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1 **Late glacial fluctuations of the Laurentide Ice Sheet in the White**  
2 **Mountains of Maine and New Hampshire, U.S.A.**

3

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16

17 **Abstract.** Prominent moraines deposited by the Laurentide Ice Sheet in northern New  
18 England document readvances, or stillstands, of the ice margin during overall  
19 deglaciation. However, until now, the paucity of direct chronologies over much of the  
20 region has precluded meaningful assessment of the mechanisms that drove these  
21 events, or of the complex relationships between ice-sheet dynamics and climate. As a  
22 step towards addressing this problem, we present a cosmogenic <sup>10</sup>Be surface-exposure  
23 chronology from the Androscoggin moraine complex, located in the White Mountains  
24 of western Maine and northern New Hampshire, as well as four recalculated ages  
25 from the nearby Littleton-Bethlehem moraine. Seven internally consistent <sup>10</sup>Be ages  
26 from the Androscoggin terminal moraines indicate that advance culminated  $\sim 13.2 \pm$

27 0.8 ka, in close agreement with the mean age of the neighboring Littleton-Bethlehem  
28 complex. Together, these two datasets indicate stabilization or advance of the ice-  
29 sheet margin in northern New England, at ~14-13 ka, during the Allerød/Greenland  
30 Interstadial I.

31

## 32 **Introduction**

33 During the last glacial maximum (LGM; ~26-19 ka), the Laurentide Ice Sheet (LIS)  
34 constituted the largest ice mass on Earth. With an area of ~13,000,000 km<sup>2</sup> and  
35 maximum surface elevations of as much as 3000 m (Denton and Hughes, 1981), the  
36 LIS exerted a considerable influence on global climate through albedo forcing (e.g.,  
37 Broccoli and Manabe, 1987), displacement of atmospheric circulation (e.g., Williams  
38 *et al.*, 1974; Hostetler and Clark, 1997; Oviatt, 1997; Birkel, 2010), and, particularly  
39 along the ice-sheet's Atlantic margins, ocean-ice sheet interactions (e.g., Bond *et al.*,  
40 1993; MacAyeal, 1993). Indeed, perturbations of the ice sheet's maritime sectors have  
41 been invoked widely as a key component – if not the cause – of Heinrich events (e.g.,  
42 Hulbe *et al.*, 2004), during which considerable volumes of ice apparently were  
43 disgorged into the North Atlantic Ocean with profound consequences for downwind  
44 ocean circulation (Broecker, 2003). Concurrently, the temperate nature and relatively  
45 low latitude of the ice sheet mean the LIS itself will have been highly sensitive to  
46 climate, particularly along southern and eastern margins where deglacial readvances  
47 have been correlated with perturbations of North Atlantic circulation (e.g., Lowell *et*  
48 *al.*, 1999; Dorion *et al.*, 2001; Borns *et al.*, 2004; Kaplan, 2007). Consequently,  
49 accurate reconstruction of past ice-sheet behavior provides the means to assess both  
50 the role of the LIS in Late Quaternary climate variability and sensitivity of the ice  
51 sheet to key events, such as Heinrich stadials and late-glacial climate reversals.

52           Comprehensive geologic mapping of glacial deposits and features over the  
53 past several decades has revealed the overall pattern of ice-sheet behavior in New  
54 England during the LGM and subsequent deglaciation. At its maximum extent, the  
55 southeastern sector of the LIS terminated at the large moraine systems off the coast of  
56 southern New England – Long Island, Martha’s Vineyard, and Nantucket Island (Fig.  
57 1; Oldale, 1982; Oldale and O’Hara, 1984; Stone and Borns, 1986; Balco *et al.*, 2002)  
58 – while farther east the ice margin lay far offshore at George’s Bank in the Gulf of  
59 Maine (Fig. 1; Pratt and Schlee, 1969; Stone and Borns, 1986). Inland from these  
60 limits, minor moraine belts and landform assemblages in southern New England  
61 document the northward retreat of the ice-sheet margin through Connecticut,  
62 Massachusetts, and into New Hampshire (Fig. 1; Koteff and Pessl, 1981; Ridge, 2004;  
63 Balco and Schaefer, 2006; Balco *et al.*, 2009). In Maine, swarms of minor moraines  
64 occur where deglaciation of the coastal lowland was accompanied by marine  
65 submergence. The distribution of glacial landforms in these areas indicates that ice-  
66 sheet retreat initially was gradual and oscillatory, with frequent stillstands and minor  
67 readvances resulting in the extensive (> 100 km in length) coastal moraine belt in  
68 Maine and New Brunswick (Borns, 1973, 1980; Borns and Hughes, 1977; Rampton *et*  
69 *al.*, 1984; Thompson and Borns, 1985; Smith and Hunter, 1989; Kaplan, 1999; Borns  
70 *et al.*, 2004). In a few places, the advances or stillstands were long enough to build  
71 extensive deposits, such as the Pond Ridge moraine and Pineo Ridge moraine  
72 complex (Borns *et al.*, 2004; Fig. 1). The scarcity of moraines and prevalence of  
73 eskers proximal to the coastal moraine belt indicates that subsequent deglaciation  
74 proceeded rapidly and without major interruption until the ice margin reached the  
75 vicinity of the Appalachian Mountains (Davis and Jacobson, 1985). There, several

76 extensive moraine complexes record fluctuations of this sector of the LIS and are the  
77 focus of this study.

78         Compared to extensive mapping studies, the glacial chronology for this SE  
79 sector of the LIS is less complete and, in places, controversial. On the basis of  
80 limiting radiocarbon ages and existing  $^{10}\text{Be}$  surface-exposure ages, we know that the  
81 ice sheet was most extensive during the LGM *sensu stricto* (Tucholke and Hollister,  
82 1973; Oldale, 1982; Stone and Borns, 1986; Stone *et al.*, 1998, 2002; Balco *et al.*,  
83 2002, 2009), consistent with other sectors of the LIS (Dyke and Prest, 1987).  
84 Recalculated  $^{10}\text{Be}$  ages from the Buzzards Bay moraine in Massachusetts (Fig. 1)  
85 suggest that the ice margin remained at or near its maximum extent until ~21 ka  
86 (Balco *et al.*, 2009). However, estimates for the onset of deglaciation vary widely,  
87 with varve (e.g., Ridge, 2004) and surface-exposure (Balco and Schaefer, 2006; Balco  
88 *et al.*, 2009) chronologies indicating a significantly earlier (by as much as ~5 ka)  
89 retreat of the LIS than that determined by recent terrestrial radiocarbon dating (e.g.,  
90 Peteet *et al.*, 2012). Nonetheless, radiocarbon ages of mollusk shells from raised  
91 marine deposits in eastern Maine indicate that the ice-sheet margin had retreated to  
92 the present-day coastline by ~17-15 ka, with the Pineo Ridge moraine complex  
93 representing a major stillstand (Smith and Hunter, 1989) and/or readvance (Borns,  
94 1980; Borns and Hughes, 1977) of the LIS towards the end of that period (e.g.,  
95 Kaplan, 1999, 2007; Dorion, 1997; Dorion *et al.*, 2001; Borns *et al.*, 2004).

96         Subsequent portions of the deglacial record are problematic. For example, the  
97 deglaciation of central and northern Maine is constrained by a relatively sparse  
98 coverage of minimum-limiting  $^{14}\text{C}$  ages, many of which are based on bulk samples  
99 and, in some cases, affected by a marine reservoir effect of uncertain magnitude  
100 (Davis and Jacobson, 1985; Ridge *et al.*, 2001; Borns *et al.*, 2004). In comparison, the

101 varve chronologies constructed for central New England (Antevs, 1922, 1928; Ridge  
102 *et al.*, 2001; Ridge, 2004) offer a seemingly precise chronology for ice-sheet retreat  
103 through that sector. However, as argued by Borns *et al.* (2004) and later by Peteet *et*  
104 *al.* (2012), these essentially "floating" chronologies are pinned to the calendar  
105 timescale in relatively few locations (see below), raising the possibility of systematic  
106 age offsets in the deglacial record. Thus, much of the chronology of ice-sheet retreat  
107 in New England is known only on a first-order basis.

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

119         To help develop the deglacial record of New England, specifically after the  
120 LIS had retreated inland of the marine limit, we present a  $^{10}\text{Be}$  surface-exposure  
121 chronology for the Androscoggin moraine complex, located in the White Mountains  
122 of western Maine and eastern New Hampshire (Figs. 1,2). Together with the  
123 neighboring Littleton-Bethlehem (L-B) moraine (Figs. 1,2; Thompson *et al.*, 1996,  
124 1999, 2002), the Androscoggin moraine represents renewed advance - or a stillstand -  
125 of the LIS in northern New England during overall deglaciation. However, until now

126 the Androscoggin moraine has not been dated directly. As a result, the significance of  
127 these prominent moraines to the deglacial record of the LIS hitherto was poorly  
128 understood. In this paper, we discuss the implications of our Androscoggin moraine  
129 dataset in conjunction with four recalculated  $^{10}\text{Be}$  ages from the L-B moraine.

130

### 131 **Geologic Setting and Methods**

132 The White Mountains of New Hampshire and western Maine contain a rich geologic  
133 record of LIS deglaciation that has been studied in detail since the mid-19<sup>th</sup> Century  
134 (e.g., Lyell, 1850; Agassiz, 1870; Thompson, 1999, and references therein). The most  
135 conspicuous depositional landforms include the L-B moraine system, a broad complex  
136 of ridges in the vicinity of the towns of Littleton and Bethlehem (Fig. 2; see  
137 Thompson *et al.*, 1999), and a group of prominent lateral and terminal moraines  
138 straddling the Maine-New Hampshire border in the Androscoggin River valley (Figs.  
139 2,3) that are the focus of this paper.

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

151 Previous work has established the age of the Androscoggin moraine complex  
152 only in a broad sense. A basal age of  $12,450 \pm 60$   $^{14}\text{C}$  yr BP (OS-7125)  
153 (14,985–14,219 cal yr BP (IntCal 13); Thompson *et al.*, 1996) from Pond of Safety  
154 (Fig. 2), located approximately 30 km north-west of the moraine in the upper Israel  
155 River valley, has been interpreted as a minimum age for the Androscoggin complex  
156 (Thompson *et al.*, 1999). However, owing to the challenging terrain and distances  
157 involved, the exact correlation between the two sites – and thus the relevance of the  
158 basal age to the moraine – has not been established unequivocally.

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

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



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

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

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

201 2.2 (Balco *et al.*, 2008), in conjunction with the NENA  $^{10}\text{Be}$  production rate (Balco *et*  
202 *al.*, 2009) and the time-independent "St" scaling (Lal, 1991; Stone, 2000). To provide  
203 an independent check of the NENA-derived ages, and to prevent circular reasoning  
204 (i.e., the L-B samples are a constituent of the NENA calibration), we also calculated  
205 ages using two recent, non-local  $^{10}\text{Be}$  production rates, one from north-eastern North  
206 America (Young *et al.*, 2013; see Table 3) and one from the Southern Alps of New  
207 Zealand (~44°S; Putnam *et al.*, 2010). In calculating moraine mean age, we also  
208 propagated analytical uncertainties with a systematic uncertainty (4.8%) associated  
209 with the NENA production rate calibration (Table 1). Although inclusion of such a  
210 systematic uncertainty is not needed for comparison between the closely located L-B  
211 and Androscoggin moraines (e.g., Fig. 5), it is a consideration when comparing to  
212 records determined by other chronological methods.

213 Our choice of St scaling reflects the close match this scheme provides to  
214 independent chronological constraints in New England (Balco *et al.*, 2009) and  
215 elsewhere (Fenton *et al.*, 2011; Briner *et al.*, 2012; Kelly *et al.*, 2013; Young *et al.*,  
216 2013). However, we stress that our conclusions are independent of our choice of  
217 scaling scheme, because at these latitudes and relatively low elevations the respective  
218 ages are statistically identical (Table 2). Analytical results and ages are given in Table  
219 1.

220

## 221 **Results**

222 The seven  $^{10}\text{Be}$  ages from the Androscoggin moraine – constituting the first surface  
223 exposure-dated moraine from the state of Maine – are given in Table 1 and are shown  
224 in Figures 3 and 5. The four recalculated surface-exposure ages from the L-B moraine  
225 are given in Table 1 and Figure 5. Together, the seven Androscoggin ages form an

226 internally consistent grouping, without outliers (i.e.,  $\pm 1\sigma$ ), ranging from  $13.7 \pm 0.6$  ka  
227 to  $12.6 \pm 0.5$  ka, and provide an average moraine age of  $13.2 \pm 0.4$  ka, or  $13.2 \pm 0.8$   
228 ka if a ~5% systematic uncertainty is propagated. Plotted as an age-probability curve,  
229 the ages exhibit a normal distribution with a peak age of 13.2 ka (Fig. 5). In  
230 comparison, the four samples collected from the L-B moraine by Balco *et al.* (2009)  
231 range from  $14.0 \pm 0.4$  ka to  $13.5 \pm 0.5$  ka, and provide a mean moraine age of  $13.8 \pm$   
232  $0.2$  ka (or  $13.8 \pm 0.7$  ka with a 5% propagated uncertainty) and a peak age of 13.8 ka  
233 (Fig. 5). As the three production rates (Balco *et al.*, 2009; Putnam *et al.*, 2010; Young  
234 *et al.*, 2013) are statistically identical, our moraine ages are consistent regardless of  
235 which rate is used. Moreover, both of the two more precise rates (Putnam *et al.*, 2010;  
236 Young *et al.*, 2013) produce exposure ages for L-B samples that are congruent with  
237 that inferred from the New England varve chronology (Ridge *et al.*, 1999; Balco *et*  
238 *al.*, 2009), supporting their use throughout eastern North America. We also point out  
239 that the very close agreement between the production rate of Young *et al.* (2013) and  
240 the New England varve chronology implies our propagation of a 5% systematic  
241 uncertainty is conservative.

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

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

257

## 258 **Discussion**

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

271 Although the peak ages of the two moraine complexes differ by  $\sim 600$  yr, their  
272 respective mean ages overlap within  $1\sigma$  uncertainty and thus the datasets are  
273 indistinguishable statistically. Therefore, we suggest that the Androscoggin and L-B  
274 moraine systems represent the same late-glacial advance/stillstand of the LIS, as  
275 hypothesized originally by Thompson *et al.* (2007). Viewed in greater detail, the small

276 discrepancy in peak age potentially reflects differences in the distribution of moraines  
277 and sampled boulders between the two sites. For instance, the sprawling geographic  
278 distribution of the L-B complex (Upham, 1904; Thompson *et al.*, 1999) indicates this  
279 moraine belt was deposited over a considerable period, while the four existing L-B  
280  $^{10}\text{Be}$  ages, from the Sleeping Astronomer and Beech Hill moraines, constrain the age  
281 of only the most distal parts of the complex. In contrast, the Androscoggin moraines,  
282 which are primarily composite landforms, exhibit a more compact spatial distribution.  
283 Our seven samples, collected from the crests of those composite moraines, thus  
284 represent the most recent period of deposition. Therefore, it is conceivable that the  
285 marginally younger peak age of the Androscoggin moraines correlates with the more  
286 proximal, as yet undated deposits in the L-B complex. Nonetheless, given the  
287 statistically indistinguishable ages of the two moraine complexes, we acknowledge  
288 that we are unable to test this model with the existing dataset.

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

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

312            At first glance, this disparity between glacial activity in northern New England  
313 and the widely accepted late-glacial temperature record for the North Atlantic region  
314 represents an ostensible paradox, in terms of the assumed climatic conditions during  
315 the YD stadial. Indeed, within New England, organic-poor, minerogenic sediments  
316 dating to the late glacial have been documented at several sites in Maine and the  
317 White Mountains and are interpreted as a return to stadial conditions during the YD  
318 (e.g., Dorion, 1997; Borns *et al.*, 2004; Dieffenbacher-Krall and Nurse, 2006).  
319 Nonetheless, we note that this pattern of Allerød advance–YD retreat has been  
320 documented elsewhere in the Northern Hemisphere, including at sites immediately  
321 adjacent the North Atlantic Ocean where YD cooling apparently was greatest (Bowen,  
322 1999; Lohne *et al.*, 2007; Bromley *et al.*, 2014). Moreover, our New England record  
323 is in accord with evidence for mild YD summertime temperatures in Greenland  
324 (Björck *et al.*, 2002) and with the absence of YD moraines (e.g., Mangerud and  
325 Landvik, 2007; Hall *et al.*, 2008; Kelly *et al.*, 2008; Briner *et al.*, 2009) or stratigraphy

326 (e.g., Wolfe, 1996) at many northern high-latitude sites. Such late-glacial behavior  
327 potentially points to the effects of extreme seasonality adjacent the North Atlantic  
328 during the YD (Denton *et al.*, 2005), with warm summers driving glacial recession  
329 and very cold winters causing the depressed mean-annual temperatures recorded by  
330 ice-core proxies. If true, the sedimentologic signature of the YD recorded in New  
331 England lakes potentially reflects periglacial remobilization of slope deposits rather  
332 than year-round cooling (Putnam and Putnam, 2009).

333         Finally, our new chronology supports the findings of Richard and Occhietti  
334 (2005) that late-glacial marine transgression of the St. Lawrence lowlands occurred  
335 later than indicated by basal radiocarbon ages from the former Champlain Sea  
336 (Rodrigues, 1992). Construction of the L-B and Androscoggin moraines would have  
337 required the persistence of a robust ice-sheet margin in northern New England at least  
338 as late as ~13.2 ka, a scenario that is supported by terrestrial radiocarbon data from  
339 elsewhere in this region (see Thompson *et al.*, 1999; Borns *et al.*, 2004) but which  
340 may be incompatible with ice-free conditions in the St. Lawrence River valley  
341 immediately north of New Hampshire as early as ~14 ka (Parent and Occhietti, 1988).  
342 This discrepancy is removed, however, if the radiocarbon dates upon which the age of  
343 the Champlain Sea is based incorporate a considerable reservoir effect, as suggested  
344 by Rodrigues (1992) and demonstrated by Richard and Occhietti (2005). Thompson *et*  
345 *al.* (2011) found evidence of a 1000-yr reservoir at a site with juxtaposed marine and  
346 terrestrial organics, located in Portland, Maine.

347

## 348 **Conclusions**

349 Our cosmogenic  $^{10}\text{Be}$  surface-exposure chronology from the Androscoggin moraine,  
350 coupled with four recalculated ages from the L-B complex, affords robust and directly

351 dated evidence for a major late-glacial fluctuation of the LIS in northern New  
352 England. Retreat of the ice sheet north of the White Mountains was interrupted by at  
353 least one advance during the Allerød, resulting in deposition of the prominent L-B and  
354 Androscoggin complexes. Seven ages from the terminal moraines of the  
355 Androscoggin complex indicate deposition  $\sim 13.2 \pm 0.8$  ka, broadly coincident with  
356 the emplacement of the Sleeping Astronomer and Beech Hill moraines in the L-B  
357 complex, leading us to conclude that the two moraine belts represent the same late-  
358 glacial climate reversal. Ultimately, our chronology indicates that the LIS in northern  
359 New England underwent readvance, or stillstand, during the Allerød. These findings  
360 have important implications for our understanding both of ice-sheet behavior in New  
361 England during the last glacial termination and the terrestrial impact of deglacial  
362 climate changes, including cold stadials, in the North Atlantic region.

363

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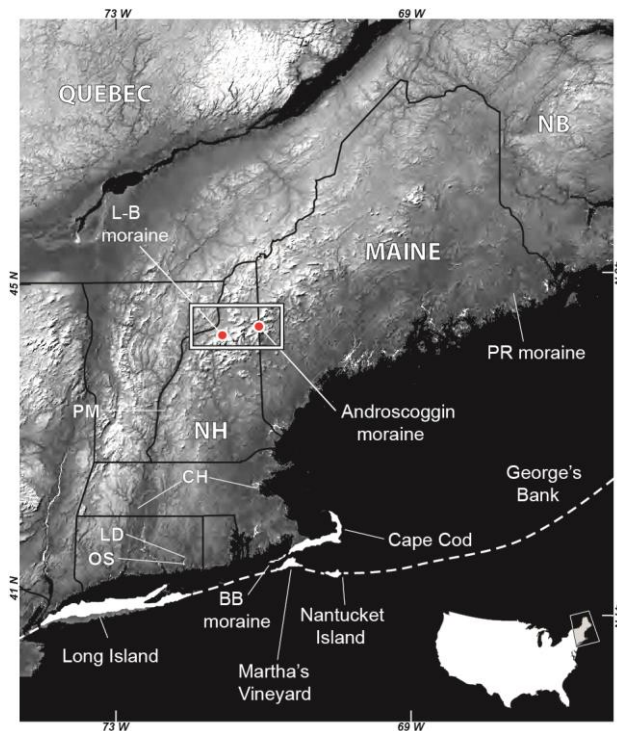
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643 **Figure and Table Captions**



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

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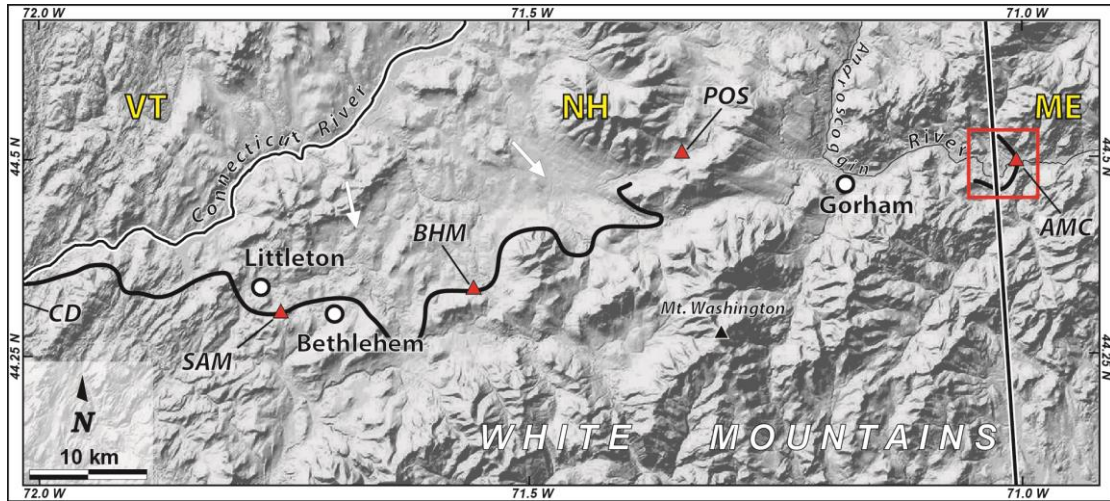
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660 **Figure 2** Topographic relief map of central New Hampshire and westernmost Maine  
 661 showing locations sites discussed in the text. *AMC* - Androscoggin moraine complex;  
 662 *SAM* - Sleeping Astronomer moraine; *BHM* - Beech Hill moraine;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

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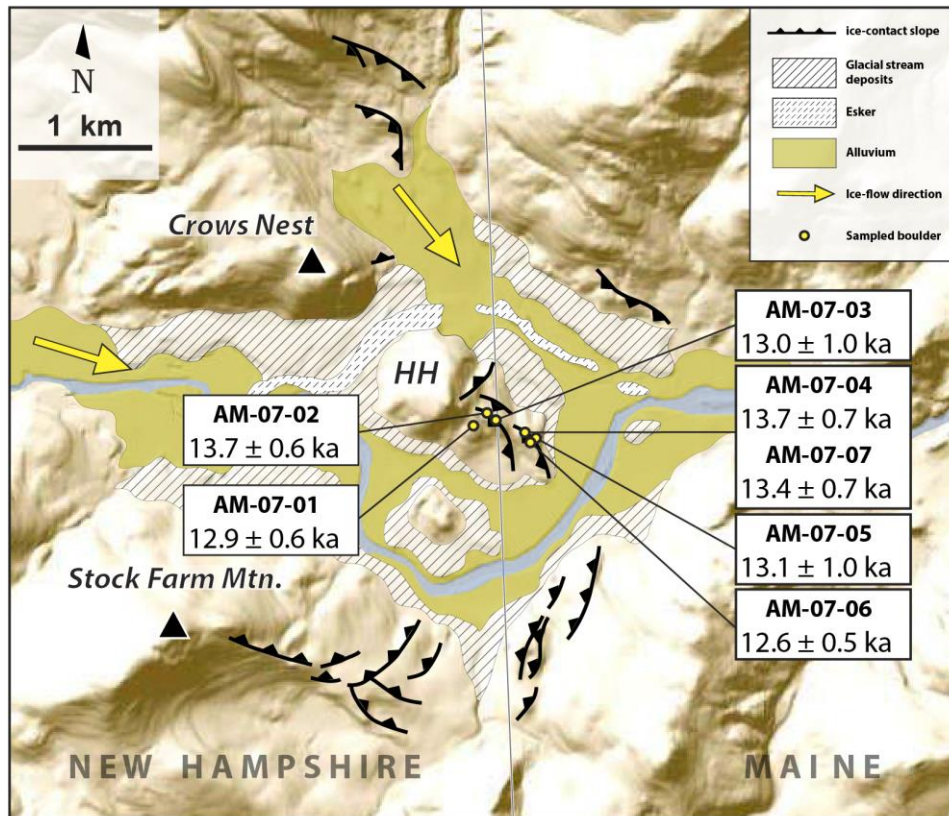
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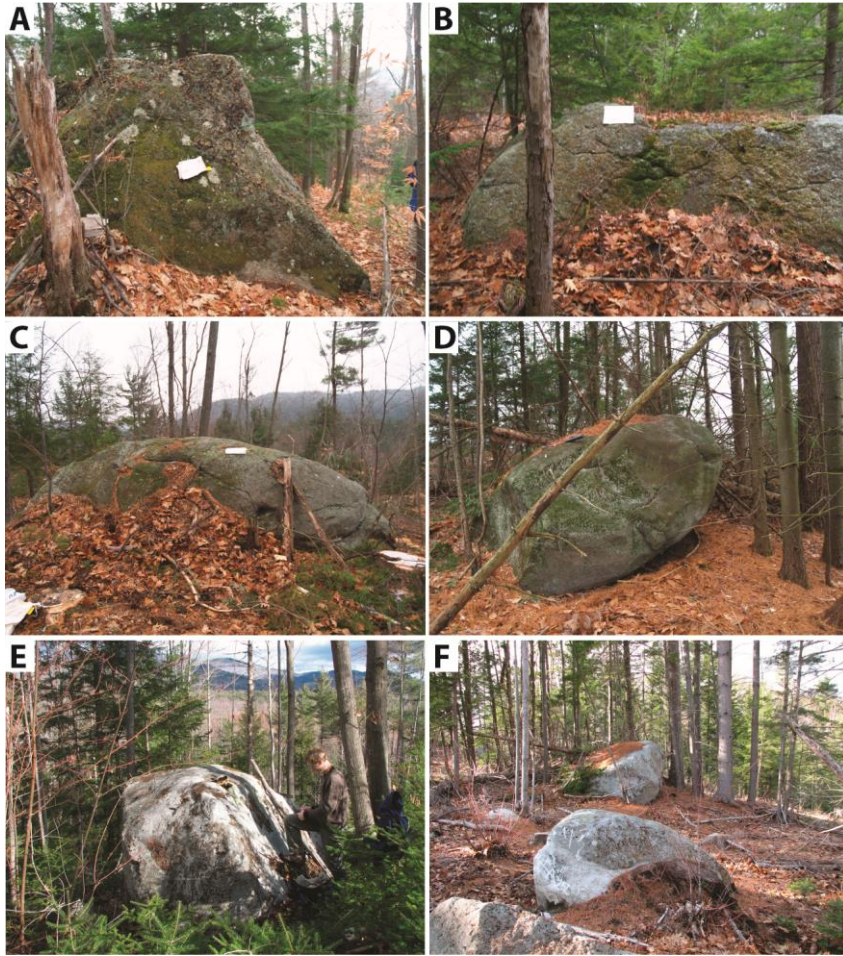


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679 **Figure 3** Geomorphic map of the Androscoggin moraine complex, showing the  
 680 general distribution of surficial deposits (adapted from Thompson and Fowler, 1989),  
 681 as well as locations and surface-exposure ages of samples from Hark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

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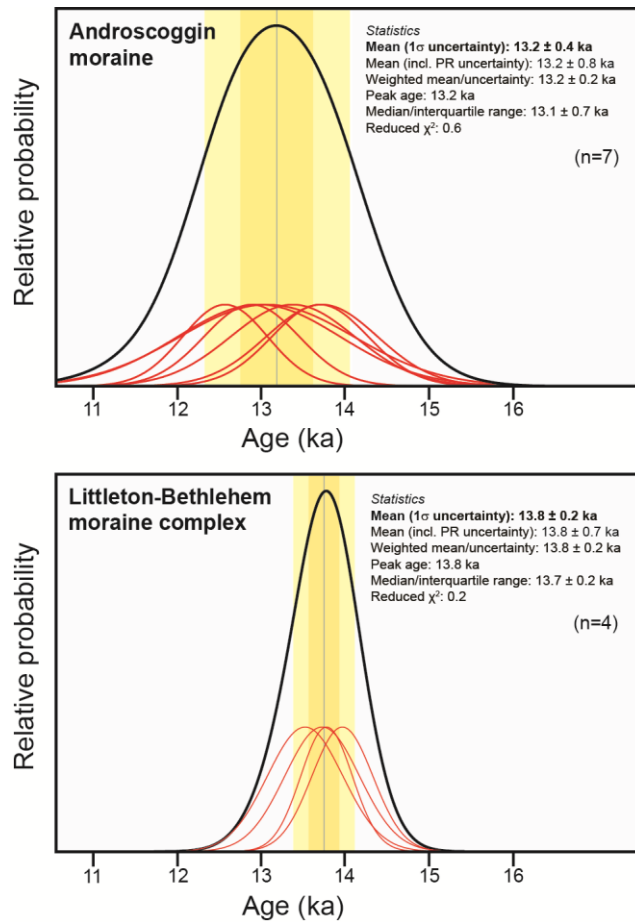




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685 **Figure 4** Glacially molded erratics on Hark Hill, Maine. Clockwise from top left: (A)  
 686 AM-07-01; (B) AM-07-02; (C) AM-07-03; (D) AM-07-04; (E) AM-07-05; (F) AM-  
 687 07-07 (foreground) and AM-07-04 (background). With the exception of AM-07-01,  
 688 which is located on a bedrock knoll inside the moraine complex, all sampled boulders  
 689 are located on well-defined moraine crests.

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692 **Figure 5** Age probability curves for the Androscoggin moraine and Littleton-  
 693 Bethlehem moraine complex, showing mean and peak ages calculated using the  
 694 NENA production rate (Balco *et al.*, 2009) and time-independent St scaling (Lal,  
 695 1991; Stone, 2000). Vertical yellow bands depict the 1 $\sigma$  (dark yellow) and 2 $\sigma$  (light  
 696 yellow) ranges of uncertainty for the mean ages. Reduced  $\chi^2$  ( $\chi^2$ ) values of < 1–2  
 697 reflect that age differences can be explained by analytical uncertainties alone  
 698 (Bevington & Robinson, 1992).

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

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

**Table 2** Comparison of  $^{10}\text{Be}$  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	$12.9 \pm 0.6$	$12.9 \pm 0.6$	$13.0 \pm 0.6$	$13.0 \pm 0.6$	$13.0 \pm 0.6$
AM-07-02	$13.7 \pm 0.6$	$13.7 \pm 0.6$	$13.8 \pm 0.6$	$13.8 \pm 0.6$	$13.8 \pm 0.6$
AM-07-03	$13.0 \pm 1.0$	$13.0 \pm 1.0$	$13.0 \pm 1.0$	$13.1 \pm 1.0$	$13.1 \pm 1.0$
AM-07-04	$13.7 \pm 0.7$	$13.7 \pm 0.7$	$13.8 \pm 0.7$	$13.8 \pm 0.7$	$13.8 \pm 0.7$
AM-07-05	$13.1 \pm 1.1$	$13.1 \pm 1.1$	$13.1 \pm 1.1$	$13.2 \pm 1.1$	$13.2 \pm 1.1$
AM-07-06	$12.6 \pm 0.5$	$12.6 \pm 0.5$	$12.6 \pm 0.5$	$12.7 \pm 0.5$	$12.7 \pm 0.5$
AM-07-07	$13.4 \pm 0.8$	$13.4 \pm 0.8$	$13.4 \pm 0.8$	$13.5 \pm 0.8$	$13.5 \pm 0.8$
06-NE-010-LIT	$13.7 \pm 0.4$	$13.7 \pm 0.5$	$13.8 \pm 0.5$	$13.8 \pm 0.5$	$13.8 \pm 0.5$
06-NE-011-LIT	$13.5 \pm 0.5$	$13.5 \pm 0.5$	$13.6 \pm 0.5$	$13.6 \pm 0.5$	$13.6 \pm 0.5$
06-NE-012-LIT	$14.0 \pm 0.4$	$13.9 \pm 0.4$	$14.0 \pm 0.4$	$14.1 \pm 0.4$	$14.0 \pm 0.4$
06-NE-013-LIT	$13.8 \pm 0.3$	$13.8 \pm 0.3$	$13.8 \pm 0.3$	$13.9 \pm 0.3$	$13.9 \pm 0.3$

**Table 3** Comparison of  $^{10}\text{Be}$  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	$12.9 \pm 0.6$	$13.2 \pm 0.6$	$12.9 \pm 0.6$
AM-07-02	$13.7 \pm 0.6$	$14.0 \pm 0.6$	$13.7 \pm 0.6$
AM-07-03	$13.0 \pm 1.0$	$13.3 \pm 1.0$	$13.0 \pm 1.0$
AM-07-04	$13.7 \pm 0.7$	$14.0 \pm 0.7$	$13.7 \pm 0.7$
AM-07-05	$13.1 \pm 1.0$	$13.4 \pm 1.1$	$13.1 \pm 1.0$
AM-07-06	$12.6 \pm 0.5$	$12.9 \pm 0.5$	$12.6 \pm 0.5$
AM-07-07	$13.4 \pm 0.7$	$13.7 \pm 0.8$	$13.4 \pm 0.7$
06-NE-010-LIT	$13.7 \pm 0.4$	$14.1 \pm 0.5$	$13.7 \pm 0.4$
06-NE-011-LIT	$13.5 \pm 0.5$	$13.8 \pm 0.5$	$13.5 \pm 0.4$
06-NE-012-LIT	$14.0 \pm 0.4$	$14.3 \pm 0.4$	$14.0 \pm 0.4$
06-NE-013-LIT	$13.8 \pm 0.3$	$14.1 \pm 0.3$	$13.8 \pm 0.3$