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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 ~13.2 $\pm$

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 icesheet margin in northern New England, at ~14-13 ka, during the Allerød/Greenland
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

extensive moraine complexes record fluctuations of this sector of the LIS and are thefocus 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 limiting radiocarbon ages and existing <sup>10</sup>Be surface-exposure ages, we know that the 80 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). Recalculated <sup>10</sup>Be ages from the Buzzards Bay moraine in Massachusetts (Fig. 1) 84 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 coverage of minimum-limiting <sup>14</sup>C ages, many of which are based on bulk samples 98 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

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.

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 LIS had retreated inland of the marine limit, we present a <sup>10</sup>Be surface-exposure 120

- 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 <sup>10</sup>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

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

regional provenance (Fig. 4), the majority of which are quartz-bearing and thus ideal

150 for <sup>10</sup>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 $^{14}$ 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 $\pm250$ yr to the complex as part of the north-east North
171	America (NENA) <sup>10</sup> 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).

Our Androscoggin surface-exposure chronology is based on <sup>10</sup>Be ages (n=7) 176 177 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 closest well-preserved moraines to the maximum terminal position of the 181 182 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 standard  $[^{10}\text{Be}/^9\text{Be} = 2.85 \times 10^{-12}$ ; Nishiizumi *et al.*, 2007] (note: ratios for the L-B 198 samples were measured relative to the KNSTD standard [ ${}^{10}Be/{}^{9}Be = 3.15 \times 10^{-12}$ ]). We 199

200 calculated surface-exposure ages using the CRONUS-Earth online calculator, version

2.2 (Balco et al., 2008), in conjunction with the NENA <sup>10</sup>Be production rate (Balco et 201 202 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 ages using two recent, non-local <sup>10</sup>Be production rates, one from north-eastern North 205 206 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**

The seven <sup>10</sup>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

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, the ages exhibit a normal distribution with a peak age of 13.2 ka (Fig. 5). In 229 230 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. Second, comparison of <sup>10</sup>Be age versus boulder height shows no significant 247 248 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

Androscoggin and L-B data sets are < 1; values of  $\sim 1-2$  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 postdepositional 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 ~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 <sup>10</sup>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  ${}^{14}C$ 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

(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).

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 <sup>10</sup>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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372

373

374

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# 643 **Figure and Table Captions**





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 -



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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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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.



- **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
- are located on well-defined moraine crests.
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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,

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<sup>2</sup> ( $\chi^2$ ) 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 <sup>10</sup>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 <sup>10</sup>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)	<sup>10</sup> Be/ <sup>9</sup> Be (10 <sup>-14</sup> )	$^{10}$ Be conc. (10 <sup>3</sup> 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	$6.653 \pm 0.26$	$69.57\pm2.9$	$13,\!700\pm580$
AM-07-03	44°23.83'N	71°0.97'W	283	1.5	2.1	0.999	5.510	0.1798	$3.052\pm0.21$	$65.54 \pm 4.9$	$12{,}980 \pm 970$
AM-07-04	44°23.81'N	71°0.83'W	273	1.5	0.8	0.999	11.3217	0.1796	$6.511 \pm 0.30$	$69.41 \pm 3.4$	$13,\!730\pm670$
AM-07-05	44°23.65'N	71°0.82'W	241	1.5	2.1	0.999	8.4227	0.1797	$4.468 \pm 0.34$	$63.53\pm5.0$	$13,070 \pm 1040$
AM-07-06	44°23.66'N	71°0.88'W	244	1.5	2.4	0.999	12.3043	0.1796	$6.238 \pm 0.23$	$61.13\pm2.3$	$12,\!570\pm510$
AM-07-07	44°23.69'N	71°0.86'W	241	1	1.8	0.999	11.8735	0.1796	$6.413 \pm 0.34$	$65.17 \pm 3.6$	$13,380 \pm 740$
									(wi	Mean age (s.d.) th PR uncertainty propagated)	<b>13,190 ± 430</b> 13,190 ± 770
06-NE-010- LIT	44°17.42'N	71°45.67'W	357	1.8	2	0.999	-	-	-	$81.80\pm2.6$	$13,\!730\pm\!450$
06-NE-011- LIT	44°17.42'N	71°45.65'W	357	1.9	2	0.999	-	-	-	$80.60\pm2.7$	$13,\!530\pm450$
06-NE-012- LIT	44°18.77'N	71°34.33'W	414	1.6	1	0.999	-	-	-	$88.30\pm2.3$	$13,970 \pm 360$
06-NE-013- LIT	44°18.88'N	71°34.38'W	412	1.6	10	0.999	-	-	-	$81.00\pm1.8*$	$13,\!780\pm310$
									(wi	Mean age (s.d.) th PR uncertainty propagated)	<b>13,750 ± 180</b> 13,750 ± 690

*Note:* All AM samples were spiked with a 1024 mg/g <sup>9</sup>Be carrier. Two procedural blanks ( ${}^{10}\text{Be}/{}^9\text{Be} = (0.491 \text{ and}-1.835\times10^{-15})$ , consisting of 0.180 ml of <sup>9</sup>Be carrier, were processed identically to the samples. Beryllium ratios of AM samples and blanks were measured relative to the 07KNSTD standard [ ${}^{10}\text{Be}/{}^9\text{Be} = 2.85\times10^{-12}$ ], while those of L-B samples and blanks were measured relative to the KNSTD standard [ ${}^{10}\text{Be}/{}^9\text{Be} = 3.15\times10^{-12}$ ]. Ages were calculated using a rock density of 2.65 g/cm<sup>3</sup> and assuming zero erosion.

**Table 2** Comparison of <sup>10</sup>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	$12.9\pm0.6$	$12.9\pm0.6$	$13.0\pm0.6$	$13.0\pm0.6$	$13.0\pm0.6$
AM-07-02	$13.7\pm0.6$	$13.7\pm0.6$	$13.8\pm0.6$	$13.8\pm0.6$	$13.8\pm0.6$
AM-07-03	$13.0\pm1.0$	$13.0\pm1.0$	$13.0\pm1.0$	$13.1\pm1.0$	$13.1\pm1.0$
AM-07-04	$13.7\pm0.7$	$13.7\pm0.7$	$13.8\pm0.7$	$13.8\pm0.7$	$13.8\pm0.7$
AM-07-05	$13.1 \pm 1.1$	$13.1 \pm 1.1$	$13.1\pm1.1$	$13.2\pm1.1$	$13.2\pm1.1$
AM-07-06	$12.6\pm0.5$	$12.6\pm0.5$	$12.6\pm0.5$	$12.7\pm0.5$	$12.7\pm0.5$
AM-07-07	$13.4\pm0.8$	$13.4\pm0.8$	$13.4\pm0.8$	$13.5\pm0.8$	$13.5\pm0.8$
06-NE-010-LIT	$13.7\pm0.4$	$13.7\pm0.5$	$13.8\pm0.5$	$13.8\pm0.5$	$13.8\pm0.5$
06-NE-011-LIT	$13.5\pm0.5$	$13.5\pm0.5$	$13.6\pm0.5$	$13.6\pm0.5$	$13.6\pm0.5$
06-NE-012-LIT	$14.0\pm0.4$	$13.9\pm0.4$	$14.0\pm0.4$	$14.1\pm0.4$	$14.0\pm0.4$
06-NE-013-LIT	$13.8\pm0.3$	$13.8\pm0.3$	$13.8\pm0.3$	$13.9\pm0.3$	$13.9\pm0.3$

**Table 3** Comparison of <sup>10</sup>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	$12.9\pm0.6$	$13.2\pm0.6$	$12.9\pm0.6$
AM-07-02	$13.7\pm0.6$	$14.0\pm0.6$	$13.7\pm0.6$
AM-07-03	$13.0\pm1.0$	$13.3\pm1.0$	$13.0\pm1.0$
AM-07-04	$13.7\pm0.7$	$14.0\pm0.7$	$13.7\pm0.7$
AM-07-05	$13.1 \pm 1.0$	$13.4 \pm 1.1$	$13.1\pm1.0$
AM-07-06	$12.6\pm0.5$	$12.9\pm0.5$	$12.6\pm0.5$
AM-07-07	$13.4\pm0.7$	$13.7\pm0.8$	$13.4\pm0.7$
06-NE-010-LIT	$13.7\pm0.4$	$14.1\pm0.5$	$13.7\pm0.4$
06-NE-011-LIT	$13.5\pm0.5$	$13.8\pm0.5$	$13.5 \pm 0.4$
06-NE-012-LIT	$14.0\pm0.4$	$14.3\pm0.4$	$14.0\pm0.4$
06-NE-013-LIT	$13.8 \pm 0.3$	$14.1\pm0.3$	$13.8\pm0.3$