



1 Ice sheet model simulations reveal polythermal ice conditions existed across the NE USA during
2 the Last Glacial Maximum

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12 **Abstract**

13 Geologic evidence of the Laurentide Ice Sheet (LIS) provides abundant constraints on the areal
14 extent of the ice sheet during the Last Glacial Maximum (LGM). Direct observations of LGM LIS
15 thickness are non-existent, however, with most geologic data across high elevation summits in the
16 Northeastern United States (NE USA) often showing signs of inheritance, indicative of weakly
17 erosive ice flow and the presence of cold-based ice. While warm-based ice and erosive conditions
18 likely existed on the flanks of these summits and throughout neighboring valleys, summit
19 inheritance issues have hampered efforts to constrain the timing of the emergence of ice-free
20 conditions at high elevation summits. These geomorphic reconstructions indicate that a complex
21 erosional and thermal regime likely existed across the southeasternmost extent of the LIS
22 sometime during the LGM, although this has not been confirmed by ice sheet models. Instead,
23 current ice sheet models simulate warm-based ice conditions across this region, with disagreement
24 likely arising from the use of low resolution meshes (e.g., >20 km) which are unable to resolve the
25 high bedrock relief across this region that strongly influenced overall ice flow and the complex
26 LIS thermal state. Here we use a newer generation ice sheet model, the Ice-sheet and Sea-level
27 System Model (ISSM), to simulate the LGM conditions of the LIS across the NE USA and at 3
28 localities with high bedrock relief (Adirondack Mountains, White Mountains, and Mount
29 Katahdin), with results confirming the existence of a complex thermal regime as interpreted by the
30 geologic data. The model uses higher-order physics, a small ensemble of LGM climate boundary
31 conditions, and a high-resolution horizontal mesh that resolves bedrock features down to 30 meters
32 to reconstruct LGM ice flow, ice thickness, and thermal conditions. These results indicate that
33 across the NE USA, polythermal conditions existed during the LGM. While the majority of this
34 domain is simulated to be warm-based, cold-based ice persists where ice velocities are slow (<15
35 m/yr) particularly across regional ice divides (e.g., Adirondacks). Additionally, sharp thermal
36 boundaries are simulated where cold-based ice across high elevation summits (White Mountains
37 and Mount Katahdin) flank warm-based ice in adjacent valleys. Because geologic data is
38 geographically limited, these high-resolution simulations can help fill gaps in our understanding
39 of the geographical distribution of the polythermal ice during the LGM. We find that the elevation
40 of this simulated thermal boundary ranges between 800-1500 meters, largely supporting geologic
41 interpretations that polythermal ice conditions existed across NE USA during the LGM, however
42 this boundary varies geographically. In general, we show that a model with finer spatial resolution
43 and higher order physics is able to simulate the polythermal conditions captured in the geologic
44 data, with model output being of potential utility for site selection in future geologic studies and
45 geomorphic interpretation of landscape evolution.



46 **1. Introduction**

47 During the Last Glacial Maximum (26.5 to 19.0 ka; Clark et al., 2009), global temperatures
48 cooled by ~6.1°C (Tierney et al., 2020) leading to the growth of expansive ice sheets and the
49 lowering of global sea level by ~130 m (Clark & Mix, 2002). As part of the North American Ice
50 Sheet complex (NAIS), the Laurentide Ice Sheet (LIS) is estimated to have contained 75-85 m
51 global sea-level equivalent (SLE; Clark & Mix, 2002) thus representing a climatically important
52 component of the cryosphere. Extending southwards from its source region in northern Canada,
53 the LIS covered most of the northeastern United States (NE USA; Fig. 1) with its terminal position
54 located along Martha's Vineyard, Massachusetts (MA), Long Island, New York (NY), and into
55 northern New Jersey and Pennsylvania to the west (Dalton et al., 2020). The retreat of the LIS
56 during the last deglaciation is constrained through numerous geochronologic studies including:
57 varve chronologies (Ridge et al., 2012), basal radiocarbon dates (Fig. 1d; Dyke et al., 2004; Dalton
58 et al., 2020), and terrestrial in-situ cosmogenic nuclide surface exposure ages of moraines (Balco
59 and Schaefer, 2006; Bromley et al., 2015; Ullman et al., 2016; Hall et al., 2017; Bromley et al.,
60 2020; Balter-Kennedy et al., 2024). While these geologic archives constrain ice margin retreat
61 well, the vertical thinning history and ultimately the volumetric change of the LIS during the last
62 deglaciation in this region remains poorly known.

63 Fortunately, because of the high vertical relief across the NE USA, studies have addressed
64 the vertical thinning history of the LIS in this region by dating glacial features along vertical
65 transects (herein referred to as dipstick studies; Bierman et al., 2015; Koester et al., 2017; Barth et
66 al., 2019; Corbett et al., 2019; Koester et al., 2020; Halsted et al., 2023). These studies indicate
67 that rapid vertical ice sheet thinning occurred coincident with ice margin retreat during the last
68 deglaciation, and predominantly during the Bølling-Allerød warm period. Through surface
69 exposure dating of bedrock features and glacial erratics, these dipstick studies commonly find the
70 presence of inherited nuclides across high elevation sites (>1200 m; Halsted et al., 2023), making
71 geologic interpretations of the onset of vertical ice thinning at these locations difficult.
72 Consequently, these data suggest that the high elevation regions of the NE USA were likely
73 covered by cold-based ice characterized by the absence of subglacial water and ultimately much
74 reduced subglacial erosion, with warm-based ice and erosive conditions flanking the valleys of
75 these high elevation regions (Halsted et al., 2023). While these geologic interpretations support
76 the existence of polythermal ice conditions across the NE USA, it is not well known how this
77 subglacial regime varied spatially and whether the existence of this boundary occurred along a
78 geographically consistent elevation. This has implications for interpreting geologic data of past
79 ice sheet retreat or thinning particularly at high elevations, as well as erosional processes that may
80 have operated during glaciation across the NE USA, as erosional patterns are closely related to the
81 ice sheet thermal regime (Lai and Anders et al., 2021). Ultimately, where these geologic archives
82 are limited in their spatial coverage, ice sheet models can be used to simulate broader
83 characteristics of this thermal regime.

84 Ice sheet models have been important tools to study LIS conditions during the LGM and
85 assess the drivers of deglacial change (Sugden, 1977; Marshall et al., 2000; Hooke and Fastook,
86 2007; Tarasov and Peltier, 2007; Gregoire et al., 2012; Moreno et al., 2023). Not only can ice
87 sheet models aid in the interpretation of geologic proxies of past ice sheet behavior, but they can
88 provide outputs that enable a more informed choice when considering field locations for sampling.
89 For example, a recent assessment of the basal thermal state of the Greenland Ice Sheet (MacGregor
90 et al., 2016; 2022), which in part relied upon output from newer generation ice sheet models, was
91 used to inform field campaigns aimed at sampling subglacial bedrock in portions of the Greenland



92 Ice Sheet that were estimated to be cold-based with low erosion (Briner et al., 2022). In regards
93 to the LIS, prior ice sheet modeling efforts were useful in identifying a broad picture of the thermal
94 behavior of the LIS, which indicated that at the LGM roughly 20-50% of the LIS was warm based
95 (Marshall and Clark, 2002; Tarasov and Peltier, 2007), including the NE USA. Some aspects of
96 these simulations (Tarasov and Peltier, 2007) do agree well with broad scale geomorphic indicators
97 of basal conditions (Klemen and Hattestrand, 1999; Briner et al., 2006; Klemen and Glasser, 2007;
98 Briner et al., 2014), which indicate that the LIS exhibited a varying subglacial thermal regime,
99 with frozen bed patches interspersed particularly along ice divides and in some cases residing
100 along sharp boundaries with warm-based ice streams or outlet glaciers. However, due to the low
101 spatial resolution of existing models (>20 km), the high topographic relief across the NE USA is
102 poorly resolved and therefore the sharp thermal boundary between cold and warm-based ice
103 identified in the geologic archives is not captured. This is likely due to the more coarsely resolved
104 models' inability to capture advective and diffusive processes at these small spatial scales.
105 Likewise, as the high relief of this region served as a control and impediment to ice flow during
106 the LGM, the impacts of ice flow on deformational and frictional heating is more poorly captured
107 in lower resolution models. This can be improved by downscaling ice sheet models to higher
108 resolution, which has shown promise (Staiger et al., 2005) in interpreting how polythermal ice
109 conditions may have influenced regional glacial geomorphology.

110 To address this shortcoming, we use a high-resolution ice sheet model to simulate the
111 thermal regime of the LIS across the NE USA during the LGM. Our model experiments are
112 designed to test whether the presence of this sharp thermal regime is simulated and to assess the
113 spatial and elevational characteristics of the basal thermal regime at regional and local scales (i.e.,
114 mountain range) where geologic evidence suggests the existence of polythermal conditions.
115 Through this work, we aim to support current geologic interpretations, while also filling gaps in
116 our understanding of this complex thermal regime where geologic constraints are limited.

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119 **2. Methods**

120 The NE USA, comprising of the states listed in Figure 1, is marked by high topographic
121 relief that spans an elevational range from sea-level to >1500 meters. In order to capture these
122 large gradients in bedrock topography and thermal boundaries within the ice sheet, we employ a
123 nested modelling approach (Briner et al., 2020; Cuzzone et al., 2022), whereby a more coarsely
124 resolved and larger model provides boundary conditions necessary for downscaling over regional
125 and local scale domains that are more finely resolved (see Figure S1).

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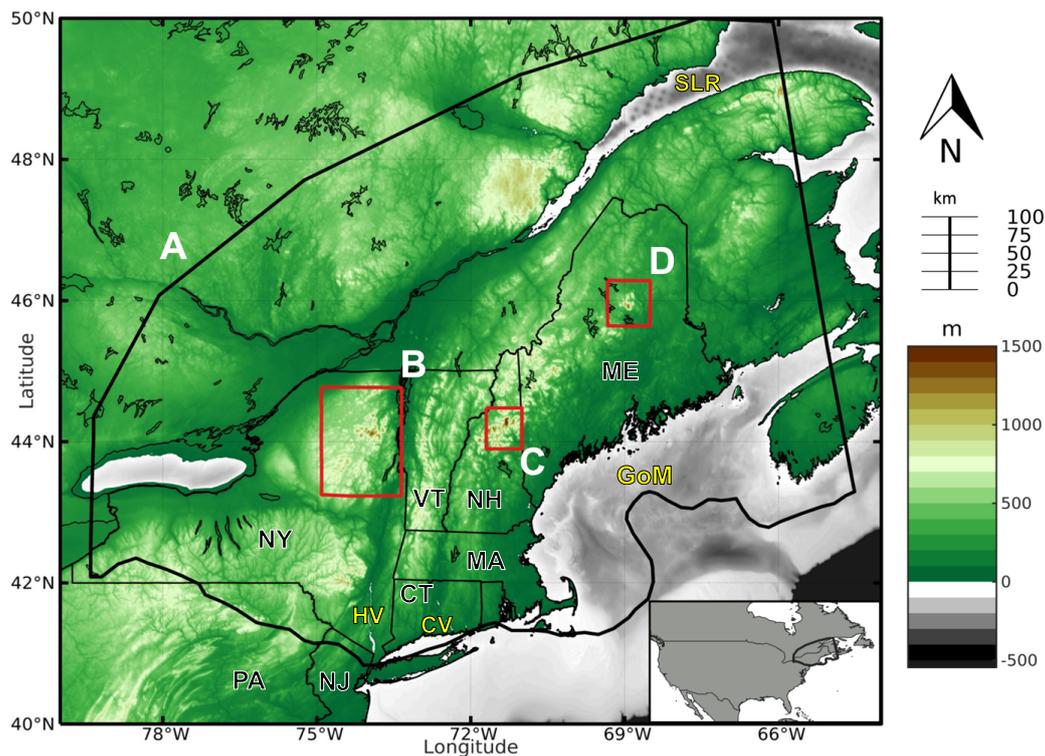


Figure 1. Bedrock topography (meters) and model domains for the regional NE USA ice sheet model (shown as black outline; A) and the local scale models shown in the red outlines across the Adirondack Mountains (B), White Mountains (C) and Mount Katahdin (D). The southern boundary of the NE USA model (A) follows the reconstructed LGM ice extent at the LGM from Dalton et al. (2020). States are abbreviated as: PA: Pennsylvania, NJ: New Jersey, CT: Connecticut, NY: New York, MA: Massachusetts, VT: Vermont, NH: New Hampshire, ME: Maine HV: Other locations described in the text are abbreviated as: Hudson Valley, CV: Connecticut Valley, GoM: Gulf of Maine, SLR: Saint Lawrence River Valley.

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2.1.1 Ice Sheet Model

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We use the Ice-sheet and Sea-level System Model (ISSM; Larour et al., 2012), a higher-order thermomechanical finite-element ice sheet model to simulate the LIS LGM conditions across the NE USA. Because of the high topographic relief across this region and associated impact on ice flow, we use a higher-order approximation to solve the momentum balance equations (Dias dos Santos et al., 2022). This ice flow approximation is a depth-integrated formulation of the higher-order approximation of Blatter (1995) and Pattyn (2003), which allows for an improved representation of ice flow compared with more traditional approaches in paleo-ice flow modelling (e.g., shallow ice approximation or hybrid approaches).

An enthalpy formulation that simulates both temperate and cold-based ice (Aschwanen et al., 2012; Serrousi et al., 2013; Rückamp et al., 2020) is used to capture the thermal state of the ice sheet. Our model contains four vertical layers and uses quadratic finite elements ($P1 \times P2$) along



141 the z axis for the vertical interpolation following Cuzzone et al. (2018) in order to better capture
142 sharp thermal gradients near the base and simulate the vertical distribution of temperature within
143 the ice. This methodology has been successfully applied to simulate the transient behavior of the
144 Greenland Ice Sheet across geologic timescales and the contemporary period (Briner et al., 2020;
145 Smith-Johnsen et al., 2020; Cuzzone et al., 2022). The ice rheology follows Glen’s flow law with
146 the ice viscosity being dependent on the simulated ice temperature following rate factors in Cuffey
147 and Paterson (2010). Surface temperature (see 2.1.3) and geothermal heat flux (Shapiro and
148 Ritzwoller, 2004) are imposed as boundary conditions in the thermal model. We use a linear
149 friction law (Budd et al., 1979), where the basal drag (τ_b) is represented as:

$$150 \tau_b = -\alpha^2 N u_b \quad (1)$$

151 where α represents the spatially varying friction coefficient, N represents the effective pressure,
152 and u_b is the basal velocity. Here $N = g(\rho_i H + \rho_w Z_b)$, where g is gravity, H is ice thickness, ρ_i is
153 the density of ice, ρ_w is the density of water, and z_b is bedrock elevation following Cuffey and
154 Paterson (2010). N evolves as the ice sheet thickness changes. The spatially varying friction
155 coefficient (α) is constructed as a function of bedrock elevation above sea level (Åkesson et al.
156 2018; Cuzzone et al., 2024):

$$157 \alpha = 200 \times \frac{\min[\max(0, z_b + 600), z_b]}{\max(z_b)} \quad (2)$$

158 where z_b is the height of the bedrock with respect to sea level. Using this parameterization, basal
159 friction is larger across high topographic relief and lower across valleys and areas below sea level,
160 which is consistent with what is found today for the Greenland Ice Sheet. The thermomechanical
161 coupling captures the impact of ice deformation and frictional heating on ice temperature, which
162 in turns affect the ice rheology through the temperature dependent rate-factor. In our approach, the
163 friction coefficient is independent of the thermal state but has been explored recently in modeling
164 LIS LGM conditions (Moreno et al., 2023). Although it is found to have a measurable impact the
165 overall simulated LGM ice volume, basal temperatures, and ice stream extent particularly for
166 Hudson Bay, these variations are small when compared with uncoupled simulations particularly
167 across the NE USA.

168 The motion of the ice front is tracked using the level-set method described in Bondzio et
169 al. (2016) and calving is simulated where the LIS interacts with the ocean based on the Von Mises
170 stress criterion (Morlighem et al., 2016). This approach approximates the calving rate as a function
171 of the tensile stresses simulated with the ice, with ice front retreat occurring when the von Mises
172 tensile strength exceeds user defined stress thresholds set for tidewater (1 MPa) and floating ice
173 (200 kPa). These values are consistent with other contemporary and paleo ice sheet modeling
174 studies (Bondzio et al., 2016; Morlighem et al., 2016; Choi et al., 2021; Cuzzone et al., 2024).

175 Glacial isostatic adjustment (GIA) is not simulated transiently, but is instead prescribed by
176 using a reconstruction of relative sea level from a global GIA model of the last glacial cycle (Caron
177 et al., 2018). Bedrock vertical motion, eustatic sea-level, and geoid change at the LGM are applied
178 to our model to account for GIA (Briner et al., 2020; Cuzzone et al., 2024).

183 2.1.2 Model Domains

184 To constrain the model boundary conditions necessary for simulating the LIS conditions
185 across the NE USA at both a regional (Figure 1; Domain A) and local scales across the Adirondack
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187 mountains, White Mountains, and Mount Katahdin (Figure 1; Domain B-D), we first simulate the
188 LGM ice conditions across the LIS (Figure S1). Our model domain follows the LGM ice extent
189 reconstructed from Dalton et al. (2020) but does not include the Cordilleran Ice Sheet and the
190 connection between Ellesmere Island and the Greenland Ice Sheet. The LIS model is built on a
191 structured mesh with a spatial resolution of 20 km. Bedrock topography is initialized from the
192 General Bathymetric Chart of the Oceans (GEBCO; GEBCO Bathymetric Compilation Group,
193 2021), which relies on the 15 arcsecond (450-meter resolution) Surface Radar Topography Mission
194 data (SRTM15_plus; Tozer et al., 2019) for the terrain model. The regional ice sheet model domain
195 (Figure 1, Domain A; Figure S1, Domain 2) covers the NE USA and extends into portions of
196 southern and maritime Quebec. Across this domain anisotropic mesh adaption is used to construct
197 a non-uniform mesh that varies based upon gradients in bedrock topography from GEBCO. The
198 spatial resolution varies from 5 km in areas of low topographic relief to 500 meters in areas of high
199 topographic relief. To gain a more detailed reconstruction of the LGM conditions across the NE
200 USA that are more consistent with the spatial scales associated with the geomorphic data (Halsted
201 et al., 2023), we downscale our results to three local areas within the NE USA (Section 2.2; step
202 3): the Adirondack Mountains, New York; the White Mountains, New Hampshire; and Mount
203 Katahdin, Maine. The local scale models (Figure 1, Domains B-D; Figure S1, Domain 3) also rely
204 on anisotropic mesh adaptation with spatial resolution varying from 400 to 30 meters based upon
205 topography from the 1 arcsecond (~30-meter resolution) SRTM product (Farr et al., 2007). The
206 three locations were chosen based on their geomorphic qualities, as each represents the highest
207 peak within their respective state, and has pre-existing geologic data on glaciation (e.g.,
208 cosmogenic surface exposure ages; Bierman et al., 2015; Barth et al., 2019). At these spatial scales,
209 the model is able to resolve the high topographic relief found between the mountain peaks and
210 neighboring valley floors.

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212 2.1.3 Climate Forcing and Surface Mass Balance scheme

213 Output of monthly mean LGM temperature and precipitation from 5 different climate
214 models is used as surface boundary conditions and to estimate the surface mass balance (SMB).
215 The LGM climate from the TraCE-21ka transient simulation of the last deglaciation is used (Liu
216 et al., 2009; He et al., 2013), as well as output from 4 models participating in the Paleoclimate
217 Modelling Intercomparison Project 4 (PMIP4; Kageyama et al., 2021). The simulated climate from
218 these models differs both with respect to the magnitude and spatial distribution of glacial
219 temperature and precipitation change relative to the preindustrial climate (Figure S2). By using a
220 diversity of surface climate forcings rather than the multi-model mean, we aim to construct a small
221 ensemble of results for the simulated LGM conditions across our model domains as the
222 representative climate model spread can impact the simulated ice sheet geometry, ice flow, and
223 thermal characteristics (Lai and Anders, 2021).

224

<i>Model</i>	<i>Spatial Resolution</i>	<i>Reference</i>
CCSM4 (TraCE-21ka)	3.75° x 3.75°	Liu et al., 2009 He et al., 2013
MPI-ESM1.2 (MPI)	1.8° x 1.8°	Mauritsen et al., 2019
MIROC-ES2L (MIROC)	2.8° x 2.8°	Ohgaito et al., 2021 Hajima et al., 2020
IPSLCM5A2 (IPSL)	3.8° x 1.9°	Sepulchre et al. (2020)
AWIESM2 (AWI)	1.8° x 1.8°	Sidorenko et al. (2019)



225 Table 1. List of climate models used for the LGM surface climate forcing and the corresponding
226 spatial resolution.
227

228 We use a positive degree day model (Tarasov and Peltier, 1999; Le Morzadec et al., 2015;
229 Cuzzone et al., 2019; Briner et al., 2020) to calculate the SMB. Our degree day factor for snow
230 melt is $5 \text{ mm } ^\circ\text{C}^{-1}\text{day}^{-1}$ and $9 \text{ mm } ^\circ\text{C}^{-1}\text{day}^{-1}$ for bare ice melt, and we use a lapse rate of $5^\circ\text{C}/\text{km}$ to
231 adjust the temperature of the climate forcings to surface elevation (Abe-Ouchi et al., 2007). The
232 hourly temperatures are assumed to have a normal distribution, of standard deviation 3.5°C around
233 the monthly mean. An elevation-dependent desertification is included (Budd and Smith, 1981),
234 which reduces precipitation by a factor of 2 for every kilometer change in ice sheet surface
235 elevation and the model accounts for the formation of superimposed ice following Janssens and
236 Huybrechts (2000). The degree day model requires inputs of monthly temperature and
237 precipitation. We apply a commonly used modeling approach to scale a contemporary climatology
238 of temperature and precipitation back to the LGM ('Anomaly Method'; Pollard et al., 2012;
239 Seguinot et al., 2016; Golledge et al., 2017; Tigchelaar et al., 2019; Briner et al., 2020; Cuzzone
240 et al., 2022). The monthly mean climatology of temperature and precipitation for the period 1979-
241 2010 from the European Center for Medium-Range Weather Forecasts ERA5 reanalysis (Hersbach
242 et al., 2020) is bilinearly interpolated onto our model mesh. Next, anomalies of the monthly mean
243 temperature and precipitation fields from the climate models (Table 1), computed as the difference
244 between the LGM and preindustrial control run, are added to the contemporary monthly mean
245 climatology to produce the monthly temperature and precipitation fields at LGM.

246 247 2.2 Experimental Setup

248 Our downscaling approach is conducted over 3 steps:
249 *Step 1 (LIS models; Figure S1 and S3):* First, we construct a model of the LGM LIS using a 2d
250 model with setup described above, and prescribe constant LGM climate (section 2.1.3). Following
251 Moreno et al. (2023), we initialize our model with ice thicknesses of 1000 m north of 50°N , and
252 allow the simulated LIS to reach equilibrium with respect to ice volume ($\sim 50,000$ years). The
253 resulting models (ice geometry and velocity) are used to initialize a 3d model that extrudes the 2d
254 model to 4 vertical layers (P1xP2 vertical finite elements; section 2.1.1). The thermal regime is
255 assumed to be in steady-state with the LGM climate, and we perform a thermomechanical steady
256 state calculation with a fixed ice sheet geometry until the ice sheet velocities are consistent with
257 ice temperature (i.e., convergence) following Seroussi et al. (2013). Lastly, we let the 3d LIS
258 model relax an additional 20,000 years (i.e., until thermal equilibrium is reached) with the LGM
259 climate as the model adjusts to the updated ice temperature and rheology.

260
261 *Step 2 (NE USA models; Figure 1):* We construct our 3d NE USA ice models following the setup
262 discussed in section 2.1 and initialize the model with ice geometry, temperature and rheology, and
263 velocity from the resulting LGM 3d LIS model. Boundary conditions of temperature, ice velocity,
264 and thickness from the LIS results are imposed as Dirichelt boundary conditions at the western,
265 northern, and eastern boundaries following Cuzzone et al. (2022). The initial ice velocity and
266 temperature are downscaled across this domain by performing a thermomechanical steady state
267 calculation (Seroussi et al., 2013). The model is then allowed to relax with constant LGM climate
268 for 20,000 years as the ice geometry, flow, and temperature adjust to the higher resolution grid.

269



270 *Step 3 (local models; Figure 1):* Similar to *Step 2*, the local scale models are initialized with ice
271 geometry, temperature and rheology, and flow with the results from the NE USA model, and
272 boundary conditions (as in *Step 2*) are applied from the NE USA model to the local scale models
273 at the North, South, East, and West boundaries. After running a thermomechanical steady state
274 calculation, the model is allowed to relax for 20,000 years with constant LGM climate as the ice
275 geometry, flow, and temperature adjust to the higher resolution grid.

276
277 Our approach above makes use of the thermomechanical steady state calculation to avoid
278 high computational expense in relaxing the 3d models for a sufficiently long time (e.g. >100,000
279 yr) until thermal equilibrium is reached. For this study, we focus the discussion of our results on
280 the simulated LGM state of the NE USA (step 2) models and local models (step 3), but provide
281 Figure S3 to illustrate the simulated LGM state for the LIS (step 1).

282 283 **3. Results**

284 285 *3.1 Northeast USA*

286 We simulate a range of possible LGM states given different climatologies, and do not
287 specifically tune our models to match reconstructions of ice extent or flowlines. The southern
288 boundary of our model domain is constrained to the maximum reconstructed LGM ice extent from
289 Dalton et al. (2020). In total we have 5 different simulations of the LIS across the NE USA during
290 the LGM (Figure S4), driven by the independent climate forcings (Table 1). The depth integrated
291 ice velocity and ice thickness for the ensemble mean (n=5) of those experiments is shown in Figure
292 2. Individually, most of the experiments simulate an LGM ice margin that reaches the
293 reconstructed terminal LGM ice extent (Figure 2B & S4) from Dalton et al. (2020), although some
294 simulate reduced ice extent particularly along the southern and eastern boundaries (Figure S4).
295 Thinner ice (<500 m) is found along the ice margin and marine terminating portions of the Atlantic
296 Ocean and Gulf of Maine, and thickens to upwards of 3500 m over the Northwest portion of the
297 model domain. Across the Northwest region, ice velocities are slow (<20 m/yr), consistent with a
298 northward trending ice divide simulated through Quebec (see Figure S3). Additionally, a regional
299 ice divide is simulated across the Adirondack mountains (Figure 2A), where ice velocities are
300 <15m/yr and ice flow vectors indicate diverging ice flow to the southwest and southeast. This
301 regional ice divide is simulated for all experiments (Figure S4), with variation in the magnitude of
302 the ice velocities (~1-25 m/yr). Faster ice flow (>50 m/yr) is found along the ice margin and
303 through areas of topographic troughs. Horizontal ice flow is fastest in marine terminating portions
304 such as the Gulf of Maine (GoM; Figure 1), the St. Lawrence River (SLR; Figure 1), and
305 throughout lower terrain such as the Hudson and Connecticut river valley (Figure 2A; HC, CV
306 Figure 1), with speeds approaching and exceeding 300 m/yr.

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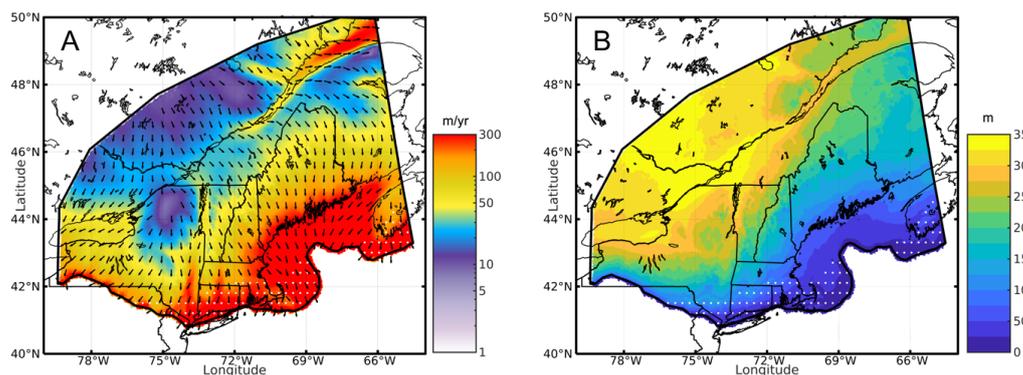


Figure 2. The modeled mean ($n = 5$; Table 1), simulated LGM depth-integrated ice velocity (A; m/yr) and simulated LGM ice thickness (B; m) across the NE USA. Vectors (A) illustrate the simulated direction of ice flow, and colors denote the magnitude of ice velocity. White stippling indicates where areas where some individual models (see Figure S4) simulate no LGM ice cover.

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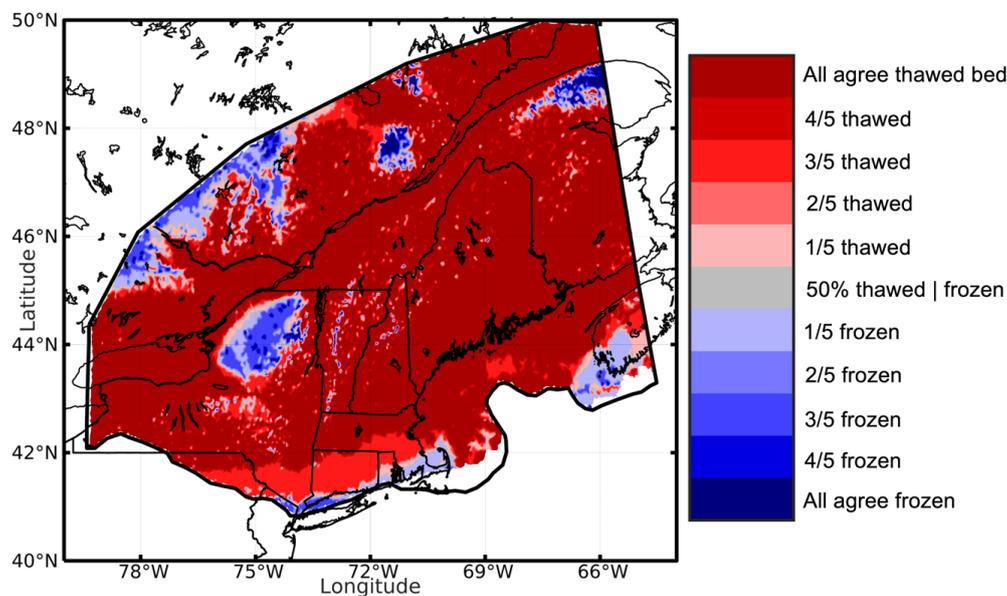


Figure 3. The agreement in the simulated LGM basal thermal state ($n = 5$) for the experiments with varying climate forcing. We assume a thawed bed is present when the pressure corrected $T_{bed} \geq -1^\circ\text{C}$ and frozen bed is present when $T_{bed} < -1^\circ\text{C}$ following MacGregor et al., 2022.

310 For each experiment, we calculate the simulated basal temperature, adjusted for pressure
 311 melting (i.e., pressure corrected temperature). Following MacGregor et al. (2022), we consider a
 312 thawed bed to be simulated where the pressure corrected $T_{bed} \geq -1^\circ\text{C}$ and frozen where $T_{bed} < -1^\circ\text{C}$.



313 The magnitude of the simulated thawed and frozen bed conditions varies across our model domain
314 (Figure S4) but follow a spatial pattern that can be summarized in Figure 3, where we show the
315 inter-model agreement between experiments for thawed and frozen bed conditions. We find that
316 approximately 70% of the model domain is thawed and only 2% of the model domain has frozen
317 bed conditions (i.e., area of domain where all experiments agree), indicating that warm-based ice
318 conditions prevailed across this portion of the LIS. Warm-based areas are coincident with areas of
319 fast ice flow (Figure 2A), where sliding dominates. Frozen bed conditions are generally simulated
320 in areas of reduced ice flow (<15 m/yr; Figure 2A), particularly along the ice divide that exists in
321 the northwest portion of the model domain, south of the St. Lawrence River, and across the
322 Adirondack mountains. Additionally, the regional model simulates the presence of cold-based ice
323 scattered across areas of high bedrock elevation (see Figure 1 for topographic map). In general,
324 each experiment using the different climate surface forcings simulate a similar spatial pattern of
325 warm-based and cold-based ice (Figure S4). Where each experiment disagrees is with respect to
326 the magnitude of basal cooling, with some experiments simulating both colder basal temperatures
327 and a wider swath of frozen bed conditions primarily across the Adirondack Mountains (Figure
328 S4; TraCE-21ka, MPI, and IPSL). This is likely related to the differences in the simulated LGM
329 surface climate between each climate model. The experiments which simulate a larger swath of
330 frozen bed conditions across the Adirondack Mountains (Figure S4; TraCE-21ka, MPI, and IPSL)
331 have a higher magnitude of simulated LGM surface cooling and higher SMB (Figure S5; TraCE-
332 21ka, MPI, and IPSL), versus the experiments which simulate less extensive frozen bed conditions
333 across the Adirondack Mountains (Figure S4 and S5; MIROC and AWI).

334

335 3.2 Local Models

336 3.2.1 Adirondacks

337 The simulated mean (n=5) LGM conditions for the Adirondacks are presented in Figure 4.
338 High topographic relief that exceeds 1000 m is characteristic of the Adirondack Mountains, with
339 43 mountain peaks exceeding 1200 m in elevation, and maximum elevations reaching 1600 m
340 (Figure 4; HPA: High Peaks area). Simulated LGM ice thickness reaches between 1600 m to 2000
341 m across the highest peaks, increases throughout the lower valleys in excess of 2000 m, and reaches
342 upwards of 3000 m across the lower elevations in the Northwest and Northeast portion of the
343 domain (Figure 4C). Consistent with the ice divide simulated in the regional model (Figure 2),
344 slow ice velocities (<10 m/yr to 25 m/yr) dominate across the Adirondacks (Figure 4D). Ice flow
345 generally trends in a south-southeastward direction, with ice velocities increasing to > 60 m/yr
346 across lower elevation valleys. Approximately 58% of the model domain is thawed and 4% of the
347 model domain has frozen bed conditions (i.e., area of domain where all experiments agree),
348 indicating that warm-based ice conditions prevailed across this portion of the LIS (Figure 4E).
349 Generally, frozen bed conditions are confined to high elevation peaks (Figure 4B), where low ice
350 velocities and thinner ice exist, making these high elevation regions more susceptible to vertical
351 advection of cold surface temperature (LGM temperature range from climate models: -24°C to -
352 18°C). However, we also find that across lower elevation regions, particularly in the Northwest
353 portion of the domain, frozen bed conditions are simulated. Here ice is thick (upwards of 3000 m;
354 Figure 4A) but is characterized by very slow ice flow (<10 m/yr). Without limited frictional and
355 deformational heating, this area was likely influenced by the vertical advection of cold surface
356 climate despite thick ice. Across this domain the thermal boundary separating frozen and thawed
357 bed (Figure 4E) is 882 m. However, if we just consider the High Peaks area (HPA; Figure 4A),
358 that thermal boundary resides at 1180 m.

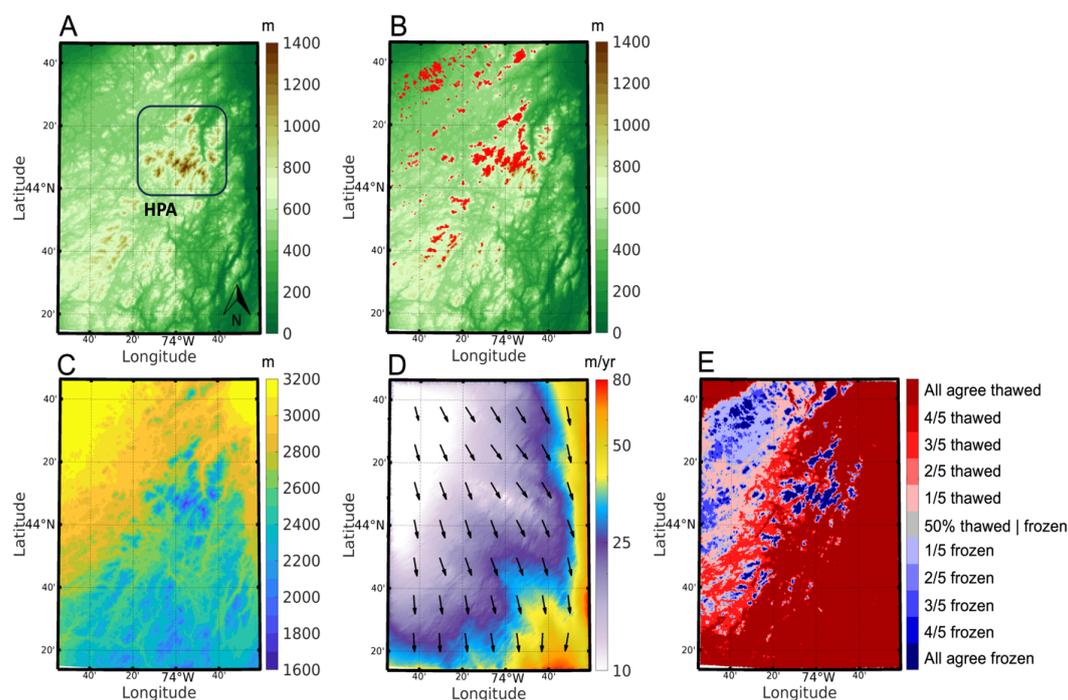


Figure 4. A) Bedrock topography for the Adirondack Mountains (m). Box highlights the High Peaks area. See Figure 1 for geographical context within the NE USA. B) Bedrock topography for the Adirondack Mountains with an overlay of areas where the models agree (n=5) on simulated frozen bed conditions. C) The model mean ice thickness (m). D) The model mean depth average ice velocity (m/yr). Note, vectors denote ice flow direction and not the magnitude of velocity. E) Modeled agreement for simulated frozen and thawed bed conditions.

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3.2.2 White Mountains

361 To the east of the Adirondack Mountains, the White Mountains comprise a series of
362 mountain ranges intersected by deeply incised valleys, with the highest peak Mount Washington
363 reaching 1912 m (Figure 5A). The simulated mean LGM conditions for the White Mountains are
364 presented in Figure 5. Ice thickness ranges between 600 m to 1200 m across high elevation peaks
365 (Figure 5C) and thickens to 2000 m in the deeply incised valleys. Across the lower elevations in
366 the Northwest portion of the model domain (Figure 5A), ice thickness reaches upwards of 2500
367 m. Ice flows southeastward across this region, with simulated ice velocities being higher than is
368 found across the Adirondack Mountains. Ice velocities are lowest in the Northwest portion of the
369 domain and across Mount Washington, where the southeastward ice flow is impeded by the high
370 elevation bedrock (35 – 60 m/yr; Figure 5D), and increases up to 150 m/yr across the lower
371 elevations in the southeast region. When considering the basal thermal regime, we find that 95%
372 of the model domain has thawed bed conditions, with only 0.5% of the domain having frozen bed
373 conditions. Frozen bed conditions are limited to high elevation sites, particularly across the
374 Presidential Range where Mt. Washington is located and Mount Lafayette and Little Haystack in
375 the Franconia Range, with a mean thermal boundary between frozen and thawed bed conditions
376 residing at 1530 m.
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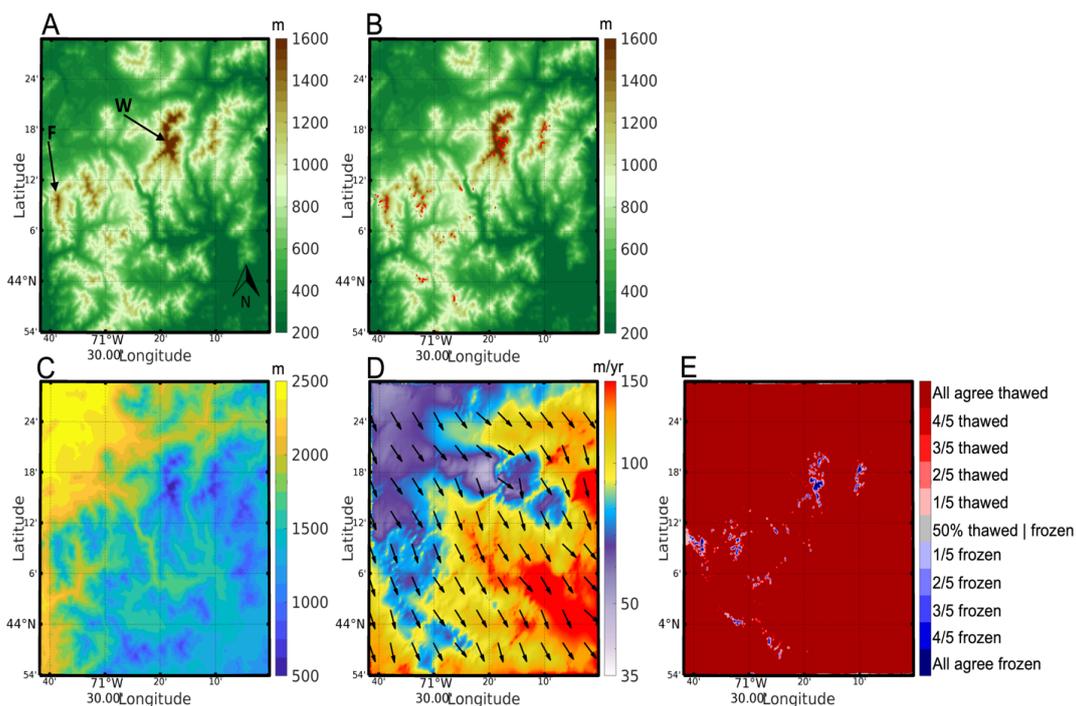


Figure 5. A) Bedrock topography for the White Mountains (m). Highlighted are Mt. Washington (W) and Mount Lafayette and Little Haystack Mountain, labeled (F) for Franconia Range. See Figure 1 for geographical context within the NE USA. B) Bedrock topography for the White Mountains with an overlay of areas where the models agree ($n=5$) on simulated frozen bed conditions. C) The model mean ice thickness (m). D) The model mean depth average ice velocity (m/yr). Note, vectors denote ice flow direction and not the magnitude of velocity. E) Modeled agreement for simulated frozen and thawed bed conditions.

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3.2.3 Mount Katahdin

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Mount Katahdin reaches 1606 m with upwards of 1300 m of relief above the surrounding valleys (Figure 6A). The simulated mean LGM conditions for Mount Katahdin are presented in Figure 6. Across the summit of Mount Katahdin, ice thicknesses reach between 500-700 m, and thickens to 1500 - 2000 m around the flanks of the mountain (Figure 6C) and the surrounding lowlands. The ice generally flows south-southeastward (Figure 6D), with ice flow diverging and slowing down to 25 m/yr upstream of Mountain Katahdin, before reaching a minimum of 10-15 m/yr at the summit. Ice flow converges on the downstream side of Mount Katahdin and reaches 50-100 m/yr across the lower elevations to the south and east. The simulated basal thermal regime indicates that approximately 97% of the domain is thawed (Figure 6E), with only a few locations across the summit of Mount Katahdin having frozen bed conditions, representing ~0.5 % of the model domain. The elevational boundary separating frozen and thawed bed conditions across this domain is simulated to be at 1318 m.

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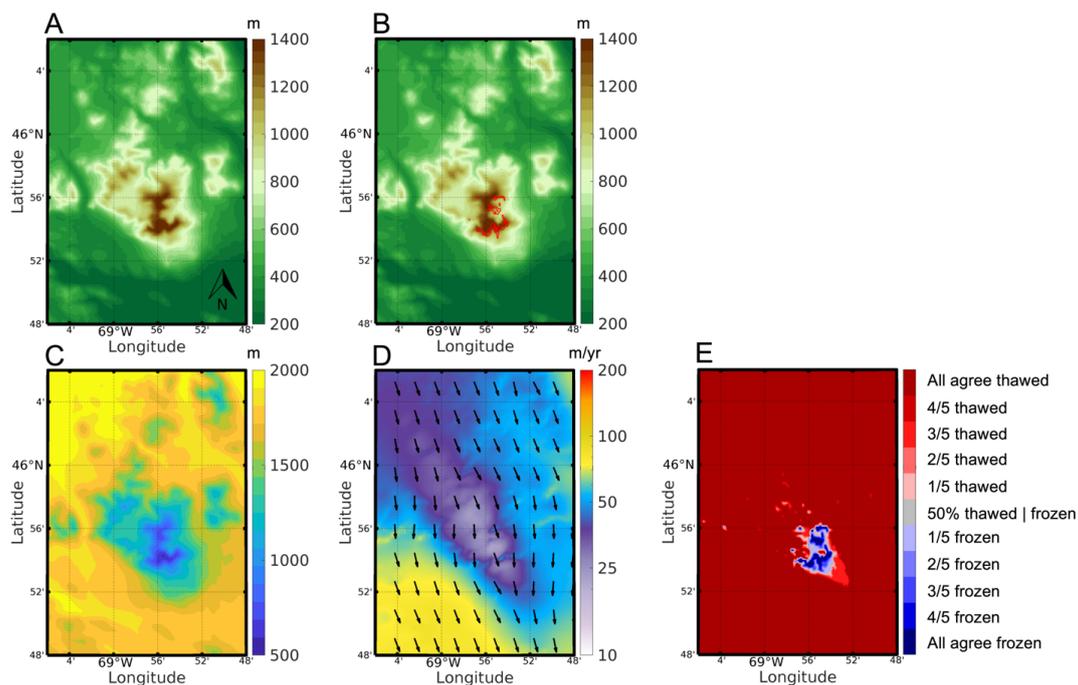


Figure 6. A) Bedrock topography for Mount Katahdin (m). See Figure 1 for geographical context within the NE USA. B) Bedrock topography for Mount Katahdin with an overlay of areas where the models agree ($n=5$) on simulated frozen bed conditions. C) The model mean ice thickness (m). D) The model mean depth average ice velocity (m/yr). Note, vectors denote ice flow direction and not the magnitude of velocity. E) Modeled agreement for simulated frozen and thawed bed conditions.

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4. Discussion

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Across the broader LIS, geomorphic evidence supports the existence of frozen bed conditions interspersed amongst regions of warm-based ice (Klemen and Hattestrand, 1999; Klemen and Glasser, 1997; Marquette et al., 2004) and along southern sectors of the LIS, where ice cover was thin (Colgan et al., 2002). More regional indicators of polythermal conditions with sharp contacts of warm-based ice in fast flowing outlet glaciers and cold-based ice on slow moving ice on uplands have been found across Arctic Canada and Baffin Island (Davis et al., 1999; Davis et al., 2006; Briner et al., 2006; 2014). Prior to direct evidence from surface exposure dating (Bierman et al., 2015), it was unknown if cold-based ice may have existed across the NE USA. Ice sheet models have been used to infer the basal thermal conditions of the LIS, with implications for better understanding large-scale and regional ice flow and controls on ice mass evolution and regional geomorphology (Sugden et al., 1977; Marshall and Clark, 2002; Tarasov and Peltier, 2007; Moreno-Parada et al., 2023). However, while these models provide possible scenarios for LGM and the deglacial evolution of the LIS thermal state, the coarse spatial resolution of existing models limits the ability to capture sharp thermal gradients that may have existed in high relief



412 terrain, with models simulating large scale warm-based conditions during the LGM across the NE
413 USA (Tarasov and Peltier, 2007). Therefore, given that geologic and geomorphic indicators of LIS
414 thermal conditions show that cold-based ice existed across areas of high relief in the NE USA
415 (Goldthwait, 1940; Davis, 1989; Halsted et al., 2023), we relied upon a downscaling procedure to
416 ensure that the underlying bedrock relief and resulting stress balance was well resolved. Our
417 results agree with geologic interpretations (Davis, 1989; Bierman et al., 2015; Halsted et al., 2023)
418 that suggest the existence of polythermal ice across the NE USA during the LGM. While the
419 majority of this region was exposed to warm-based ice conditions (70%; Figure 3), frozen bed
420 conditions are simulated across areas of low ice velocity (i.e. ice divides) and high elevations (2%
421 of area in Figure 3). It is worth noting that while the focus of this study was on the NE USA, areas
422 of frozen bed ice conditions are simulated for portions Maritime Canada (Figure 3), which agree
423 reasonably well with geologic interpretations from this region (Olejczyk and Gray, 2007).

424 Because a majority of the geologic data constraining the thermal regime of the LIS across
425 this region is found in areas of high relief, we focused our modeling on three specific locations.
426 Regionally, geologic interpretations posit that the Adirondack Mountains may have acted as an
427 impediment to the south-southeast flow of the LIS during glacial expansion and the LGM (Franzi
428 et al., 2016). Our simulations suggest that ice flow across this region was slow (<15 m/yr; Figure
429 2A, 4A) in response to a divergence of ice flow around the mountainous terrain. Consequently, a
430 regional ice divide and frozen bed conditions are simulated across portions of this region where
431 ice velocities are <10 m/yr and in areas where the bedrock relief is high. This is further supported
432 when looking at the individual model experiments. Our experiments relied on a small ensemble
433 of simulated surface climate at the LGM from climate model experiments, each simulating a
434 varying magnitude of LGM cooling and precipitation change. Those experiments using colder
435 LGM boundary conditions and which simulated higher SMB (Figure S5; TraCE-21ka, MPI, and
436 IPSL), simulated a stronger magnitude and a wider swath of cold based ice conditions across the
437 Adirondacks (Figure S4; TraCE-21ka, MPI, and IPSL). Such conditions are similar to what we
438 observe across the thick ice divides of modern ice sheets (i.e., Greenland Ice Sheet; MacGregor et
439 al., 2022), where in the absence of heat generation due to frictional heating, vertical advection of
440 cold surface climate dictates the basal thermal state (Lai and Anders, 2021). We also find that
441 frozen bed conditions existed across other areas of high bedrock relief during the LGM (Figure 3)
442 where frozen bed patches are simulated above warm-based ice at lower elevations. Across both
443 the White Mountains and Mount Katahdin (Figure 5&6), upstream ice flow slowed as it
444 encountered resistance from the underlying high bedrock relief. Only where both ice velocity and
445 thickness are relatively low, and bedrock elevation is high, is the presence of frozen bed conditions
446 simulated. Where ice thickness and driving stresses increase downstream of these bedrock
447 features, ice velocities increase and warm-based conditions are simulated.

448 Regional geologic interpretations support that a thermal boundary between cold-based (low
449 erosive) ice and warm-based (erosive ice) existed at ~1200 m across areas of high bedrock relief
450 in the NE USA (Halsted et al., 2022). Terrestrial cosmogenic nuclide (TCN) surface exposure ages
451 from various peaks across the NE USA exhibit signs of nuclide inheritance which is interpreted to
452 reflect inefficient erosion by previous ice coverage. TCN ages from the peaks of Mount Katahdin
453 (ME), Mount Washington (NH), and Little Haystack Mountain (NH), exhibit signs of nuclide
454 inheritance suggestive of a thermal boundary below those locations (Bierman et al., 2015). While
455 warm-based ice indicators (e.g., roche moutonnée and lodgement till) are found on Mt. Washington
456 between 1680 m and 1820 m, a lack of age control on those features means the timing of erosion
457 relative to the LGM is inconclusive. Across Mount Katahdin and the White Mountains, we



458 simulate a thermal boundary of 1318 m and 1530 m respectively, which is in reasonable agreement
459 with the geologic assessment.

460 Across the Adirondack Mountains, frozen bed conditions are simulated both in areas of
461 high bedrock relief and lower elevation sites that reside under a simulated regional ice divide,
462 making the interpretation more complicated. While the region wide thermal boundary is simulated
463 to be at 882 m, if we only consider the High Peaks Area where geologic data of ice thinning exists
464 (Barth et al., 2019), the thermal boundary resides at 1180 m. For high elevation sites, Barth et al.
465 (2019) present an alternative hypothesis that the high-elevation TCN ages suggest early ice-sheet
466 thinning ~20 ka in lieu of reflecting nuclide inheritance and that the regional thermal boundary
467 likely existed above 1560 m. Nevertheless, our simulations confirm that this thermal boundary was
468 not spatially constant, and instead varied geographically (Bierman et al., 2015; Corbett et al., 2018;
469 Koester et al., 2021).

470 While our experiments offer a spatially high-resolution reconstruction of the LGM basal
471 thermal conditions of the LIS across the NE USA that agree well with independent assessments
472 from geologic reconstructions, our methodology assumes that the LIS is in equilibrium with the
473 LGM surface climate, which is likely not a true reflection of LGM conditions. It is noted that
474 while Tarasov and Peltier (2007) simulate the LIS thermal state across the last glacial cycle and
475 find that maximum areal coverage of frozen bed conditions occurred during the LGM, since the
476 LIS experienced a transiently evolving climate prior to the LGM as the surface climate cooled, our
477 simulations may reflect a colder thermal state than may have been experienced. Additionally, the
478 spatially varying basal friction coefficient is constant and not thermodynamically coupled in our
479 simulations. While this coupling has recently been shown to have greatest influence in areas of
480 ice streaming across the LIS with attendant feedbacks on ice surface lowering and ultimately ice
481 thickness, a thermodynamically coupled friction may promote colder basal conditions through
482 feedbacks between ice flow, ice temperature, and basal friction (Moreno et al., 2023). Although
483 we cannot fully address how these limitations and assumptions affect our results, since our
484 simulations agree well with geologic interpretations that polythermal conditions and ultimately a
485 regime of differential erosion existed across the NE USA, we attain a level of confidence that we
486 are indeed simulating thermal conditions that generally reflected LGM conditions. However,
487 future work should evaluate how these thermal conditions may have changed in response to
488 deglacial LIS change as our contemporary assessment of geomorphic and geologic indicators
489 likely integrates the full glacial and deglacial history.

490 Because of the difficulties in conducting dipstick studies aimed at constraining vertical
491 thinning histories, our regional and local scale modeling framework may prove helpful for making
492 more informed choices on sample site selection in places where model simulations suggest warm-
493 based, and ultimately erosive ice conditions, such is currently being done for fieldwork in the
494 Adirondack Mountains (Barker et al., 2024). Additionally, since the results presented here support
495 broader geologic interpretations that polythermal ice conditions likely existed across the NE USA
496 (Halsted et al., 2023), such output may be useful in geomorphological interpretations of differential
497 erosion and relief generation as well as transport processes of glacial erratics from lower elevation,
498 warm-based areas (Bierman et al., 2015). Lastly, such a frame work shows promise in applications
499 to other regions of the LIS where geologic and geomorphic indicators suggest the existence of
500 sharp thermal contacts and erosional history, such as across portions of Arctic Canada and Baffin
501 Island (Briner et al., 2006; 2014).

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504 **5. Conclusions**

505 In this study, we use a numerical ice sheet model to simulate at high spatial resolution,
506 steady-state LGM basal thermal conditions for the LIS across the NE USA and at 3 specific
507 locations characterized by high bedrock relief. LGM climate boundary conditions are used from
508 a small ensemble of climate model simulations, each with a varying degree of LGM cooling and
509 precipitation change relative to preindustrial climate. Our results illustrate that during the LGM,
510 the LIS across the NE USA was mainly warm-based and ultimately erosive, yet exhibited
511 polythermal ice conditions, as simulations reveal that cold-based ice existed across this region in
512 areas of high bedrock elevation and slow ice flow (i.e., ice divides). At local scales, we find that
513 within the Adirondack Mountains, a regional ice divide is simulated during the LGM, characterized
514 by low ice velocities (<15 m/yr) and a wide swath of cold-based ice that spans a large elevational
515 range. Across the White Mountains and Mount Katahdin, ice velocities are generally higher, with
516 cold-based ice conditions being simulated only amongst the highest elevation peaks. Where
517 existing models (Marshall and Clark, 2002; Tarasov and Peltier, 2007; Gregoire et al., 2012) lack
518 sufficient resolution to capture these features, for the first time we simulate a complex thermal
519 regime that may have existed across this region reflective of the highly variable topography of the
520 region.

521 The results presented here support the conclusions from a large dataset of TCN surface
522 exposure ages that relate nuclide inheritance across high relief areas of the NE USA to the presence
523 of cold-based and low erosive conditions sometime during the LGM and last deglaciation (Halsted
524 et al., 2023). These studies largely support that a thermal boundary of ~1200 m in elevation
525 separated cold-based ice at higher elevations and warm-based ice at lower elevations. While our
526 simulations support this conclusion, they also illustrate that this thermal boundary was not spatially
527 consistent and instead varied geographically. Additionally, the results here are supportive of lower
528 latitude polythermal ice conditions existing across the LIS during the LGM (Bierman et al., 2015;
529 Colgan et al., 2002). The existence of polythermal ice conditions across this region has
530 implications with respect to glacial geomorphology, as the erosive character of the ice sheet is
531 closely tied to the basal thermal regime. Since dipstick studies (Bierman et al., 2015; Koester
532 et al., 2017; Barth et al., 2019; Corbett et al., 2019; Koester et al., 2020; Halsted et al., 2023) have
533 the potential to provide critical constraints on paleo ice sheet thinning, yet interpretations can be
534 hindered by the existence of cold-based ice (i.e. low erosion and thus nuclide inheritance), studies
535 like this may aid in future study site selection (e.g. Briner et al., 2022; Barker et al., 2024) as this
536 downscaling procedure can be applied to specific sites of interest. Additionally, this downscaling
537 approach may be useful for geomorphological assessments in other areas of the LIS where
538 polythermal conditions may have existed (Davis et al., 1999; Davis et al., 2006; Briner et al., 2014;
539 Staiger et al., 2005), by providing another metric to evaluate landscape evolution.

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542 **Code and data availability**

543 The simulations performed for this paper made use of the open-source Ice-Sheet and Sea-level
544 System Model (ISSM) and are publicly available at <https://issm.jpl.nasa.gov/> (Larour et al., 2012).
545 Model output described in this study can be found at <https://doi.org/10.5281/zenodo.12665418>
546 (Cuzzone et al., 2024). This includes the simulated output of LGM ice velocity (x and y
547 components as well), ice thickness, and the simulated thermal agreement for cold and warm-based
548 ice across the NE USA, the Adirondack Mountains, the White Mountains, and Mount Katahdin.

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550 **Author contributions.** JC, AB, and KB conceived the study. JC conducted the model setup and
551 conducted the experiments with input from MM. JC analyzed model output with help from AB,
552 KB, and MM. JC and AB wrote the manuscript with input from KB and MM.

553

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555 competing interests.

556

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