the present-day coastline by ~17-15 ka, with the Pineo Ridge moraine complex representing a major stillstand (Smith and Hunter, 1989) and/or readvance (Borns, 1980; Borns and Hughes, 1977) of the LIS towards the end of that period (e.g., Kaplan, 1999, 2007; Dorion, 1997; Dorion *et al*., 2001; Borns *et al*., 2004).

Subsequent portions of the deglacial record are problematic. For example, the deglaciation of central and northern Maine is constrained by a relatively sparse coverage of minimum-limiting ^{14}C ages, many of which are based on bulk samples and, in some cases, affected by a marine reservoir effect of uncertain magnitude (Davis and Jacobson, 1985; Ridge *et al.*, 2001; Borns *et al*., 2004). In comparison, the varve chronologies constructed for central New England (Antevs, 1922, 1928; Ridge *et al*., 2001; Ridge, 2004) offer a seemingly precise chronology for ice-sheet retreat through that sector. However, as argued by Borns *et al*. (2004) and later by Peteet *et al*. (2012), these essentially "floating" chronologies are pinned to the calendar timescale in relatively few locations (see below), raising the possibility of systematic age offsets in the deglacial record. Thus, much of the chronology of ice-sheet retreat in New England is known only on a first-order basis.

Together, the glacial stratigraphy and existing chronology for New England provide some insight into ice-sheet behavior during the last glacial-interglacial transition. However, the temporal resolution is insufficient to reveal the fine structure of deglaciation or to assess the response of the LIS to millennial-scale climate forcing. This is particularly true for the late-glacial period (~15-11.6 ka), during which the climate system underwent high-magnitude, potentially abrupt transitions on both regional and global scales. For example, although several studies have reported evidence for a cryospheric response in New England to such events as the Younger Dryas stadial (YD) (e.g., Lamothe, 1992; Borns *et al*., 2004) and Heinrich stadials (e.g., Balco *et al*., 2002; Ridge, 2004; Kaplan, 2007), these correlations have yet to be tested and therefore remain equivocal.

To help develop the deglacial record of New England, specifically after the LIS had retreated inland of the marine limit, we present a ^{10}Be surface-exposure chronology for the Androscoggin moraine complex, located in the White Mountains of western Maine and eastern New Hampshire (Figs. 1,2). Together with the neighboring Littleton-Bethlehem (L-B) moraine (Figs. 1,2; Thompson *et al*., 1996, 1999, 2002), the Androscoggin moraine represents renewed advance - or a stillstand - of the LIS in northern New England during overall deglaciation. However, until now the Androscoggin moraine has not been dated directly. As a result, the significance of these prominent moraines to the deglacial record of the LIS hitherto was poorly understood. In this paper, we discuss the implications of our Androscoggin moraine dataset in conjunction with four recalculated ^{10}Be ages from the L-B moraine.

Geologic Setting and Methods

The White Mountains of New Hampshire and western Maine contain a rich geologic record of LIS deglaciation that has been studied in detail since the mid-19th Century (e.g., Lyell, 1850; Agassiz, 1870; Thompson, 1999, and references therein). The most conspicuous depositional landforms include the L-B moraine system, a broad complex of ridges in the vicinity of the towns of Littleton and Bethlehem (Fig. 2; see Thompson *et al.*, 1999), and a group of prominent lateral and terminal moraines straddling the Maine-New Hampshire border in the Androscoggin River valley (Figs. 2,3) that are the focus of this paper.

The Androscoggin moraines, comprising approximately twenty sections of lateral and terminal ridges located on both sides of the valley (Fig. 3) (Stone, 1880; Thompson and Fowler, 1989), are some of the most prominent moraines in the White Mountains (Upham, 1904). As detailed by Thompson and Fowler (1989), the arcuate distribution of the moraines defines the former terminus of a ~3 km-wide, east-flowing glacier tongue and its subsequent separation into discrete sub-lobes during initial stages of retreat. The Androscoggin moraines exhibit considerable relief (exceeding 30 m in places) and sharp, well-preserved crests, indicating little post-glacial reworking. Moreover, the crests are mantled with boulders of local and regional provenance (Fig. 4), the majority of which are quartz-bearing and thus ideal for ^{10}Be surface-exposure dating (see below).

Previous work has established the age of the Androscoggin moraine complex only in a broad sense. A basal age of 12,450 ± 60 ^{14}C yr BP (OS-7125) (14,985−14,219 cal yr BP (IntCal 13); Thompson *et al.*, 1996) from Pond of Safety (Fig. 2), located approximately 30 km north-west of the moraine in the upper Israel River valley, has been interpreted as a minimum age for the Androscoggin complex (Thompson *et al*., 1999). However, owing to the challenging terrain and distances involved, the exact correlation between the two sites – and thus the relevance of the basal age to the moraine – has not been established unequivocally.

Age control notwithstanding, Thompson *et al.* (1999) suggested on the basis of relative position that the Androscoggin moraine might predate the extensive L-B moraine, whereas a subsequent interpretation proposed that the two complexes might correspond to the same advance (Thompson *et al*., 2007). Because the two moraine systems are not laterally continuous, and owing to heavy forest cover, it is not possible to determine their age relationship from mapping alone.

Compared to the Androscoggin moraine, the age of the L-B complex is better resolved owing to its direct association with the New England varve record at Comerford Dam, on the Vermont-New Hampshire border (Ridge *et al.*, 1996; Fig. 2). Indeed, the correlation of the moraines with a till deposited between varve years 7154 and 7305 (Ridge and Larsen, 1990; Ridge *et al*., 1999) allowed Balco *et al*. (2009) to assign a mean age of 13,840 ± 250 yr to the complex as part of the north-east North America (NENA) ^{10}Be production-rate calibration study. Balco *et al*. (2009) also collected four surface-exposure samples from two sites − the Sleeping Astronomer and Beech Hill moraines (Fig. 2) − located close to the distal edge of the L-B moraine belt. In the current study, we present recalculated ages for those samples using the beryllium concentrations published by Balco *et al.* (2009).

Our Androscoggin surface-exposure chronology is based on ^{10}Be ages (*n*=7) from coarse-grained granite boulders, all but one (AM-07-01) of which are located on the crests of two terminal moraines protruding eastward from Hark Hill (Figs. 3, 4). Sample AM-07-01 is located on a bedrock knoll immediately inside the moraine complex (Fig. 3). We targeted these prominent landforms because they represent the closest well-preserved moraines to the maximum terminal position of the Androscoggin glacier. Samples comprise the upper few centimeters (≤ 5 cm) of rock from the boulders' horizontal top surfaces and were collected with a hammer and chisel. Although sampled surfaces exhibited pronounced glacial molding, the absence of striae and polish suggests a minor degree of post-depositional erosion, as reported by Balco *et al.* (2009) for samples from the L-B moraines. As argued below, however, this process is unlikely to have had a significant effect on exposure age (see Results).

To reduce the likelihood of shielding effects due to snow or vegetation cover, we sampled only boulders greater than 1 m in relief (Table 1). Additionally, we calculated chi-squared (χ^2) values for both the AM and L-B data sets to assess the impact, if any, of post-depositional geological and/or environmental processes on exposure-age distributions (see Results).

Samples were prepared at the University of Maine and Lamont-Doherty Earth Observatory. We separated quartz following established heavy-liquid and HF-leaching procedures, after which beryllium was extracted following the methods described by Schaefer *et al*. (2009). Beryllium ratios of samples and blanks were measured at the Lawrence-Livermore CAMS facility relative to the 07KNSTD standard [^{10}Be/^{9}Be = 2.85×10^{-12}; Nishiizumi *et al*., 2007] (note: ratios for the L-B samples were measured relative to the KNSTD standard [^{10}Be/^{9}Be = 3.15×10^{-12}]). We calculated surface-exposure ages using the CRONUS-Earth online calculator, version

2.2 (Balco *et al*., 2008), in conjunction with the NENA ^{10}Be production rate (Balco *et al*., 2009) and the time-independent "St" scaling (Lal, 1991; Stone, 2000). To provide an independent check of the NENA-derived ages, and to prevent circular reasoning (i.e., the L-B samples are a constituent of the NENA calibration), we also calculated ages using two recent, non-local ^{10}Be production rates, one from north-eastern North America (Young *et al*., 2013; see Table 3) and one from the Southern Alps of New Zealand (~44°S; Putnam *et al*., 2010). In calculating moraine mean age, we also propagated analytical uncertainties with a systematic uncertainty (4.8%) associated with the NENA production rate calibration (Table 1). Although inclusion of such a systematic uncertainty is not needed for comparison between the closely located L-B and Androscoggin moraines (e.g., Fig. 5), it is a consideration when comparing to records determined by other chronological methods.

Our choice of St scaling reflects the close match this scheme provides to independent chronological constraints in New England (Balco *et al*., 2009) and elsewhere (Fenton *et al.*, 2011; Briner *et al.*, 2012; Kelly *et al.*, 2013; Young *et al.*, 2013). However, we stress that our conclusions are independent of our choice of scaling scheme, because at these latitudes and relatively low elevations the respective ages are statistically identical (Table 2). Analytical results and ages are given in Table 1.

Results

The seven ^{10}Be ages from the Androscoggin moraine – constituting the first surface exposure-dated moraine from the state of Maine – are given in Table 1 and are shown in Figures 3 and 5. The four recalculated surface-exposure ages from the L-B moraine are given in Table 1 and Figure 5. Together, the seven Androscoggin ages form an internally consistent grouping, without outliers (i.e., ±1σ), ranging from 13.7 ± 0.6 ka to 12.6 ± 0.5 ka, and provide an average moraine age of 13.2 ± 0.4 ka, or 13.2 ± 0.8 ka if a~5% systematic uncertainty is propagated. Plotted as an age-probability curve, the ages exhibit a normal distribution with a peak age of 13.2 ka (Fig. 5). In comparison, the four samples collected from the L-B moraine by Balco *et al.* (2009) range from 14.0 ± 0.4 ka to 13.5 ± 0.5 ka, and provide a mean moraine age of 13.8 ± 0.2 ka (or 13.8 ± 0.7 ka with a 5% propagated uncertainty) and a peak age of 13.8 ka (Fig. 5). As the three production rates (Balco *et al*., 2009; Putnam *et al*., 2010; Young *et al*., 2013) are statistically identical, our moraine ages are consistent regardless of which rate is used. Moreover, both of the two more precise rates (Putnam *et al*., 2010; Young *et al.*, 2013) produce exposure ages for L-B samples that are congruent with that inferred from the New England varve chronology (Ridge *et al*., 1999; Balco *et al.,* 2009), supporting their use throughout eastern North America. We also point out that the very close agreement between the production rate of Young *et al*. (2013) and the New England varve chronology implies our propagation of a 5% systematic uncertainty is conservative.

While we acknowledge the possibility of geologic or environmental processes affecting our Androscoggin ages, several lines of evidence point to these processes having had minimal impact on our age calculations. First, despite the evidence for minor erosion of boulder surfaces, both the glacial molding of selected boulders and the internal consistency of our data set (Figs. 3, 5) suggest this effect has been minor. Second, comparison of ^{10}Be age versus boulder height shows no significant relationship (Table 1), as would be expected with snow shielding. Third, the Androscoggin moraines are well defined and sharp crested, indicating that post-depositional modification has been negligible. Finally, chi-squared values for both the

Androscoggin and L-B data sets are < 1; values of ~1−2 indicate that age differences can be explained by analytical uncertainties alone (Bevington and Robinson, 1992). Processes such as snow shielding, substantial boulder erosion, and moraine deflation should cause each boulder to have a different exposure history. Given that our ages are statistically indistinguishable from one another, we conclude that such post-depositional processes have had a negligible effect on our data set.

Discussion

The prominent nature of the Androscoggin moraines suggests that this complex, like the L-B moraine belt (Thompson *et al*., 1999), represents an advance or stillstand of the LIS during overall retreat. However, owing to dense forest cover and challenging terrain, the exact relationship between the two complexes has remained ambiguous. Our new ages for the Androscoggin moraine and recalculated ages for the L-B moraine provide important insight into the deglaciation of northern New England. As noted above, we discuss the data as calculated using the local NENA production rate, but stress that our interpretations would remain unchanged using other recently published, indistinguishable production rates (see Table 3). Indeed, it is noteworthy that these three production rates produce results that are indistinguishable statistically despite having been calibrated in starkly different geographic and environmental contexts (i.e., New England, Arctic, Southern Alps).

Although the peak ages of the two moraine complexes differ by ~600 yr, their respective mean ages overlap within 1σ uncertainty and thus the datasets are indistinguishable statistically. Therefore, we suggest that the Androscoggin and L-B moraine systems represent the same late-glacial advance/stillstand of the LIS, as hypothesized originally by Thompson *et al*. (2007). Viewed in greater detail, the small discrepancy in peak age potentially reflects differences in the distribution of moraines and sampled boulders between the two sites. For instance, the sprawling geographic distribution of the L-B complex (Upham, 1904; Thompson *et al*., 1999) indicates this moraine belt was deposited over a considerable period, while the four existing L-B ^{10}Be ages, from the Sleeping Astronomer and Beech Hill moraines, constrain the age of only the most distal parts of the complex. In contrast, the Androscoggin moraines, which are primarily composite landforms, exhibit a more compact spatial distribution. Our seven samples, collected from the crests of those composite moraines, thus represent the most recent period of deposition. Therefore, it is conceivable that the marginally younger peak age of the Androscoggin moraines correlates with the more proximal, as yet undated deposits in the L-B complex. Nonetheless, given the statistically indistinguishable ages of the two moraine complexes, we acknowledge that we are unable to test this model with the existing dataset.

We also note that our surface-exposure data conflict with the original interpretation of the Pond of Safety radiocarbon chronology. By that model, the ^{14}C date from the basal sediments of the pond constitutes a minimum-limiting age for the Androscoggin moraines but a maximum-limiting age for the L-B complex, moraines of which are located proximal (relative to ice-flow direction) to the pond (Thompson *et al*., 1996, 1999). However, our data suggest that the Androscoggin moraines postdate the onset of sedimentation in - and thus deglaciation of - the pond, and therefore must be younger. This conflict is resolved by our interpretation of the Androscoggin and L-B moraines as representing the same advance. Because Thompson *et al.* (1996, 1999) have established that Pond of Safety represents a maximum-limiting age for the L-B complex, it follows that this basal date must also constitute a maximum age for the Androscoggin moraines.

Together, our tightly grouped ages from the Androscoggin River valley and the recalculated ages from Balco *et al.* (2009) show that the ice-sheet margin advanced during the Allerød (GI-1a-c; Lowe *et al*., 2008), resulting in formation of the Androscoggin and L-B moraine complexes. This interpretation is reinforced by the maximum-limiting basal radiocarbon age (14,985−14,219 cal yr BP) from Pond of Safety, located distal to the L-B complex in the upper Israel River valley. Moreover, as surface-exposure ages typically represent the culmination of a glacial advance close to the onset of retreat, our Androscoggin chronology also might suggest that this sector of the LIS retreated during the subsequent YD stadial. This interpretation is supported by the absence of nearby moraines up-valley from the Androscoggin terminus (Borns *et al*., 2004).

At first glance, this disparity between glacial activity in northern New England and the widely accepted late-glacial temperature record for the North Atlantic region represents an ostensible paradox, in terms of the assumed climatic conditions during the YD stadial. Indeed, within New England, organic-poor, minerogenic sediments dating to the late glacial have been documented at several sites in Maine and the White Mountains and are interpreted as a return to stadial conditions during the YD (e.g., Dorion, 1997; Borns *et al*., 2004; Dieffenbacher-Krall and Nurse, 2006). Nonetheless, we note that this pattern of Allerød advance–YD retreat has been documented elsewhere in the Northern Hemisphere, including at sites immediately adjacent the North Atlantic Ocean where YD cooling apparently was greatest (Bowen, 1999; Lohne *et al.*, 2007; Bromley *et al*., 2014). Moreover, our New England record is in accord with evidence for mild YD summertime temperatures in Greenland (Björck *et al*., 2002) and with the absence of YD moraines (e.g., Mangerud and Landvik, 2007; Hall *et al.*, 2008; Kelly *et al.*, 2008; Briner *et al*., 2009) or stratigraphy (e.g., Wolfe, 1996) at many northern high-latitude sites. Such late-glacial behavior potentially points to the effects of extreme seasonality adjacent the North Atlantic during the YD (Denton *et al*., 2005), with warm summers driving glacial recession and very cold winters causing the depressed mean-annual temperatures recorded by ice-core proxies. If true, the sedimentologic signature of the YD recorded in New England lakes potentially reflects periglacial remobilization of slope deposits rather than year-round cooling (Putnam and Putnam, 2009).

Finally, our new chronology supports the findings of Richard and Occhietti (2005) that late-glacial marine transgression of the St. Lawrence lowlands occurred later than indicated by basal radiocarbon ages from the former Champlain Sea (Rodrigues, 1992). Construction of the L-B and Androscoggin moraines would have required the persistence of a robust ice-sheet margin in northern New England at least as late as ~13.2 ka, a scenario that is supported by terrestrial radiocarbon data from elsewhere in this region (see Thompson *et al*., 1999; Borns *et al*., 2004) but which may be incompatible with ice-free conditions in the St. Lawrence River valley immediately north of New Hampshire as early as ~14 ka (Parent and Occhietti, 1988). This discrepancy is removed, however, if the radiocarbon dates upon which the age of the Champlain Sea is based incorporate a considerable reservoir effect, as suggested by Rodrigues (1992) and demonstrated by Richard and Occhietti (2005). Thompson *et al.* (2011) found evidence of a 1000-yr reservoir at a site with juxtaposed marine and terrestrial organics, located in Portland, Maine.

Conclusions

Our cosmogenic ^{10}Be surface-exposure chronology from the Androscoggin moraine, coupled with four recalculated ages from the L-B complex, affords robust and directly dated evidence for a major late-glacial fluctuation of the LIS in northern New England. Retreat of the ice sheet north of the White Mountains was interrupted by at least one advance during the Allerød, resulting in deposition of the prominent L-B and Androscoggin complexes. Seven ages from the terminal moraines of the Androscoggin complex indicate deposition ~13.2 ± 0.8 ka, broadly coincident with the emplacement of the Sleeping Astronomer and Beech Hill moraines in the L-B complex, leading us to conclude that the two moraine belts represent the same late-glacial climate reversal. Ultimately, our chronology indicates that the LIS in northern New England underwent readvance, or stillstand, during the Allerød. These findings have important implications for our understanding both of ice-sheet behavior in New England during the last glacial termination and the terrestrial impact of deglacial climate changes, including cold stadials, in the North Atlantic region.

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Figure and Table Captions

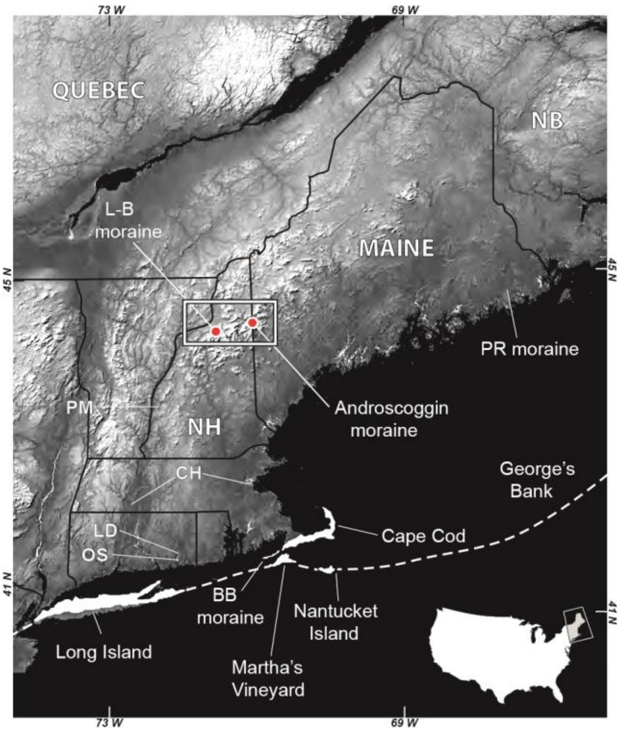

Figure 1 Map of New England and SE Canada showing the maximum extent of the Laurentide Ice Sheet during the LGM (dashed line). Also shown are locations of sites mentioned in the text (*L-B moraine* - Littleton-Bethlehem moraine; *PR moraine* - Pineo Ridge moraine; *BB moraine* - Buzzards Bay moraine; *NH* - New Hampshire; *NB* - New Brunswick), as well as prominent moraines corresponding to deglacial ice-marginal positions (*CH* - Chicopee Readvance moraines (see Ridge, 2004); *LD* - Ledyard moraines, *OS* - Old Saybrook moraines (see Balco and Schaefer, 2006); *PM* - Perry Mountain moraine (see Balco *et al*., 2009)).

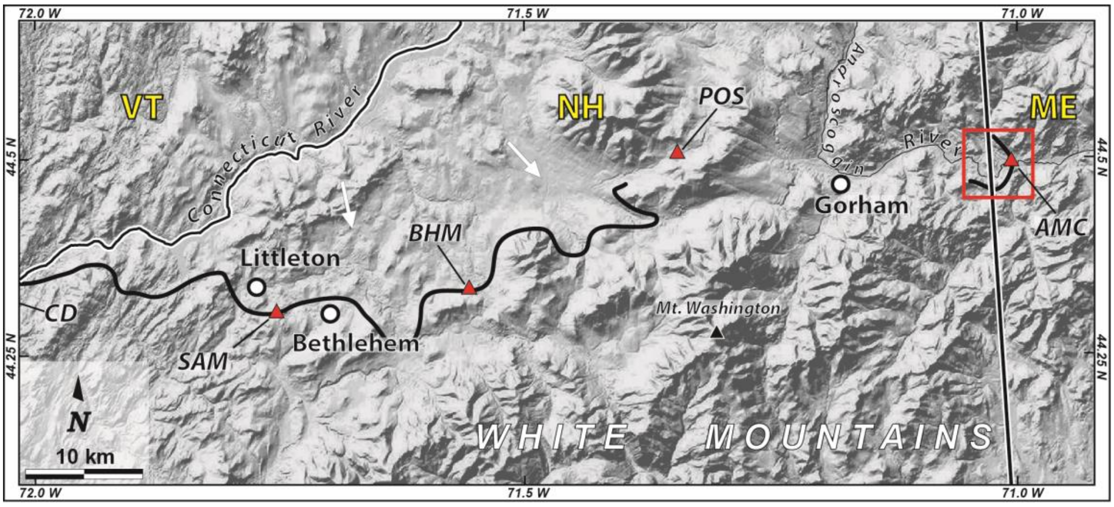

Figure 2 Topographic relief map of central New Hampshire and westernmost Maine showing locations sites discussed in the text. *AMC* - Androscoggin moraine complex; *SAM* - Sleeping Astronomer moraine; *BHM* - Beech Hill moraine; *CD* - Comerford Dam; *POS* - Pond of Safety. Thick black line represents the Littleton-Bethlehem moraine complex (adapted from Thompson *et al*., 1999: for more detailed maps of the BHM and SAM sections, we refer readers to that publication). White arrows indicate general ice-flow direction. Red square indicates area depicted in Figure 3.

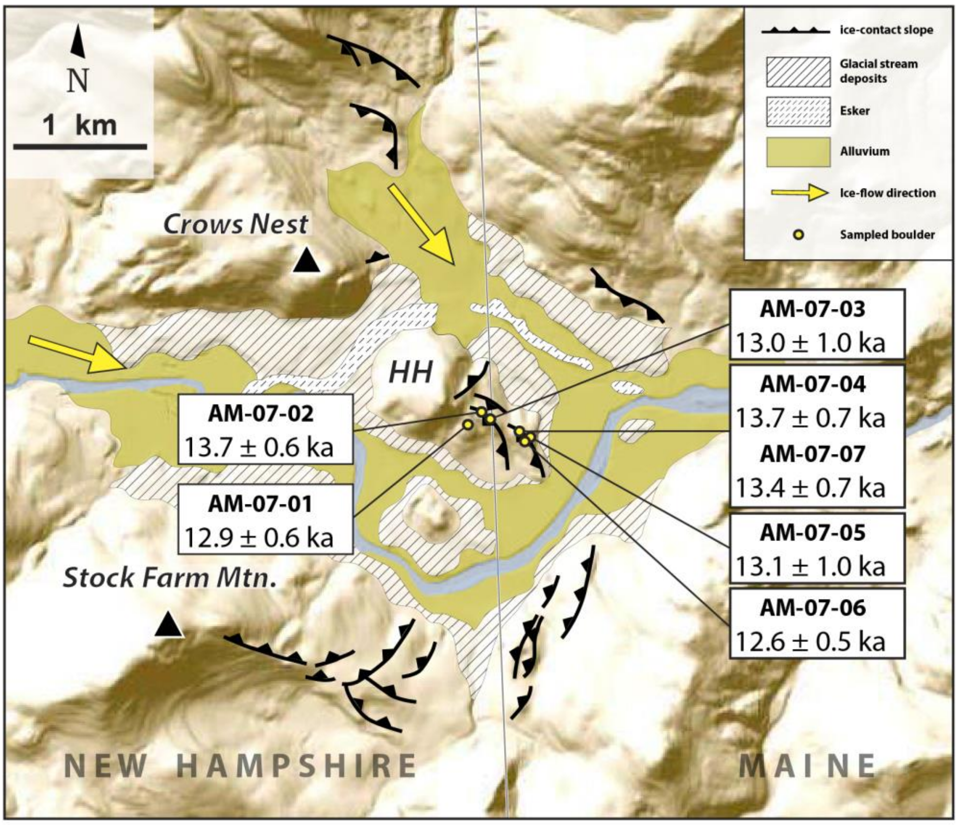

Figure 3 Geomorphic map of the Androscoggin moraine complex, showing the general distribution of surficial deposits (adapted from Thompson and Fowler, 1989), as well as locations and surface-exposure ages of samples from Hark Hill (*HH*). Non-specified areas comprise till and/or bedrock.

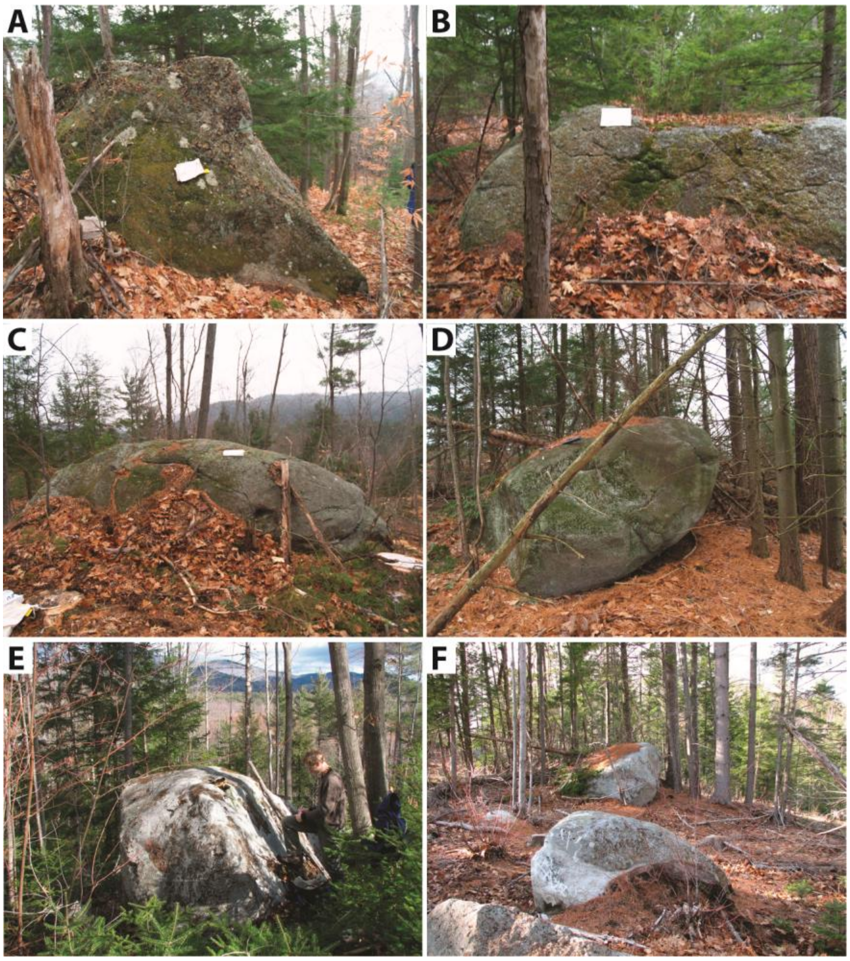

Figure 4 Glacially molded erratics on Hark Hill, Maine. Clockwise from top left: (A) AM-07-01; (B) AM-07-02; (C) AM-07-03; (D) AM-07-04; (E) AM-07-05; (F) AM-07-07 (foreground) and AM-07-04 (background). With the exception of AM-07-01, which is located on a bedrock knoll inside the moraine complex, all sampled boulders are located on well-defined moraine crests.

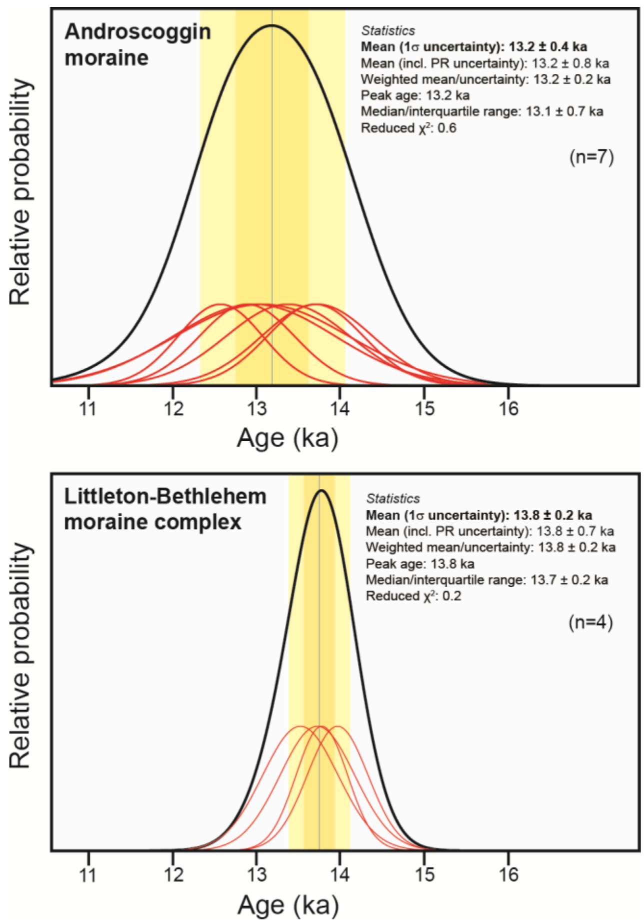

Figure 5 Age probability curves for the Androscoggin moraine and Littleton-Bethlehem moraine complex, showing mean and peak ages calculated using the NENA production rate (Balco *et al*., 2009) and time-independent St scaling (Lal, 1991; Stone, 2000). Vertical yellow bands depict the 1σ (dark yellow) and 2σ (light yellow) ranges of uncertainty for the mean ages. Reduced chi^2 (χ^2) values of < 1−2 reflect that age differences can be explained by analytical uncertainties alone (Bevington & Robinson, 1992).

Table 1 Sample details and ^{10}Be surface-exposure ages for the Androscoggin (AM) and Littleton-Bethlehem (LIT) moraine samples. Data for the LIT samples from Balco *et al*. (2009). All exposure ages shown are calculated using the NENA production rate of Balco *et al*. (2009) and St scaling (Lal, 1991; Stone, 2000). The L-B ages are identical if calculated with either the (independent) rate in Young *et al.* (2013) or Balco *et al.* (2009) (see text). Asterisk denotes the average ^{10}Be concentration of two replicates of sample 06-NE-013-LIT (Balco *et al*., 2009). For mean moraine age (arithmetic), we also present an error that includes propagation of the standard deviation as well as the uncertainty (4.8%) for the NENA production rate calibration.

Sample	Latitude	Longitude	Altitude (m)	Boulder height (m)	Thickness (cm)	Horizon correction	Quartz mass (g)	Carrier mass (mg)	$^{10}Be/^{9}Be$ (10^{-14})	^{10}Be conc. (10^{3} atoms/g)	Exposure age (yr)
AM-07-01	44°23.84'N	71°1.13'W	279	2	1.0	0.998	7.3058	0.1798	3.998 ± 0.15	65.37 ± 2.9	12,890 ± 570
AM-07-02	44°23.71'N	71°1.01'W	288	1.5	2.0	0.999	11.551	0.1797	6.653 ± 0.26	69.57 ± 2.9	13,700 ± 580
AM-07-03	44°23.83'N	71°0.97'W	283	1.5	2.1	0.999	5.510	0.1798	3.052 ± 0.21	65.54 ± 4.9	12,980 ± 970
AM-07-04	44°23.81'N	71°0.83'W	273	1.5	0.8	0.999	11.3217	0.1796	6.511 ± 0.30	69.41 ± 3.4	13,730 ± 670
AM-07-05	44°23.65'N	71°0.82'W	241	1.5	2.1	0.999	8.4227	0.1797	4.468 ± 0.34	63.53 ± 5.0	13,070 ± 1040
AM-07-06	44°23.66'N	71°0.88'W	244	1.5	2.4	0.999	12.3043	0.1796	6.238 ± 0.23	61.13 ± 2.3	12,570 ± 510
AM-07-07	44°23.69'N	71°0.86'W	241	1	1.8	0.999	11.8735	0.1796	6.413 ± 0.34	65.17 ± 3.6	13,380 ± 740
										Mean age (s.d.)	**13,190 ± 430**
										(with PR uncertainty propagated)	13,190 ± 770
06-NE-010-LIT	44°17.42'N	71°45.67'W	357	1.8	2	0.999	-	-	-	81.80 ± 2.6	13,730 ± 450
06-NE-011-LIT	44°17.42'N	71°45.65'W	357	1.9	2	0.999	-	-	-	80.60 ± 2.7	13,530 ± 450
06-NE-012-LIT	44°18.77'N	71°34.33'W	414	1.6	1	0.999	-	-	-	88.30 ± 2.3	13,970 ± 360
06-NE-013-LIT	44°18.88'N	71°34.38'W	412	1.6	10	0.999	-	-	-	81.00 ± 1.8*	13,780 ± 310
										Mean age (s.d.)	**13,750 ± 180**
										(with PR uncertainty propagated)	13,750 ± 690

Note: All AM samples were spiked with a 1024 mg/g ^{9}Be carrier. Two procedural blanks ($^{10}Be/^{9}Be$ = (0.491 and−1.835×10^{-15}), consisting of 0.180 ml of ^{9}Be carrier, were processed identically to the samples. Beryllium ratios of AM samples and blanks were measured relative to the 07KNSTD standard [$^{10}Be/^{9}Be$ = 2.85×10^{-12}], while those of L-B samples and blanks were measured relative to the KNSTD standard [$^{10}Be/^{9}Be$ = 3.15×10^{-12}]. Ages were calculated using a rock density of 2.65 g/cm^3 and assuming zero erosion.

Table 2 Comparison of ^{10}Be surface-exposure ages for the Androscoggin and Littleton-Bethlehem moraine samples generated using the NENA production rate and various scaling schemes: St - time-independent (Lal, 1991/Stone, 2000); Lm - time dependent (Lal, 1991/Stone, 2000); De (Desilets *et al*., 2006); Du (Dunai, 2001); Li (Lifton *et al*., 2005). All ages were calculated using the CRONUS-Earth online calculator, version 2.2 (Balco *et al*., 2008).

Sample	Exposure age: St (ka)	Exposure age: Lm (ka)	Exposure age: De (ka)	Exposure age: Du (ka)	Exposure age: Li (ka)
AM-07-01	12.9 ± 0.6	12.9 ± 0.6	13.0 ± 0.6	13.0 ± 0.6	13.0 ± 0.6
AM-07-02	13.7 ± 0.6	13.7 ± 0.6	13.8 ± 0.6	13.8 ± 0.6	13.8 ± 0.6
AM-07-03	13.0 ± 1.0	13.0 ± 1.0	13.0 ± 1.0	13.1 ± 1.0	13.1 ± 1.0
AM-07-04	13.7 ± 0.7	13.7 ± 0.7	13.8 ± 0.7	13.8 ± 0.7	13.8 ± 0.7
AM-07-05	13.1 ± 1.1	13.1 ± 1.1	13.1 ± 1.1	13.2 ± 1.1	13.2 ± 1.1
AM-07-06	12.6 ± 0.5	12.6 ± 0.5	12.6 ± 0.5	12.7 ± 0.5	12.7 ± 0.5
AM-07-07	13.4 ± 0.8	13.4 ± 0.8	13.4 ± 0.8	13.5 ± 0.8	13.5 ± 0.8
06-NE-010-LIT	13.7 ± 0.4	13.7 ± 0.5	13.8 ± 0.5	13.8 ± 0.5	13.8 ± 0.5
06-NE-011-LIT	13.5 ± 0.5	13.5 ± 0.5	13.6 ± 0.5	13.6 ± 0.5	13.6 ± 0.5
06-NE-012-LIT	14.0 ± 0.4	13.9 ± 0.4	14.0 ± 0.4	14.1 ± 0.4	14.0 ± 0.4
06-NE-013-LIT	13.8 ± 0.3	13.8 ± 0.3	13.8 ± 0.3	13.9 ± 0.3	13.9 ± 0.3

Table 3 Comparison of ^{10}Be surface-exposure ages for the Androscoggin and Littleton-Bethlehem moraine samples generated using several recent production rates: NENA (Balco *et al.*, 2009); Arctic (Young *et al.*, 2013); New Zealand (Putnam *et al.*, 2010). All ages were calculated using the CRONUS-Earth online calculator, version 2.2 (Balco *et al.*, 2008), and St scaling (Lal, 1991; Stone, 2000).

Sample	Exposure age: NENA (ka)	Exposure age: NZ (ka)	Exposure age: Arctic (ka)
AM-07-01	12.9 ± 0.6	13.2 ± 0.6	12.9 ± 0.6
AM-07-02	13.7 ± 0.6	14.0 ± 0.6	13.7 ± 0.6
AM-07-03	13.0 ± 1.0	13.3 ± 1.0	13.0 ± 1.0
AM-07-04	13.7 ± 0.7	14.0 ± 0.7	13.7 ± 0.7
AM-07-05	13.1 ± 1.0	13.4 ± 1.1	13.1 ± 1.0
AM-07-06	12.6 ± 0.5	12.9 ± 0.5	12.6 ± 0.5
AM-07-07	13.4 ± 0.7	13.7 ± 0.8	13.4 ± 0.7
06-NE-010-LIT	13.7 ± 0.4	14.1 ± 0.5	13.7 ± 0.4
06-NE-011-LIT	13.5 ± 0.5	13.8 ± 0.5	13.5 ± 0.4
06-NE-012-LIT	14.0 ± 0.4	14.3 ± 0.4	14.0 ± 0.4
06-NE-013-LIT	13.8 ± 0.3	14.1 ± 0.3	13.8 ± 0.3