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# The Deglaciation of the Americas during the Last Glacial Termination

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91 **Abstract**

92  
93 50 This paper reviews current understanding of deglaciation in North, Central and South  
94 51 America from the Last Glacial Maximum to the beginning of the Holocene. Together  
95 52 with paleoclimatic and paleoceanographic data, we compare and contrast the pace of  
96 53 deglaciation and the response of glaciers to major climate events. During the Global Last  
97 54 Glacial Maximum (GLGM, 26.5-19 ka), average temperatures decreased 4° to 8°C in the  
98 55 Americas, but precipitation varied strongly throughout this large region. Many glaciers  
99 56 in North and Central America achieved their maximum extent during the GLGM, whereas  
100 57 others advanced even farther during the subsequent Heinrich Stadial 1 (HS-1). Glaciers  
101 58 in the Andes also expanded during the GLGM, but that advance was not the largest,  
102 59 except on Tierra del Fuego. HS-1 (17.5-14.6 ka) was a time of general glacier thickening  
103 60 and advance throughout most of North and Central America, and in the tropical Andes;  
104 61 however, glaciers in the temperate and subpolar Andes thinned and retreated during this  
105 62 period. During the Bølling-Allerød interstadial (B-A, 14.6-12.9 ka), glaciers retreated  
106 63 throughout North and Central America and, in some cases, completely disappeared. Many  
107 64 glaciers advanced during the Antarctic Cold Reversal (ACR, 14.6-12.9 ka) in the tropical  
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121 65 Andes and Patagonia. There were small advances of glaciers in North America, Central  
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123 66 America and in northern South America (Venezuela) during the Younger Dryas (12.9-  
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125 67 11.7 ka), but glaciers in central and southern South America retreated during this period,  
126  
127 68 except on the Altiplano where advances were driven by an increase in precipitation.  
128  
129 69 Taken together, we suggest that there was a climate compensation effect, or ‘seesaw’,  
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131 70 between the hemispheres, which affected not only marine currents and atmospheric  
132  
133 71 circulation, but also the behavior of glaciers. This seesaw is consistent with the opposing  
134  
135 72 behavior of many glaciers in the Northern and Southern Hemispheres.

136  
137 73 *Key Words:* Deglaciation, Termination-I, Americas, Late Pleistocene, Glacial  
138  
139 74 Chronology

## 140 76 **1. Introduction**

141  
142 77 This paper focuses on the evolution of glaciation in the Americas during the Last Glacial  
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144 78 Termination. The American continents extend 15,000 km from 70°N to 55°S and are  
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146 79 characterized on their Pacific margins by mountain ranges that are continuous over this  
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148 80 distance and, in most cases, now have glaciers or had them during the last glacial period  
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150 81 of the Pleistocene. Knowledge of the activity of these glaciers has increased enormously  
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152 82 in recent years ([Palacios, 2017](#)). This knowledge provides us an opportunity to study how  
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154 83 American glaciers behaved during the Last Glacial Termination in the context of the  
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156 84 asynchronous climatic setting of the two hemispheres. The largely north-south orientation  
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158 85 and nearly continuous extent of mountain ranges in the Americas provide a unique  
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160 86 opportunity to understand synoptic latitudinal variations in global paleoclimate.

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162 87 The Last Glacial Termination is generally considered to span the time period between the  
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164 88 Global Last Glacial Maximum (GLGM) and the beginning of the current interglacial  
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166 89 period, the Holocene ([Cheng et al., 2009](#); [Denton et al., 2010](#)). It has also been referred  
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168 90 to as Termination I, given that it is the last in a series of similar transitions between  
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170 91 Pleistocene glacials and interglacials ([Emiliani, 1955](#); [Broecker and van Donk, 1970](#);  
171  
172 92 [Cheng et al., 2009](#); [Deaney et al., 2017](#)).

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174 93 The motivation for this review paper is that there have been few attempts to summarize,  
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176 94 synthesize, and compare evidence for late Pleistocene glacier activity across the entire  
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178 95 extent of the Americas. Our objective is to review current understanding of the evolution  
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180 96 of glaciers in both North and South America throughout the Last Glacial Termination and

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180 97 discuss whether the contrasts between the hemispheres implied by paleoclimatic and  
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182 98 paleoceanographic models are reflected in the behavior of the glaciers. Given the  
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184 99 continuous nature of the processes involved in the planet's recent glacial history, parsing  
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186 100 deglaciation into periods requires simplification of the climatic mechanisms. Different  
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188 101 and opposite changes may occur at different latitudes, with variable response times. In  
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190 102 the present article, we have however selected deglaciation phases in accordance with the  
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192 103 current state of knowledge and with scientific tradition.

192 104 Section 2 introduces the study regions and then summarizes how each of the regions has  
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194 105 responded to major climatic changes caused by the different forcings that drove  
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196 106 deglaciation. Section 3 presents the methods that we have used in this work to select study  
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198 107 areas, represent graphically the glacial evolution of each area, and compare glacial  
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200 108 chronologies and paleoclimatic aspects of areas. Section 4 reviews the spatial and  
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202 109 temporal variability of the GLGM in the Americas. The next sections review the  
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204 110 behaviors of these glaciers during deglaciation, notably the Heinrich 1 Stadial (HS-1)  
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206 111 (Section 5), the Bølling-Allerød (B-A) interstadial and the Antarctic Cold Reversal  
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208 112 (Section 6), and the Younger Dryas (YD) (Section 7). These sections are followed by a  
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210 113 discussion (Section 8) in which we: 1) consider uncertainties in numeric ages obtained  
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212 114 on glacial landforms (Section 8.1); 2) summarize knowledge of climate evolution during  
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214 115 the Last Glacial Termination based on research on marine sediments and polar ice cores  
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216 116 (Section 8.2); 3) compare our results with the climatic evolution summarized in Section  
217  
218 117 8.2; and 4) compare our results with published research on glacier activity on other  
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220 118 continents during the Last Glacial Termination (Sections 8.3 to 8.5).

## 216 119

## 218 120 **2. Study Areas**

220 121 Our review proceeds from north to south (Figs. 1 and 2). The study begins with the  
221  
222 122 Laurentide Ice Sheet (LIS) (Fig. 1), which contributed most to sea-level rise during the  
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224 123 Last Glacial Termination ([Lambeck et al., 2014](#)) and was capable of greatly disrupting  
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226 124 the coupled ocean-atmosphere system during deglaciation ([Broccoli and Manabe, 1987a](#);  
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228 125 [Heinrich, 1988](#); [Clark, 1994](#); [Barber et al., 1999](#); [Hemming, 2004](#)). We summarize the  
229  
230 126 most recent syntheses about LIS deglaciation ([Dyke, 2004](#); [Stokes 2017](#)), enabling  
231  
232 127 comparisons with deglaciation in the American mountains. Alaska is traversed by high  
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234 128 mountains and was only partially glaciated during the GLGM. We examine this region as  
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236 129 an unusual example of mountain glaciation at northern high latitudes ([Briner et al., 2017](#)).

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239 130 We next describe the Cordilleran Ice Sheet (CIS) in southwestern Canada and adjacent  
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241 131 United States, from roughly 48°N to 52°N, which removed or buried much of the  
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243 132 preceding alpine glacial record (Clague, 2017), and the North Cascades in Washington  
244  
245 133 State from 47°N to 49°N, which provide an excellent record of the early part of the Last  
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247 134 Glacial Termination (Porter, 1976; Porter et al., 1983; Riedel et al., 2010; Riedel, 2017).  
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249 135 The climate of these areas is strongly influenced by the location of the northern westerlies.  
250  
251 136 There was widespread alpine glaciation in the central sector of western North America:  
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253 137 to the west in the Sierra Nevada Mountains in California; to the east in the Rocky  
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255 138 Mountain/Yellowstone region, and in between the numerous mountain ranges of the  
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257 139 Basin and Range Province (Fig. 1). Glaciation in the western U.S. has been the subject of  
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259 140 numerous recent studies (Licciardi et al., 2001, 2004; Munroe et al., 2006; Licciardi and  
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261 141 Pierce, 2008; Refsnider et al., 2008; Thackray, 2008; Laabs et al., 2009; Young et al.,  
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263 142 2011; Shakun et al., 2015b; Leonard et al., 2017a, 2017b; Licciardi and Pierce, 2018;  
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265 143 Dahms et al., 2018, 2019). In the interior, we mainly focus on the greater Yellowstone  
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267 144 glacial system and adjacent mountain ranges around 44-45°N where new glacial  
268  
269 145 syntheses are available (Larsen et al., 2016; Licciardi and Pierce, 2018; Pierce et al., 2018;  
270  
271 146 Dahms et al., 2018, 2019); and the Rocky Mountains of Colorado at 37-41°N, for which  
272  
273 147 there are also some recent contributions (Ward et al., 2009; Young et al., 2011; Leonard  
274  
275 148 et al., 2017a, 2017b; Brugger et al., 2019). The Sierra Nevada, from 36° to 38°N, is one  
276  
277 149 of the most-studied mountain ranges in North America, and numerous syntheses have  
278  
279 150 been written on its glacial history (Gillespie and Zehfuss, 2004; Gillespie and Clark,  
280  
281 151 2011; Phillips, 2016, 2017).  
282  
283 152 Southward, the combined effects of lower elevation and higher ELA result in the limited  
284  
285 153 presence of glacial landforms in the southern United States and northern Mexico.  
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287 154 However, in central Mexico, at about 19°N, the high volcanoes (>5000 m above sea level,  
288  
289 155 asl) of the Trans-Mexican Volcanic Belt were glaciated (Fig. 1). Elevations decrease  
290  
291 156 again in southern Mexico, and there are two mountain ranges in Central America (>3800  
292  
293 157 m asl) that hosted glaciers during the Late Pleistocene: Sierra Altos Cuchumatanes in  
294  
295 158 Guatemala and the Cordillera de Talamanca in Costa Rica. There are some recent  
159 syntheses of the glacial history of central Mexico (Vázquez-Selem and Heine, 2011;  
160 Vázquez-Selem and Lachniet, 2017) and the Central American glaciated ranges (Lachniet  
161 and Selzer, 2002; Roy and Lachniet, 2010; Cunningham et al., 2019; Potter et al., 2019).

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298 162 [Lachniet and Vázquez-Selem \(2005\)](#) and [Vázquez-Selem and Lachniet \(2017\)](#) recently  
299 163 summarized the history of Quaternary glaciation for this entire region.

301  
302 164 In South America (Fig. 2), the crest and high valleys of the Andes from the north at 11°N  
303 165 to the south at 55°S (a distance of over 7200 km) were glaciated during the last glacial  
304 166 cycle. The northern Andes are located between latitudes 11°N and 4°S, and include ranges  
305 167 in Venezuela, Colombia, and Ecuador. The Venezuelan Andes consist of two main ranges  
306 168 oriented northeast to southwest between 7°N and 10° N, named Sierra de Perijá and the  
307 169 Mérida Andes; the latter contains abundant glacial landforms and extant glaciers.  
308 170 Numerous dating studies have been performed on glacial landforms in that region (e.g.  
309 171 [Schubert, 1974](#); [Bezada, 1989](#); [Mahaney et al., 2000](#); [Dirszowsky et al., 2005](#);  
310 172 [Wesnousky et al., 2012](#); [Angel et al., 2013, 2016, 2017](#); [Carcaillet et al., 2013](#); [Guzmán,](#)  
311 173 [2013](#); [Angel, 2016](#);). The Colombian Andes consist of three parallel ranges extending  
312 174 from 1°N to 11°N: the Cordillera Occidental (western), Cordillera Central and Cordillera  
313 175 Oriental (eastern). Studies have been carried out in the Cordillera Central and Cordillera  
314 176 Oriental involving radiocarbon dating of paleosols and glaciofluvial and glacial  
315 177 sediments, and more recently surface exposure <sup>10</sup>Be dating ([Thouret et al., 1996](#);  
316 178 [Clapperton, 2000](#); [Helmens, 2004, 2011](#); [Jomelli et al., 2014](#)). The Ecuadorian Andes  
317 179 extend from 1°N to 4°S and include the Eastern and Western Cordilleras. Glaciation  
318 180 studies in these ranges have relied mainly on radiocarbon dating of glaciolacustrine and  
319 181 till sediments ([Clapperton et al., 1997a](#); [Rodbell et al., 2002](#); [La Frenierre et al., 2011](#)).

320  
321 182 The central Andes extend the length of Peru, western Bolivia and northern Chile, and  
322 183 comprise two parallel ranges in which the highest areas have glacial landforms and extant  
323 184 glaciers (Fig. 2). Databases have been compiled to inform paleoclimate modeling and to  
324 185 compare glacier activity in Peru and Bolivia (e.g. [Mark et al., 2005](#)). Cosmogenic nuclide  
325 186 exposure dating methods have improved knowledge of Late Pleistocene glacial evolution,  
326 187 but there are significant challenges in interpreting the data ([Smith et al., 2005, 2008](#); [Zech](#)  
327 188 [et al., 2008](#); [Glasser et al., 2009](#); [Licciardi et al., 2009](#); [Rodbell et al., 2009](#); [Smith and](#)  
328 189 [Rodbell, 2010](#); [Blard et al., 2013](#); [Jomelli et al., 2011, 2014](#); [Bromley et al., 2016](#); [Martin](#)  
329 190 [et al., 2018](#)). Several other studies in this region focus on the time of deglaciation ([He et](#)  
330 191 [al., 2013](#); [Shakun et al., 2015b](#); [Stansell et al., 2015, 2017](#)). A recent synthesis of Late  
331 192 Pleistocene glacial evolution has been published for the entire region ([Mark et al., 2017](#)),  
332 193 and this review has since been complemented by additional paleoglacier chronologies  
333 194 ([Ward et al., 2017](#); [Martin et al., 2018](#))

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356  
357 195 In southern Peru (Fig. 2), western Bolivia and northern Chile, the western Andean range  
358 196 marks the west edge of the Altiplano and Puna Plateau, a closed basin that contains the  
360 197 great lakes of Titicaca (3806 m asl), Poopó (3685 m asl), and Salar de Uyuni (3653 m  
362 198 asl). Here, typical elevations are 4000-5000 m asl; the basin is surrounded by the Western  
363 199 and the Eastern Andes, where large volcanoes reach elevations greater than 6000 m asl.  
365 200 In the north of this area, precipitation is delivered mainly by the South American monsoon  
366 201 in the summer months, but to the south, this gives way to extratropical systems related to  
368 202 the southern westerly winds in austral winter. This transition region, between 18°S and  
370 203 30°S, is an area of persistent aridity known as the Arid Diagonal (De Martonne, 1934).

372 204 The western cordillera of the Andes in northern Chile crosses the Arid Diagonal between  
373 205 18°S and 27°S (Fig. 2). Between Nevado Sajama (18.1°S) and Cerro Tapado (30.2°S),  
374 206 there are few modern glaciers because of limited precipitation (Casassa et al., 2007), but  
375 207 glacial deposits can be mapped as far south as ~24°S on the north side of the Arid  
376 208 Diagonal and as far north as 27°S on the south side (Jenny et al., 1996). A few small rock  
377 209 glaciers and permanent snowfields exist on very high peaks throughout the Arid  
378 210 Diagonal, where ELAs reach >6000 m asl (Ward et al., 2017).

383 211 Moving farther south to the northern part of the Argentine Andes between 22°S and 36°S  
384 212 (Fig. 2), there are two different atmospheric circulation patterns, which again are  
385 213 separated by the Arid Diagonal. The Arid Diagonal crosses this section of the Andes  
386 214 between 25°S and 27°S. Most of the precipitation north of the Arid Diagonal falls during  
387 215 the South American summer monsoon season. South of the Arid Diagonal precipitation  
388 216 falls mainly during the austral winter months and is related to southerly sourced westerly  
389 217 winds. The locations where most precipitation is related to the South American summer  
390 218 monsoon are Tres Lagunas (Zech et al., 2009), Nevado de Chañi (Martini et al., 2017a),  
391 219 Sierra de Quilmes (Zech et al., 2017), and Sierra de Aconquija (D’Arcy et al., 2019). The  
392 220 locations where most precipitation is related to the southern westerlies are: the Ansilta  
393 221 range (Terrizzano et al., 2017), Cordon del Plata (Moreiras et al., 2017) and Las Leñas  
394 222 valley (Zech et al., 2017). Reviews of the glacial chronology of the entire region were  
395 223 carried out by Zech et al. (2017) and more recently by D’Arcy et al. (2019).

404 224 From 36°S to the southernmost tip of South America, the Patagonian Andes are a complex  
405 225 mountainous region with numerous present-day glaciers and two large ice fields (Campo  
406 226 de Hielo Patagónico Norte and Sur) (Fig. 2). Mapping of major moraine systems  
407 227 throughout Patagonia and early geochronological work have provided a broad framework



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228 that underpins our knowledge of the glacial history of this region (e.g., [Caldenius, 1932](#);  
229 [Feruglio, 1950](#); [Flint and Fidalgo, 1964, 1969](#); [Mercer, 1969, 1976](#)). During the GLGM,  
230 there was a large ice sheet, the Patagonian Ice Sheet (PIS), that extended 2000 km along  
231 the crest of the range, from 38°S to 55°S. With the exception of northern Patagonia, the  
232 western outlet glaciers of the Patagonian Ice Sheet terminated in the Pacific Ocean,  
233 whereas eastern outlets terminated on land. The deglaciation chronology and pattern of  
234 land-terminating outlets of the PIS have been the subject of much research ([Denton et al.,](#)  
235 [1999a](#); [Glasser et al., 2004, 2008](#); [Kaplan et al., 2008](#); [Moreno et al., 2009](#); [Rabassa and](#)  
236 [Coronato, 2009](#); [Rodbell et al., 2009](#); [Hein et al., 2010](#); [Harrison and Glasser, 2011](#); [Boex](#)  
237 [et al., 2013](#); [Mendelova et al., 2017](#)).

238 The southern tip of South America, from the Strait of Magellan to Cape Horn, comprises  
239 hundreds of islands (Fig. 2). The largest island, Isla Grande de Tierra del Fuego, is  
240 dominated on its west side by the Cordillera Darwin, a mountain range with peaks over  
241 2000 m asl. This range is currently covered by large glaciers, some of which reach the  
242 sea. The climate of Tierra del Fuego is strongly affected by the southern westerlies, and  
243 precipitation declines rapidly from the Pacific to the Atlantic coast. [Hall et al. \(2017a\)](#)  
244 and [Hall et al. \(2019\)](#) published recent syntheses of the glacial history of Tierra del Fuego  
245 during Last Glacial Termination and the Holocene, respectively.

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### 247 **3. Methods**

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248 *3.1 Selection of studies in each area*

249 It is impossible to include all available information on the deglaciation of the Americas  
250 in detail in a single review paper. However, this does not preclude us from carrying out a  
251 comparative analysis of the Late Glacial history of the two continents based on recent  
252 advances in knowledge that we seek to provide here. With this objective in mind, we  
253 selected regions where studies of Late Glacial history are most advanced and  
254 geographically representative. For each selected region, we review recent publications  
255 that are key to understanding the glacial history from the Last Glacial Maximum to the  
256 beginning of the Holocene, including the most up-to-date review papers or syntheses from  
257 specific regions.

258 *3.2 Graphical expression of glacier extent for each interval*

259 The figures in this paper illustrate generalised glacier extent for each of the intervals  
260 discussed below (Figs. 3, 4, 5, 6, and 7). A common metric is required to compare glacier  
261 advances and extent across the vast area of the Americas. Many researchers consider the  
262 Equilibrium Line Altitude (ELA) to be the best measure of climate-driven changes in  
263 glacier extent (Rea, 2009), although it may introduce errors in paleoclimatic  
264 reconstructions in active tectonic mountain ranges (Mitchell and Humphries, 2015) and  
265 is perhaps less helpful for large continental ice sheets. We are unable to use ELA in our  
266 review for three reasons. First, many studies do not report ELAs for glacial events.  
267 Second, the ELAs reported in the papers we surveyed were calculated using different  
268 methods and thus may not be comparable between regions. Third, reported ELAs are  
269 generally local values and may not be representative of regional climate. For example, an  
270 ELA reconstructed for a heavily shaded glacier in a north-facing cirque in the Northern  
271 Hemisphere will yield a much lower value than one reconstructed for an exposed glacier  
272 on the south side of the same mountain. To be quantitatively useful, both of these must  
273 be normalized to the climatic ELA – the zero-mass-balance elevation of a horizontal  
274 unshaded surface. To evaluate the climatic ELA, one must model the mass balance of the  
275 glacier using a digital elevation representation of the basin. Modeling the physical mass  
276 balance of a large number of glaciers spanning the entire Americas is beyond the scope  
277 of this paper.

278 We therefore use a simple, easy-to-compute metric that is based on observational data –  
279 the relative extent of a glacier (Et), expressed as a percentage and quantified as  
280 follows:

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534  
535 281 
$$E_t(\%) = \frac{Z_t - Z_P}{Z_{LLGM} - Z_P}$$
  
536  
537

538 282 where  $Z_t$  is the elevation of the glacier terminus during the period in question,  $Z_P$  is the  
539 283 elevation at the end of the Little Ice Age, prior to the anthropogenic period, and  $Z_{LLGM}$  is  
540 284 the terminal elevation at the Local Last Glacial Maximum (LLGM). For areas that have  
541 285 no historic glaciers, we use the highest elevation in the catchment as a default value for  
542 286  $Z_P$ . In the figures, we have grouped  $E_t$  values for each climate region in 20% intervals.  
543 287 Some ice masses, notably the Laurentide and Patagonian ice sheets, did not uniformly  
544 288 descend downslope from high-elevation accumulation areas, but rather expanded from  
545 289 higher elevation accumulation across vast expanses of relatively flat terrain. For these  
550 290 areas, we used the terminal position (in kilometers) relative to the late Holocene, or final,  
551 291 position as a metric of relative extent.

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554 292 *3.3 ELA depression in the Americas*  
555 293

556 293 In the text and tables, we refer to the approximate decrease in ELA for each period with  
557 294 respect to the current ELA. We acknowledge that glacier extents can be affected by  
558 294 hypsometry, but in this broad review paper and, as noted above, we are not in a position  
559 295 to perform original mass-balance modeling of a large number of glaciers spanning the  
560 296 entire Americas, which itself could be the subject of a large research project. ELA  
561 297 depression data included in our tables are based on cited peer-reviewed papers. The values  
562 298 should be considered approximations, but are useful for comparing how glaciers in each  
563 299 region responded to climate during each of the periods we discuss and for testing  
564 300 hypotheses of the large-scale driving mechanisms.  
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569 302 *3.4 Dating glacial landforms*  
570 303

571 303 Our work compares chronological data obtained over recent decades through cosmogenic  
572 304 nuclide, surface-exposure dating methods. New scaling models and reference production  
573 305 rates have considerably changed the interpretation and chronological framing of many  
574 306 glacial landforms in recent years (e.g. [Kaplan et al., 2011](#); [Blard et al., 2013 a,b](#); [Kelly et al., 2015](#);  
575 307 [Martin et al., 2015, 2017](#); [Borchers et al., 2016](#); [Marrero et al., 2016](#), [Philips et al., 2016](#)).  
576 308 However, the degree of uncertainty in the production rates of most terrestrial  
577 309 cosmogenic isotopes, especially those that do not derive from quartz, can be greater than  
578 310 the amount of time separating many of the phases of deglaciation ([Marrero et al., 2016](#)),  
579 311 making it difficult to relate a glacier landform to a particular short period in the past.  
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592  
593 312 Nevertheless, we account for these differences by identifying scaling factors explicitly or  
594 313 providing citations to relevant publications that allow the reader to be informed.

596  
597 314 Other possible problems may compromise the validity of cosmogenic nuclide exposure  
598 315 ages. Exposure ages can be misleading if the dated glacial landforms are found to have  
600 316 had previous exposure to radiation or have been eroded out of till subsequent to glacier  
602 317 retreat (Blard et al., 2014; Briner et al., 2016; Çiner et al., 2017). Many of the dated glacial  
604 318 landforms discussed in this review are boulders on the crests of moraines, and their  
605 319 apparent ages must be interpreted in the context of advances or stillstands of glacier  
606 320 fronts. Care must be taken when interpreting these ages (Kirkbride and Winkler, 2012)  
608 321 because, once constructed, a moraine may not stabilize for a long time (Putkonen et al.,  
609 322 2008; Heyman et al., 2011). Moreover, weathering and exhumation since stabilization  
611 323 commonly remove grains from the surfaces of boulders, leading to ages that are younger  
612 324 than those of the moraines on which they lie (Briner et al., 2005; Hein, 2009; Heyman et  
614 325 al., 2011; Oliva and Ruiz-Fernández, 2015). Frequently, glacier fronts are limited in their  
616 326 advance by previously formed moraines and, in such cases, the glacier may deposit new  
617 327 boulders on old moraines. This process can be repeated several times and form a single  
619 328 moraine ridge that is the product of multiple advances (Osborn, 1986; Winkler and  
621 329 Matthews, 2010; Schimmelpfennig et al., 2014). The elevation of sample sites, which  
622 330 have changed frequently through the time on account of glacio-isostatic adjustments, is  
624 331 essential in calculating cosmogenic ages. Moreover, the pattern of these changes is very  
625 332 difficult to know, which also introduces uncertainty in cosmogenic ages (Jones et al.,  
627 333 2019). Snow can reduce the exposure of surfaces to cosmogenic radiation and, in most  
629 334 cases, it is difficult to judge the impact of variations in snow cover over the thousands of  
630 335 years that a surface was exposed (Schildgen et al., 2005). Finally, in some cases, glaciers  
632 336 can advance or retreat independently of climatic forcing (Quincey et al., 2011; Ó Cofaigh  
633 337 et al., 2019).

635 338 These potential problems may not necessarily be solved by collecting and analyzing a  
636 339 larger number of samples from the same glacial landform. If altered boulders or boulders  
637 340 with prior radiation exposure are sampled, the statistic only increases the error (Palacios,  
640 341 2017). Placing the results of cosmogenic nuclide exposure dating within a suitable  
642 342 geomorphological context is far more important than the statistics themselves. This  
643 343 context provides grounds for discarding impossible results and preferentially weighting  
645 344 others. For this reason, this review has relied not only on surface exposure ages and the

650  
651  
652 345 most recent production rates, but also on radiocarbon ages and regional geomorphological  
653  
654 346 contexts that strengthen age interpretations and indicate the degree to which they might  
655  
656 347 be in error.

657  
658 348 *3.5 Possible errors and uncertainties in ages reported in this review*

659  
660 349 The studies on which this review is based have been conducted over many decades,  
661  
662 350 during a period when dating methods and standards have markedly changed. Although  
663  
664 351 our focus is on recent literature, which is based on current knowledge, we include  
665  
666 352 pertinent older studies that report ages calculated using earlier protocols. Given the many  
667  
668 353 hundred studies and tens of thousand ages involved, reconciliation of chronological  
669  
670 354 differences resulting from different methods would constitute a major research project  
671  
672 355 and one that is far beyond the scope of this review. To that end, we therefore caution  
673  
674 356 readers that the patterns we draw from the literature are a starting point for more detailed  
675  
676 357 comparisons between specific study areas; we encourage readers to thoroughly evaluate  
677  
678 358 and, if necessary, recompute ages reported in the literature for such purposes.

679  
680 359 However, as shown below, ages cited in this paper are still comparable, because  
681  
682 360 systematic uncertainties resulting from different methods and production rates are lower  
683  
684 361 than 5% for most of the ages that we discuss here. Most of the ages cited in the text have  
685  
686 362 been calculated using the  $^{10}\text{Be}$  isotope and are derived from rocks containing quartz.  
687  
688 363 Some ages cited in the text have been calculated using  $^{36}\text{Cl}$  and  $^3\text{He}$  in rocks without  
689  
690 364 quartz, commonly volcanic rocks. Recent literature has shown that the ages derived from  
691  
692 365 these three isotopes are comparable, albeit with different uncertainties (Phillips, 2016,  
693  
694 366 2017; Barth et al., 2019).

695  
696 367 Balco and Schaefer (2006), Thompson et al. (2017), Corbett et al., (2019), and Barth et  
697  
698 368 al. (2019) have recently recalculated  $^{36}\text{Cl}$  ages cited in the Laurentide sections of this  
699  
700 369 paper and have concluded that they differ little from previously published ages. All  $^{10}\text{Be}$   
701  
702 370 ages from Alaska have been calculated using similar production rates: the Arctic value of  
703  
704 371 Young et al. (2013) or the NENA value of Balco et al. (2009). Menounos et al. (2017)  
705  
706 372 report  $^{10}\text{Be}$  ages for the area of the Cordilleran Ice Sheet that are consistent with both  
707  
708 373 previously and subsequently published ages from this region. Recently,  $^{10}\text{Be}$  ages for the  
374  
375 Rocky Mountains/Yellowstone region have been calculated or recalculated by Shakun et  
376  
377 al. (2015a), Dahms et al. (2018), Licciardi and Pierce (2018), and Pierce et al. (2018), and  
shown to be internally consistent and consistent with the other ages in North America.  
Sierra Nevada  $^{36}\text{Cl}$  and  $^{10}\text{Be}$  ages are taken from Phillips (2016) and Phillips (2017) and

709  
710  
711 378 are based on CRONUS-Earth production rates (Borchers et al., 2016, Marrero et al.,  
712 379 2016a, Phillips et al., 2016). Whole-rock cosmogenic  $^{36}\text{Cl}$  ages on moraine boulders and  
714 380 glacially polished rock surfaces in Mexico and Central America are based on calculations  
716 381 and recalculations using CRONUScalc (Marrero et al., 2016a, 2016b). Ages from Mexico  
717 382 reported by Vázquez-Selem and Lachniet (2017) and Central America reported by Potter  
719 383 et al. (2019) and Cunningham et al. (2019) are based on different scaling models, but the  
720 384 age differences are less than 2.5%. In conclusion, all cosmogenic ages from North and  
722 385 Central America cited in the text have been calculated or recalculated in the past three  
724 386 years, and possible differences are likely less than 5%.

726 387 Turning to South America, the Late Glacial chronology in the northern Andes is mainly  
727 388 based on radiocarbon and  $^{10}\text{Be}$  cosmogenic ages. Many  $^{10}\text{Be}$  ages were recomputed using  
729 389 the Cosmic Ray Exposure Program (CREP, <http://crep.cirpa.cnrs-nancy.fr/#/>) (Martin et  
730 390 al., 2017) and the synthetic High Andes  $^{10}\text{Be}$  production rate reported by Martin et al.  
732 391 (2015). Jomelli et al. (2014) notably homogenized and recalculated 477 published  $^{10}\text{Be}$   
734 392 and  $^3\text{He}$  surface exposure ages from the Peruvian and Bolivian Andes, spanning the past  
735 393 15,000 years. After the publication of the paper by Jomelli et al. (2014), Martin et al.  
737 394 (2015) proposed a new empirical  $^{10}\text{Be}$  production rate for the Tropical Andes that is  
739 395 similar, within uncertainties, to those proposed by Blard et al. (2013a) and Kelly et al.  
740 396 (2015). Jomelli et al (2017) and Martin et al (2018) adopted this new production rate and  
742 397 reported recalculated new ages, which we follow in this paper. A recent review by Mark  
743 398 et al. (2017) provides additional information on the Late Glacial chronology of the  
745 399 Peruvian and Bolivian Andes. Alcalá-Reygosa et al. (2017) report  $^{36}\text{Cl}$  ages from volcanic  
746 400 areas in the Peruvian Andes, which they calculated using the spreadsheet developed by  
748 401 Schimmelpfennig (2009) and Schimmelpfennig et al. (2009). Bromley et al. (2011)  
749 402 provide  $^3\text{He}$  ages for the same area. The  $^{36}\text{Cl}$ ,  $^3\text{He}$ , and radiocarbon ages from the Peruvian  
751 403 Andes are consistent with one another (Blard et al. 2013a, 2013b; Bromley et al., 2019).  
753 404  $^{10}\text{Be}$  ages from northern Chile were calculated or recalculated by Ward et al. (2015, 2017)  
754 405 based on a protocol similar to that used in the Peruvian and Bolivian Andes.  $^{36}\text{Cl}$  ages  
756 406 from northern Chile (Ward et al., 2017) were calculated using CRONUS-Earth  
757 407 production rates and the LSDn routine in CRONUScalc (Marrero et al., 2016b). Recently,  
759 408 D'arcy et al. (2019) recalculated all the  $^{10}\text{Be}$  ages from the Central Andes of Argentina  
761 409 using local High Andes production rates (Kelly et al., 2015; Martin et al., 2015, 2017).  
762 410 Ages from Patagonia and Tierra del Fuego were recalculated by the CRONUS-Earth

768  
769  
770 411 online exposure age calculators (v. 2.2) (Balco et al., 2008), using the time-dependent  
771  
772 412 Lal/Stone scaling model and the "Patagonian" production rate of Kaplan et al. (2011). As  
773  
774 413 in North and Central America, the South American cosmogenic ages differ slightly  
775  
776 414 between some regions, but the errors do not exceed 5% and thus do not change the overall  
777  
778 415 conclusions of the paper.

779 416 We have converted <sup>14</sup>C ages from all the regions to calendar year ages using CALIB 7.1.

## 780 781 417 **4. The Manifestation of the Global Last Glacial Maximum (26.5-19 ka) in the** 782 783 418 **Americas and the Start of the Last Glacial Termination**

### 784 785 419 *4.1 The Global Last Glacial Maximum*

786  
787 420 The first period analyzed covers the time between 26.5 ka and 19 ka, when most of the  
788  
789 421 northern ice sheets and many mountain glaciers reached their maximum extent in the last  
790  
791 422 glacial cycle (Clark et al., 2009). This period coincides with the time of minimum sea  
792  
793 423 level and is characterized by a quasi-equilibrium between the cryosphere and climate  
794  
795 424 (Clark et al., 2009). Following standard usage (Clark et al., 2009; Hughes et al., 2013),  
796  
797 425 we have called this period the 'Global Last Glacial Maximum' (GLGM). Clark et al.  
798  
799 426 (2009) note that many ice masses, especially mountain glaciers, achieved their maximum  
800  
801 427 extents prior to or after this period, and that the term 'Local Last Glacial Maximum'  
802  
803 428 (LLGM) should be used to describe local maxima in particular regions. They further  
804  
805 429 proposed 20-19 ka for the beginning of deglaciation, which was the time when most of  
806  
807 430 the northern ice sheets began to retreat, sea level and temperatures started to increase,  
808  
809 431 followed by an increase in the concentration of CO<sub>2</sub> in the atmosphere. Hughes et al.  
810  
811 432 (2013), in an exhaustive review of the chronology of the LLGM throughout the world,  
812  
813 433 show that not only did many mountain glaciers achieve their maximum extents before the  
814  
815 434 GLGM, but some northern ice sheets did as well. They acknowledge, however, the  
816  
817 435 fundamental role that the Laurentide Ice Sheet played in deglaciation, where the LLGM  
818  
819 436 broadly coincides with the GLGM.

### 814 815 437 *4.2 Laurentide Ice Sheet*

816  
817 438 The extent of glaciation during the LGM is summarized in Figure 3. The large extent of  
818  
819 439 the LIS was the result of planetary cooling, but its very existence also had an effect on  
820  
821 440 the evolution of mountain glaciers. The GLGM broadly coincides with the maximum size  
822  
823 441 of the Laurentide Ice Sheet (LIS) (Dyke et al., 2002; Clark et al., 2009; Stokes, 2017).  
824  
825 442 Despite the difficulty of precisely dating the maximum extent of the ice sheet, it is widely  
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443 accepted that different sectors of the LIS reached their local maxima at different times  
444 during the broad interval of the GLGM. For example, it has been suggested (Dyke et al.,  
445 2002) that the northwestern, northeastern and southern margins likely attained their  
446 maximum positions relatively early (~28-27 ka), whereas the southwestern and  
447 northernmost limits were probably reached slightly later (~25-24 ka). More recently,  
448 others have suggested that the northwestern margin, in the vicinity of the Mackenzie  
449 River delta, may have reached its maximum position at less than 20 ka (Murton et al.,  
450 2007; Kennedy et al., 2010; Lacelle et al., 2013) and possibly as late as 17-15 ka (Murton  
451 et al., 2015). If correct, this relatively late advance to a LLGM ice extent may have been  
452 aided by eustatic sea-level rise and the opening of the Arctic Ocean along the Beaufort  
453 Sea coastline, which provided a source of moisture and increased precipitation in the  
454 region (Lacelle et al., 2013).

455 Irrespective of the regional asynchronicity in the time of the local glacial maximum, it is  
456 likely that the LIS existed at its near-maximum extent for several thousand years, which  
457 would indicate that its mass balance was in equilibrium with the climate for a prolonged  
458 period of time (Dyke et al., 2002). Indeed, initial deglaciation is thought to have been  
459 slow prior to 17 ka (Dyke et al., 2002), and, as noted above, glaciers in some regions may  
460 have been advancing (e.g. in the far northwest). Possible exceptions to the generally slow  
461 recession include the major lobes of the southern margin of the ice sheet and the marine-  
462 based southeastern margin around the Atlantic Provinces. More rapid retreat of these  
463 margins was likely caused by ice-stream drawdown (Shaw et al., 2006, 2018; Margold et  
464 al., 2018) and, in the southeast, by eustatic sea-level rise (Dyke, 2004). In contrast, retreat  
465 of the land-based southern margin is thought to have been driven mainly by orbital forcing  
466 (Clark et al., 2009; Gregoire et al., 2015). Based on 22 <sup>10</sup>Be surface exposure ages on  
467 boulders on GLGM moraines in Wisconsin, Ullman et al. (2015a) dated the initial retreat  
468 of the ice sheet to as early as 23±0.6 ka, which coincided with a small increase in boreal  
469 summer insolation. <sup>10</sup>Be ages on samples 10-15 km up-ice from these moraines indicate  
470 a marked acceleration in retreat after ca. 20.5 ka that coincided with increased insolation  
471 prior to any increase in atmospheric carbon dioxide. This lends support to the notion that  
472 orbital forcing was the primary trigger for deglaciation of the LIS (see also Gregoire et  
473 al., 2015 and Heath et al., 2018).

474 Although increased insolation is thought to have triggered the initial retreat of the  
475 southern margin of the ice sheet (Ullman et al., 2015a), it is interesting to note that the



886  
887  
888  
889 476 overall net surface mass balance likely remained positive for much of the early part of  
890 477 deglaciation (Ullman et al., 2015b). However, the ice sheet was clearly shrinking, which  
891 478 implies that the primary mechanism of mass loss was dynamic discharge/calving from  
892 479 major marine-based ice streams (Margold et al., 2015, 2018; Ullman et al., 2015b; Robel  
893 480 and Tziperman, 2016; Stokes et al., 2016). Indeed, ~25% of the ice sheet's perimeter was  
894 481 occupied by streaming ice at the global LGM, compared to ~10% at 11 ka (Stokes et al.,  
895 482 2016). Only when summer temperatures increased by 6-7°C relative to the LGM did the  
896 483 overall net surface mass balance turn increasingly negative (Ullman et al., 2015b).  
897 484 Numerical modelling suggests that this occurred soon after ~11.5 ka and resulted in the  
898 485 rapid retreat of the land-based southern and western margins of the LIS (Ullman et al.,  
899 486 2015b). The rapid retreat of these terrestrial margins contrasts with the generally slow  
900 487 retreat of the northern and eastern marine-based margins and resulted in a highly  
901 488 asymmetric pattern of retreat towards the major dispersal centers in the east (Dyke and  
902 489 Prest, 1987; Margold et al., 2018).

#### 910 490 *4.3 Alaska*

911 491 Alaska is located at latitudes similar to the northern LIS, but was covered largely by  
912 492 mountain glaciers during the GLGM. Thus, it is of interest to understand its glacial  
913 493 evolution as a first link between the large ice sheet and mountain glaciers. The best  
914 494 available evidence from Alaska suggests that glaciers expanded during Marine Isotope  
915 495 Stage (MIS) 2 (the Late Wisconsinan glaciation in local terminology), in step with the  
916 496 GLGM. Although maximum ages constraining the advance phase are sparse, constraints  
917 497 on LGM culmination date to ~21 ka in several regions spanning the state, including the  
918 498 Ahklun Mountains (Kaufman et al., 2003), the Alaska Range (Tulenko et al., 2018), the  
919 499 Brooks Range and Arctic Alaska (Pendleton et al., 2015). The best available maximum  
920 500 age for the LGM glacier advance in Alaska – ~24 ka – is arguably from the Ahklun  
921 501 Mountains (Kaufman et al., 2003, 2012). Deglaciation in Alaska commenced as early as  
922 502 ~21 ka. Recognizing that cosmogenic nuclide exposure ages of moraine boulders  
923 503 represent the culmination of an advance, mean exposure ages of LGM terminal moraine  
924 504 boulders (~21 ka) mark the transition from maximum glacier conditions to ice retreat and  
925 505 terminal moraine stabilization. Moraines up-valley of terminal moraines were formed in  
926 506 the Ahklun Mountains (Manley et al., 2001), and marine sediments were deposited within  
927 507 LGM extents in Cooke Inlet (Reger et al., 2007) as early as ~20 ka. In the Alaska Range,  
928 508 the first moraines up-valley of the LGM terminal moraines were deposited ~20 ka

945  
946  
947 509 (Tulenko et al., 2018). In at least one or two valleys in the Brooks Range that are  
948  
949 510 accurately dated, glaciers receded well up-valley between ~21 ka and ~17 ka (Pendleton  
950  
951 511 et al., 2015).

952  
953 512 Climate conditions in Alaska during the GLGM are not well known, but several lines of  
954  
955 513 evidence indicate that conditions were much more arid than today (e.g. Finkenbinder et  
956  
957 514 al., 2014; Dorfman et al., 2015). Data on temperature changes during the LGM are scarce.  
958  
959 515 Some paleoecological evidence exists from the Brooks Range suggesting summer  
960  
961 516 temperatures 2-4°C colder than the present (Kurek et al., 2009), and pollen data from  
962  
963 517 across Beringia suggest summer temperatures were ~4°C lower (Viau et al., 2008). On  
964  
965 518 the other hand, climate modeling indicates rather warm conditions in Alaska during the  
966  
967 519 LGM, associated with persistent shifts in atmospheric circulation related to Laurentide  
968  
969 520 and Cordilleran ice sheet size (Otto-Bliesner et al., 2006; Löffverström and Liakka, 2016;  
970  
971 521 Liakka and Löffverström, 2018).

969 522 The largest gaps in knowledge regarding the timing of the LGM and initial deglaciation  
970  
971 523 in Alaska are related to the spatial pattern of glacier change across the state and complex  
972  
973 524 climate forcing. High-resolution chronologies from moraine sequences from single  
974  
975 525 valleys are scarce. Furthermore, few quantitative paleoclimate data exist, and the existing  
976  
977 526 records of glaciation and snowline depression have yet to be reconciled with climate  
978  
979 527 modeling results that show relatively warm LGM conditions.

#### 979 528 *4.4 Cordilleran Ice Sheet and North Cascades*

981 529 Glaciers in western Canada were expanding into lowland areas on the flanks of the Coast  
982  
983 530 and Rocky Mountains during the GLGM, contributing to development of the CIS  
984  
985 531 (Clague, 2017). The CIS was not fully formed at the GLGM; large areas of southern  
986  
987 532 British Columbia remained ice-free several thousand years later. Alpine glaciers in the  
988  
989 533 southern Coast Mountains advanced into lowlands near Vancouver, British Columbia,  
990  
991 534 after 25.8 ka during the Coquitlam stade in local terminology (Hicock and Armstrong,  
992  
993 535 1981; Hicock and Lian, 1995; Lian et al., 2001). To the south, alpine glaciers in the North  
994  
995 536 Cascades achieved their maximum MIS 2 extents between 25.3 ka and 20.9 ka, about the  
996  
997 537 same time as the GLGM (Kaufman et al., 2004; Riedel et al., 2010). The alpine advances  
998  
999 538 at these sites ended with the Port Moody interstade sometime after 21.4 ka, when glaciers  
1000  
1001 539 in the southern Coast Mountains and the North Cascades retreated (Hicock et al., 1982,  
1002  
1003 540 1999; Hicock and Lian, 1995; Riedel et al., 2010) (Fig. 8).

1004  
1005  
1006 541 Regional pollen and macrofossil data and glacier reconstructions indicate that the climate  
1007  
1008 542 that led to the alpine glacial advance in the North Cascades was the coldest and driest  
1009  
1010 543 period in MIS 2 (Barnosky et al., 1987; Thackray, 2001; Riedel et al., 2010). Glacier  
1011  
1012 544 ELAs fell by 750-1000 m from west to east across the range in response to a reduction in  
1013  
1014 545 mean annual surface air temperature of ~8°C and a significant reduction in precipitation  
1015  
1016 546 (Porter et al., 1983; Bartlein et al., 1998, 2011; Liu et al., 2009). The primary reasons for  
1017  
1018 547 the relatively arid climate were likely the lower sea surface temperatures in the Pacific  
1019  
1020 548 Ocean, the greater distance to the coastline and large-scale changes in the atmosphere  
1021  
1022 549 caused by formation of continental ice sheets (Hicoek et al., 1999; Grigg and Whitlock,  
1023  
1024 550 2002; Thackray, 2008). Paleoclimatic simulations produced by global climate models  
1025  
1026 551 suggest that three large-scale controls on climate have been especially important in the  
1027  
1028 552 Pacific Northwest during Late Glacial time (Broccoli and Manabe, 1987a, 1987b;  
1029  
1030 553 COHMAP Members, 1988; Bartlein et al., 1998; Whitlock et al., 2000). First, the  
1031  
1032 554 Laurentide Ice Sheet (LIS) influenced both temperature and atmospheric circulation.  
1033  
1034 555 Second, variations in the seasonal distribution of insolation as a result of the Earth's  
1035  
1036 556 orbital variations affected temperature, effective precipitation and atmospheric  
1037  
1038 557 circulation. Third, changes in atmospheric concentrations of CO<sub>2</sub> and other greenhouse  
1039  
1040 558 gases affected temperatures on centennial and millennial timescales (Sowers and Bender,  
1041  
1042 559 1995).

#### 1037 560 *4.5 Rocky Mountains/Yellowstone region*

1039 561 The Rocky Mountains allow us to link glacier behavior from the LIS to the north and the  
1040  
1041 562 CIS to the northwest with lower latitudes, where only small glaciers formed during the  
1042  
1043 563 period of maximum glacial expansion (Figs. 8, 9, 10, 11, 12, and 13). Recent ages from  
1044  
1045 564 Colorado confirm that a number of valley glaciers reached their LLGM extent ~21-20 ka  
1046  
1047 565 (Brugger et al., 2019) at roughly the same time as the GLGM (known locally as the  
1048  
1049 566 Pinedale Glaciation), while in other valleys glaciers continued to advance, re-advance or  
1050  
1051 567 remain in the same position for several thousand years until ~17 ka (see Brugger et al.,  
1052  
1053 568 2019, and references therein).

1052 569 In the Greater Yellowstone region, glaciers of the Beartooth Uplift and High Absaroka  
1053  
1054 570 Range appear to have reached their maximum extents ~20 ka (Licciardi and Pierce, 2018).  
1055  
1056 571 A similar pattern is evident on the eastern slope of the Teton Range, where the oldest  
1057  
1058 572 moraines date to 19.4±1.7 ka (Pierce et al., 2018). Differences in the ages of LLGM limits  
1059  
1060 573 in valleys surrounding the Yellowstone Plateau are likely due to local topographic factors

1063  
1064  
1065 574 at the margins of the Yellowstone Ice Cap rather than general climate forcing (Young et  
1066 al., 2011; Leonard et al., 2017a, 2017b; Pierce et al., 2018; Laabs et al., in preparation).  
1067 575  
1068  
1069 576 Ages of ~23-21 ka from terminal moraines of four valley glaciers in the Wind River  
1070 Range, about 150 km southeast of Yellowstone Park, show that the LLGM also generally  
1071 577  
1072 578 coincides with the GLGM (Phillips et al., 1997; Shakun et al., 2015a; Dahms et al., 2018).  
1073  
1074 579 Deglaciation seems to have been swift here; ice appears to have receded to 2.6 km behind  
1075 its terminus in the Middle Popo Agie valley by ~19 ka and 13 km upvalley from its  
1076 580  
1077 581 terminus in the adjacent North Fork valley by ~18-17 ka. Glaciers in both valleys  
1078  
1079 582 apparently receded 19 km and 27 km to their respective cirque riegels by 17-16 ka  
1080 583 (Dahms et al., 2018). The glacier in the Pine Creek valley receded nearly 30 km from its  
1081 584  
1082 584 terminus at Fremont Lake by 14-13 ka (Shakun et al., 2015a).  
1083  
1084 585 Many glaciers in the Rocky Mountains of Colorado reached their maximum extents  
1085 586 during the GLGM, with the outermost moraines abandoned ~22-20 ka (Ward et al., 2009;  
1086 587 Dühnforth and Anderson, 2011; Young et al., 2011; Schweinsberg et al., 2016; Leonard  
1088 588 2017a, 2017b; Brugger et al., 2019). In some cases, extensive deglaciation followed  
1089 589 shortly after 20 ka (Ward et al., 2009), but elsewhere glaciers remained at, or had re-  
1090 590 advanced to, near their maximum extents as late as 17-16 ka (Briner, 2009; Young et al.,  
1091 591 2011; Leonard et al., 2017a, 2017b), well after the end of the GLGM. In some instances,  
1092 592 these 17-16 ka moraines are the outermost moraines of the last glaciation.  
1093  
1094  
1095  
1096  
1097 593 The near complete absence of modern glaciers in the Colorado Rocky Mountains makes  
1098 594 it difficult to estimate ELA depressions at the GLGM, although in the San Juan Mountains  
1099  
1100 595 of southwestern Colorado it appears that they were lowered by at least 900 m (Ward et  
1101 596 al., 2009). Recent numerical modeling of paleo-glaciers in several Colorado ranges  
1102 597 indicates a rather modest GLGM temperature depression of 4.5°-6.0°C compared to  
1103 598 present-day temperatures, assuming no change in precipitation (Dühnforth and Anderson,  
1104 599 2011; Leonard et al., 2017a, 2017b). In contrast, work in the Mosquito Range suggests a  
1105 600 temperature depression of 7.5°-8.1°C (Brugger et al., 2019). Earlier work, using different  
1106 601 paleo-glaciological approaches, indicates somewhat greater GLGM temperature  
1107 602 depressions in the Colorado Rocky Mountains (Leonard, 1989, 2007; Brugger and  
1108 603 Goldstein, 1999; Brugger, 2006, 2010; Refsnider et al., 2008). Global and regional  
1109 604 climate models suggest that precipitation in the northern Rocky Mountains was  
1110 605 significantly reduced compared to the present. In contrast, the southernmost Rocky  
1111 606 Mountains in New Mexico were wetter at the GLGM than at present, and the central

1122  
1123  
1124 607 Rocky Mountains of Colorado and Wyoming experienced close to modern precipitation  
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1126 608 ([Oster et al., 2015](#)).

1127  
1128 609 *4.6 Sierra Nevada*

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1130 610 Few moraines from the early GLGM period in the Sierra Nevada have been directly dated,  
1131  
1132 611 perhaps because such moraines were less extensive than those built during the local  
1133 612 maximum and were thus obliterated by the later advances ([Phillips et al., 2009](#)). However,  
1134  
1135 613 there is abundant evidence of a cooling climate during the early GLGM from nearby  
1136 614 lacustrine records. For example, cores collected from Owens Lake, just east of the range  
1137  
1138 615 ([Smith and Bischoff, 1997](#)), record a rise in juniper pollen, which is considered an  
1139 616 indicator of cold temperature, between 30 ka and 25 ka, reaching a maximum between 25  
1140  
1141 617 ka and 20 ka ([Woolfenden, 2003](#)). In the same cores, total organic carbon, which  
1142  
1143 618 decreases as input of glacial rock flour increases, falls from about 4% to near zero  
1144  
1145 619 between 30 ka and 25 ka ([Benson, 1998a, b](#)). Similar patterns are observed in sediments  
1146 620 from Mono Lake ([Benson, 1998a, b](#)), which also received direct discharge from glaciated  
1147  
1148 621 valleys in the Sierra Nevada. The inference from lacustrine records that glaciation reached  
1149 622 near-maximum extent at about 25 ka is confirmed by a fortuitously preserved terminal  
1150  
1151 623 moraine in the valley of Bishop Creek, located at about 95% of the maximum LGM extent  
1152 624 and dated to  $26.5 \pm 1.7$  ka ([Phillips et al., 2009](#)) (Fig. 14).

1153  
1154 625 Cosmogenic and radiocarbon data for GLGM glaciation in the Sierra Nevada have  
1155 626 recently been compiled and updated by [Phillips \(2016, 2017\)](#). Both  $^{10}\text{Be}$  and  $^{36}\text{Cl}$  surface-  
1156  
1157 627 exposure dating yields ages ranging from 21 ka to 18 ka for the GLGM moraines (Tioga  
1158  
1159 628 3 in local terminology). Radiocarbon ages are slightly younger (19-18 ka), but this is  
1160  
1161 629 because they are on organic matter accumulated in depressions behind the Tioga 3  
1162 630 terminal moraines and thus date to the earliest stages of retreat. The spacing of recessional  
1163  
1164 631 moraines indicates that retreat was at first slow, but then accelerated ([Phillips, 2017](#)).

1165  
1166 632 In summary, glaciers advanced in the Sierra Nevada steadily after about 30 ka, achieving  
1167  
1168 633 positions slightly short of their maximum extents by 26 ka. They then were relatively  
1169  
1170 634 stable for the next 5 ka, but advanced slightly between 22 ka and 21 ka to their all-time  
1171  
1172 635 maximum limits of the last glacial cycle. Minor retreat from this maximum position began  
1173  
1174 636 at 19 ka and accelerated rapidly after 18.0 ka to 17.5 ka. [Plummer \(2002\)](#) attempted to  
1175  
1176 637 quantify both temperature and precipitation variations in the Sierra Nevada region during  
1177 638 the GLGM by simultaneously solving water and energy balance equations for glaciers  
1178  
1179 639 and closed-basin lakes. He concluded that precipitation during the peak LGM-maximum

1181  
1182  
1183 640 period (21-18 ka) was about 140% of historical levels and temperature was 5-6°C colder  
1184  
1185 641 than today.

#### 1186 1187 642 *4.7 Mexico and Central America*

1188  
1189 643 The highest mountains in Mexico and Central America were glacier covered during the  
1190 644 GLGM. There the LLGM overlaps part of the GLGM. In central Mexico, <sup>36</sup>Cl exposure  
1191 645 ages of moraines from the maximum advance are between 21 ka and 19 ka. Moraines  
1192 646 were deposited as late as 15-14 ka in the mountains near the Pacific (Tancítaro, 3840 m  
1193 647 asl) and Gulf of Mexico (Cofre de Perote, 4230 m asl), but minor recession occurred  
1194 648 around 17 ka in the interior (Iztaccíhuatl, 5286 m asl) (Vázquez-Selem and Heine, 2011).  
1195 649 Boulders on recessional moraines built inside the moraines from the local maximum have  
1196 650 yielded exposure ages between ~14.5 ka and >13 ka, and exposure ages on glacial polish  
1200 651 associated with recession range from 15 ka to 14 ka (Vázquez-Selem and Lachniet, 2017).  
1201  
1202 652 In Cerro Chirripó (3819 m asl), Costa Rica, the local maximum is ~25 ka to 23 ka based  
1203 653 on <sup>36</sup>Cl ages (Potter et al., 2019), whereas <sup>10</sup>Be exposure ages of lateral and recessional  
1204 654 moraines are between ca. 18.3 ka and ~16.9 ka (Cunningham et al., 2019). They are thus  
1205 655 younger than recessional moraines on mountains in central Mexico at a similar elevation  
1206 656 (e.g. Tancítaro, 3840 m asl). No ages exist for the glaciated Altos Cuchumatanes (3837  
1207 657 m asl) of Guatemala, although a maximum around the time of the GLGM is probable  
1208 658 based on data from central Mexico and Costa Rica (Roy and Lachniet, 2010).

1209  
1210  
1211 659 ELAs in the region during the LLGM were depressed 1000-1500 m compared to modern  
1212 660 values (equivalent to 6-9°C of cooling), which is consistent with ELA depression around  
1213 661 the world during the GLGM (Lachniet and Vázquez-Selem, 2005).

#### 1214 1215 662 *4.8 Northern Andes*

1216 663 A widespread advance in the Northern Andes during the GLGM is not clear, and the  
1217 664 limited chronological data available preclude robust interpretations. In the Venezuelan  
1218 665 Andes, temperatures during the GLGM have been estimated to be around 8°C cooler than  
1219 666 present, according to palynological analysis and a paleo-ELA reconstruction (Schubert  
1220 667 and Rinaldi, 1987; Stansell et al., 2007). Some outermost moraines have been dated to  
1221 668 around 21 ka in the Sierra Nevada based on <sup>10</sup>Be ages (modified ages from Angel, 2016;  
1222 669 updated ages from Carcaillet et al., 2013). The Las Tapias terminal moraine at 3100 m  
1223 670 asl in the Sierra Santo Domingo, northeastern Sierra Nevada, yielded ages of 18.2±1.0 ka  
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1242 671 (n=3) (Angel, 2016), and a glacier advance in the Cordillera de Trujillo has been dated to  
1243  
1244 672 around 17 ka (Bezada, 1989; Angel, 2016) (Fig. 15).

1245  
1246 673 Climate in the Colombian Andes during the GLGM was cold and dry (van Geel and van  
1247  
1248 674 der Hammen, 1973; Thouret et al., 1996). In this region, there are only a few ages from  
1249  
1250 675 scattered valleys and it is difficult to evaluate glacier extent during the GLGM. However,  
1251  
1252 676 in Páramo Peña Negra, close to Bogota, two moraine complexes between 3000 m and  
1253  
1254 677 3550 m asl were built between ~28 ka and 16 ka (Helmens, 1988). Paleo-ELAs were on  
1255  
1256 678 average 1300 m lower than modern, likely driven by 6–8°C colder temperatures (Mark  
1257  
1258 679 and Helmens, 2005). A till ('drift 3') on the western slopes of the Sierra Nevada del Cocuy  
1259  
1260 680 may date to the GLGM. The onset of sedimentation in Laguna Ciega, which is located on  
1261  
1262 681 this till at 2900 m asl has been radiocarbon-dated to ca. 27.0-24.5 ka BP (van der Hammen  
1263  
1264 682 et al., 1981).

1265  
1266 683 The glacial chronology of the Ecuadorian Andes is poorly constrained and does not allow  
1267  
1268 684 clear conclusions to be drawn about glacier extent. There are some indications of possible  
1269  
1270 685 advances around the time of the GLGM, such as in the Rucu Pichincha and the Papallacta  
1271  
1272 686 valley (Heine and Heine, 1996) and in Cajas National Park (Hansen et al., 2003).  
1273  
1274 687 Brunschön and Behling (2009) suggest that climate was cold and wet during the GLGM  
1275  
1276 688 in the southern Ecuadorian Andes based on a pollen record and the upper timberline  
1277  
1278 689 position in Podocarpus National Park.

#### 1274 690 *4.9 Peruvian and Bolivian Andes*

1276 691 Evidence for the extent and chronology of past glacier advances in Peru and Bolivia at  
1277  
1278 692 the GLGM comes from moraine chronologies and lake sediment records that provide a  
1279  
1280 693 suite of ages before and after 21 ka (Clayton and Clapperton, 1997; Blard et al., 2013).  
1281  
1282 694 The time of the local maximum glacier expansion, based on the average cosmogenic ages  
1283  
1284 695 of moraine groups, is ~25 ka, but there are large uncertainties (up to 7 ka), making the  
1285  
1286 696 exact time of the LLGM uncertain. It also remains uncertain whether LLGM moraines  
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1288 697 were constructed during a long still-stand or a re-advance that erased the previous  
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1290 698 maximum limit (Mark et al., 2017). A close examination of site records reveals that,  
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1292 699 although the LLGM was close to the GLGM, there was, in some places, a larger local  
1293  
1294 700 maximum extension of glaciers before the GLGM (Farber et al., 2005; Smith et al., 2005;  
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1296 701 Rodbell et al., 2008). In the southern part of the Altiplano, maximum glacier extents of  
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1298 702 the last glaciation are probably as old as 60 ka (e.g. Blard et al., 2014) (Figs. 16 and 17).  
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1300 703 Lakes Titicaca and Junin, which are outside glacial moraines, have provided sediment

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1301 704 records that indicate deglaciation was underway by 22-19.5 ka (Seltzer et al., 2000, 2002;  
1302 705 Baker et al., 2001a, 2001b; Rodbell et al., 2008). On the Coropuna volcano, located in  
1303 706 southern Peru, <sup>3</sup>He ages indicate that the LLGM happened ~25 ka and deglaciation began  
1304 707 at ~19 ka (Bromley et al., 2009).

1308 708 Temperatures decreased ~6° C during the GLGM in the Peruvian and Bolivian Andes  
1309 709 (Mark et al., 2005), and precipitation was slightly higher than today, as indicated by the  
1310 710 Sajsi paleo-lake cycle (Seltzer et al., 2002; Blard et al., 2011, 2013). Therefore, a  
1311 711 temperature increase was probably the main driver of deglaciation between 19 ka and 17  
1312 712 ka. However, precipitation variations likely played an important role in some regions  
1313 713 where a late deglaciation is reported, such as in the vicinity of the paleo-lake Tauca, on  
1314 714 the central Altiplano (Martin et al., 2018).

#### 1319 715 *4.10 Southern Bolivia and Northern Chile*

1322 716 Glacier extent at the GLGM in the western cordillera of the Andes, adjacent to the Arid  
1323 717 Diagonal, is unclear. Glacial deposits and landforms north of the Arid Diagonal that have  
1324 718 been investigated include those at Cerro Uturunco (Blard et al., 2014), El Tatio and  
1325 719 Sairecabur (Ward et al., 2017), and Cerro La Torta and the Chajnantor Plateau (Ward et  
1326 720 al., 2015). Deposits in the subtropics south of the Arid Diagonal include those in Valle  
1327 721 de Encierro (Zech et al., 2006) and Cordón de Doña Rosa (Zech et al., 2007). At most  
1328 722 sites north of the Arid Diagonal, a set of degraded moraines lies 2-5 km outside one or  
1329 723 two sets of closely nested, sharper-crested moraines, which in turn are outside smaller  
1330 724 younger up-valley moraines (Jenny et al., 1996). A few <sup>10</sup>Be and <sup>36</sup>Cl ages suggest that  
1331 725 the outer degraded moraines date to MIS 6 (191-130 ka; Ward et al., 2015). Greater  
1332 726 precision is not possible with available data, but this interval corresponds to the age of a  
1333 727 broad bajada along the Salar de Atacama based on <sup>36</sup>Cl ages on terrace surfaces and a  
1334 728 depth profile (Cesta and Ward, 2016).

1343 729 A set of more prominent moraines inside these degraded moraines marks the maximum  
1344 730 expansion of glaciers after MIS 6 (Ward et al., 2017). Widely scattered <sup>10</sup>Be and <sup>36</sup>Cl  
1345 731 ages, ranging from about 90 to 20 ka (45-35 ka modal age), have been obtained from  
1346 732 boulders on these moraines both north and south of the Arid Diagonal (Ward et al., 2017)  
1347 733 (Fig. 18). Eight boulders on the sharp crest of the LLGM moraine at El Tatio yielded six  
1348 734 <sup>36</sup>Cl ages between 41 ka and 19.8 ka, with outliers at 82 and 57 ka. At Cerro La Torta,  
1349 735 one LLGM moraine boulder yielded a <sup>10</sup>Be age of 24.7±1.8 ka and glaciated bedrock just



1358  
1359  
1360 736 inside the LLGM limit returned a  $^{10}\text{Be}$  exposure age of  $31\pm 2.4$  ka. Similar ages have been  
1361  
1362 737 obtained from the terminal moraines at Cerro Uturuncu, south of paleolake Tauca on the  
1363 738 Bolivian Altiplano, with 8 of 12  $^3\text{He}$  ages between 46 and 33 ka (Blard et al., 2014).  
1364  
1365 739 Similarly, a single boulder on the outer terminal moraine in Encierro Valley yielded a  
1366 740  $^{10}\text{Be}$  age of  $35\pm 2$  ka (Zech et al., 2006), and 9 of 13  $^{10}\text{Be}$  samples from the outermost  
1367  
1368 741 moraines, drift, and outwash at Cordón de Doña Rosa returned ages ranging from 49 to  
1369  
1370 742 36 ka (Zech et al., 2007).

1371  
1372 743 If the local LGM moraines date to 49-35 ka, they were built at about the same time as the  
1373  
1374 744 Incahuasi highstand, during which a deep lake formed in the Pozuelos Basin in Argentina  
1375 745 (McGlue et al., 2013), and during a period when glaciers in the subtropical Argentine  
1376  
1377 746 Andes expanded (see Section 4.11).

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1379  
1380 747 Exposure ages on bedrock inside the prominent LLGM moraines (Blard et al., 2014;  
1381 748 Ward et al., 2015) indicate that deglaciation was underway by 20-17 ka. Assuming these  
1382  
1383 749 ages are valid, deglaciation of the western cordillera in northern Chile may have preceded  
1384  
1385 750 that of the Altiplano.

1386  
1387 751 The scatter in cosmogenic ages on moraines in this region may be due to differences in  
1388  
1389 752 dating methods.  $^{36}\text{Cl}$  production is environmentally sensitive, and production rates are  
1390 753 less certain than those for  $^{10}\text{Be}$ . However,  $^{10}\text{Be}$  ages on the same features also exhibit  
1391  
1392 754 scatter (Ward et al., 2015). For example, the LLGM moraines bordering the former 200  
1393 755  $\text{km}^2$  ice cap on the Chajnantor Plateau (4500-5500 m asl) have yielded both  $^{10}\text{Be}$  and  $^{36}\text{Cl}$   
1394  
1395 756 ages ranging from 141 to 43 ka. However,  $^{10}\text{Be}$  and  $^{36}\text{Cl}$  exposure ages on glaciated  
1396  
1397 757 bedrock beneath the most prominent moraines are younger and less scattered (30-18 ka),  
1398 758 and one boulder on a small moraine  $\sim 1$  km inboard of the LLGM margin yielded an age  
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1400 759 of  $26.7\pm 2.8$  ka, similar to the bedrock ages (Ward et al., 2017). Additionally, the youngest  
1401  
1402 760 bedrock exposure ages (20-18 ka) are from downvalley sites, near the terminal moraines,  
1403 761 whereas ages higher on the plateau are older (30-26 ka) (Ward et al., 2015). This pattern  
1404  
1405 762 cannot be explained by retreat of the glacier margin; rather it suggests that the LLGM  
1406 763 moraines contain a significant component of older reworked material with cosmogenic  
1407  
1408 764 inheritance. It is also consistent with the lesser, but still considerable, scatter seen in the  
1409  
1410 765 ages on LLGM valley glacier moraines in the region.

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1418  
1419 766 Reliable estimates of temperature and precipitation in northern Chile during the GLGM  
1420  
1421 767 will require more precise dating of the glacial deposits there. [Kull and Grosjean \(2000\)](#)  
1422 768 performed glacier-climate modeling to reconstruct precipitation associated with  
1423  
1424 769 construction of the major sharp-crested moraine at the El Tatio site. They assumed a  
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1426 770 regional temperature depression of  $\sim 3.5^{\circ}\text{C}$ , consistent with that at ca. 17 ka, and  
1427 771 concluded that an additional 1000 mm/yr of precipitation over modern would be required  
1428  
1429 772 to generate a glacier of the appropriate size. If instead the sharp-crested El Tatio moraines  
1430 773 date to the GLGM, as suggested by [Ward et al. \(2017\)](#), temperatures were likely 5-7 C  
1431  
1432 774 lower than today and less precipitation would be required. For example, assuming a 5.7 C  
1433  
1434 775 temperature depression typical of the GLGM in this area, [Kull et al. \(2002\)](#) estimated that  
1435 776 a  $580\pm 150$  mm/yr increase over modern precipitation would be required to explain the  
1436  
1437 777 LLGM deposits at a different western cordillera site (Encierro Valley).

#### 1438 1439 778 *4.11 Central Andes of Argentina*

1440  
1441 779 The maximum expansion of glaciers in the Argentine Andes occurred before the GLGM,  
1442  
1443 780 between 50-40 ka and before 100 ka ([Zech et al., 2009, 2017](#); [Martini et al., 2017a](#); [Luna](#)  
1444 781 [et al., 2018](#); [D’Arcy et al., 2019](#)). However, there was a generalized glacier expansion  
1445  
1446 782 during the GLGM between 22° and 35° S (Fig. 19). North of the Arid Diagonal, the  
1447 783 LLGM is dated to 25-20 ka based on an average of 10  $^{10}\text{Be}$  ages on both sides of Nevado  
1448  
1449 784 de Chañi ([Martini et al., 2017a](#)). The advance on the east side of Nevado de Chañi was  
1450  
1451 785 less pronounced than that on the west side. Glaciers advanced between  $\sim 22$  ka and  $\sim 19$   
1452 786 ka in the Laguna Grande valley and at  $\sim 20$  ka in the Peña Negra valley, both in the Tres  
1453  
1454 787 Lagunas area ([Zech et al., 2009, 2017](#)). M2 moraines in the Sierra de Aconquija were  
1455 788 built at  $\sim 22$  ka ([D’Arcy et al., 2019](#)). There are no moraines firmly dated to the GLGM  
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1457 789 in the Sierra de Quilmes ([Zech et al., 2017](#)), but pronounced undated lateral moraines in  
1458  
1459 790 the Nevado del Chuscha valley might be of that age. Based on the geomorphology and  
1460 791 chronology of the moraine sequence in the same valley, [Zech et al. \(2017\)](#) concluded that  
1461  
1462 792 these lateral moraines must have been deposited between 44 ka and 18 ka.

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1464 793 There is no consensus about precipitation levels in the subtropical Andes north of the  
1465 794 Arid Diagonal during the GLGM. Available evidence from the nearby arid Altiplano  
1466  
1467 795 suggests climate was only moderately wetter than present ([Baker et al., 2001a, 2001b](#);  
1468 796 [Placzek et al., 2006](#)). Speleothem records from the western Amazon, the Peruvian Andes,  
1469  
1470 797 the Pantanal and southeastern Brazil all indicate wetter conditions during the GLGM

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1477  
1478 798 (Cruz et al., 2005; Wang et al., 2007; Kanner et al., 2012; Cheng et al., 2013; Novello et  
1479 al., 2017) due to an intensification of the South American summer monsoon.  
1480 799

1481  
1482 800 The glacial chronology south of the Arid Diagonal is poorly constrained. Moraines  
1483 801 coincident with the GLGM have been found in the Ansilta range and Las Leñas valley.  
1484 802 Lateral moraines in the Ansilta range have been dated to 28-19 ka based on four <sup>10</sup>Be  
1485 803 ages, and a prominent lateral moraine in Las Leñas valley was built between 22 ka and  
1486 804 20 ka (Terrizano et al., 2017). Other possible evidence of GLGM glacial activity comes  
1488 805 from the Cordon del Plata range, where one boulder on the Agostura I moraine was dated  
1490 806 to 19 ka (Moreiras et al., 2017). Two <sup>10</sup>Be ages (31 ka and 23 ka) on a moraine close to  
1491 807 Nahuel Huapi lake, near Bariloche in northern Patagonia, suggest a GLGM age (Zech et  
1492 808 al., 2017). An end moraine in the Rucachoroi valley yielded two <sup>10</sup>Be ages and Zech et  
1493 809 al. (2017) assigned an age to an end moraine in the Rucachoroi valley to 21 ka based on  
1494 810 two <sup>10</sup>Be ages. South of the Arid Diagonal, there is evidence of wetter conditions during  
1495 811 the GLGM compared to today (Kaiser et al., 2008; Moreno et al., 2018).  
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#### 1501 812 *4.12 Patagonia*

1502  
1503 813 The time of the LLGM of the Patagonian Ice Sheet (PIS) is, unsurprisingly, variable,  
1504 814 given the broad latitudinal range of the Patagonian Andes (38°–55°S). In most cases,  
1505 815 Patagonian glaciers achieved their maximum extents earlier than the GLGM, during MIS  
1506 816 3 (Darvill et al., 2015; Garcia et al., 2018). Detailed stratigraphic and chronologic data  
1507 817 exist in the Chilean Lake District (41°S) on the northwest side of the former ice sheet  
1508 818 (Denton et al., 1999a; Moreno et al., 2015, 2018). Here, multiple radiocarbon-based  
1509 819 chronologies bracket the time of local major expansions of piedmont lobes at ~33.6,  
1510 820 ~30.8, ~26.9, ~26 and 17.8 ka (Denton et al., 1999a; Moreno et al., 2015). There is a  
1511 821 significant gap in glacial chronologies for the area between 41° and 46°S, except for the  
1512 822 Cisnes valley (44°S) where moraines dating to the end of the GLGM (<sup>10</sup>Be mean age ~20  
1513 823 ka) are inside more distal moraines that are assumed to date to earlier phases of the last  
1514 824 glacial cycle (de Porras et al., 2014; Garcia et al., 2019). However, the more distal  
1515 825 moraines are undated, consequently it remains unclear whether or not the pattern of more  
1516 826 extensive MIS 3 advances persists southward in central Patagonia. Farther south,  
1517 827 additional studies have been done in the area currently occupied by the cross-border lakes  
1518 828 of Lago General Carrera/Buenos Aires (46.5°S) and Lago Cochrane/Pueyrredón (47.5°S).  
1519 829 In the former area, ages of ~26 ka have been obtained for the local maximum extent of  
1520 830 the PIS (Kaplan et al., 2004, 2011; Douglass et al., 2006), coincident with the GLGM.  
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1537 831 However, earlier glacial activity, at 34-31 ka, is suggested by Optically stimulated  
1538  
1539 832 Luminescence (OSL) ages on buried sediments (Smedley et al., 2016). In the latter area  
1540  
1541 833 (Lago Cochrane/Pueyrredón), the LLGM has been dated at ~29 ka, and possibly ~35 ka,  
1542  
1543 834 with moraines of the GLGM located immediately up-ice (Hein, 2009, 2010, 2017).

1544 835 Exposure dating in southern Patagonia indicates that the LLGM was far more extensive  
1545  
1546 836 than subsequent GLGM advances. For example, the Bahía Inútil–San Sebastián ice lobe  
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1548 837 (53°S) expanded 100 km farther at ~45 ka and ~30 ka (Darvill et al., 2015a) than later  
1549  
1550 838 advances during the GLGM at ~20 ka (McCulloch et al., 2005a; Kaplan et al., 2008). The  
1551  
1552 839 pattern is repeated farther north where the Torres del Paine and Última Esperanza ice  
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1554 840 lobes (51°S) reached their local maximum extents at ~48 ka, with subsequent advances  
1555  
1556 841 dated to 39.2 ka and 34 ka, and a far less extensive GLGM advance at 21.5 ka (Sagredo  
1557  
1558 842 et al., 2011; García et al., 2018). Single exposure ages from the San Martín valley (49°S)  
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1560 843 tentatively suggest local maximum glacier expansion at ~39 ka, with a less extensive  
1561  
1562 844 GLGM advance at ~24 ka (Glasser et al., 2011).

1561 845 Considered together, the chronologies demonstrate that the LLGM in Patagonia occurred  
1562  
1563 846 at different times, but largely during MIS 3. Presently, there is no satisfactory mechanism  
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1565 847 to adequately explain the timing of this local glacial maximum, although possible  
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1567 848 explanations include regional insolation and coupled ocean-atmosphere interactions,  
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1569 849 including the influence of the southern westerly winds, sea surface temperatures,  
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1571 850 Southern Ocean stratification and Antarctic sea ice extent (Darvill et al., 2015a, 2016;  
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1573 851 Moreno et al., 2015; García et al., 2018). Compared to the LLGM, the onset of  
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1575 852 deglaciation is more closely coupled throughout Patagonia and centered at 17.8 ka with  
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1577 853 some local variation, which is concurrent with warming of the mid to high latitudes in the  
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1579 854 Southern Hemisphere (Kaplan et al., 2004, 2007; McCulloch et al., 2005a; Douglass et  
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1581 855 al., 2006; Hein et al., 2010, 2017; Sagredo et al., 2011; Murray et al., 2012; García et al.,  
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1583 856 2014, 2019; Henríquez et al., 2015; Moreno et al., 2015, 2018, 2019; Bendle et al., 2017;  
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1585 857 Mendelova et al., 2017; Vilanova et al., 2019).

#### 1582 858 *4.13 Tierra del Fuego*

1584 859 Caldenius (1932) constructed the first map of the Darwin ice field at the LLGM. The map  
1585  
1586 860 has not been greatly modified since that time, and the exact position of the ice limits  
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1588 861 around large parts of the Cordillera Darwin are poorly constrained. Former ice extent is  
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1590 862 best understood where glaciers flowing northeastward from the mountains contributed to

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1595  
1596 863 extensive lobes in the Straits of Magellan and Bahía Inútil (Clapperton et al., 1995;  
1597 864 Rabassa et al., 2000; Bentley et al., 2005; McCulloch et al., 2005a; Coronato et al., 2009;  
1598 865 Darvill et al., 2014) (Fig. 20). Surface exposure ages of glacial landforms in Tierra del  
1600 866 Fuego suggest that these lobes achieved their maximum extents by ~25 ka and remained  
1601 866 there until ~18 ka (McCulloch et al., 2005b; Kaplan et al., 2008; Evenson et al., 2009).  
1602 867  
1603 867 However, several belts of ice-marginal landforms occur outside these moraines  
1604 868 (Caldenius, 1932; Clapperton et al., 1995; McCulloch et al., 2005a; Evenson et al., 2009),  
1605 869 and existing exposure age data have yielded conflicting results. Some of these outer  
1606 870 moraines have been assigned pre-GLGM ages, but an analysis of weathering of erratic  
1607 871 boulders suggests that most, if not all, of them may date to the last glaciation (Darvill et  
1608 872 al., 2015b). On the southern flank of the Cordillera Darwin, outlet glaciers formed an ice  
1609 873 stream in Beagle Channel that terminated near the Atlantic Ocean (Caldenius, 1932;  
1610 874 Rabassa et al., 2000, 2011; Coronato et al., 2004, 2009), but remains undated. Moreover,  
1611 875 there is no convincing evidence on the Pacific Coast for the position of the GLGM ice-  
1612 876 sheet margin, and reconstructions range from extensive ice on the continental shelf  
1613 877 (Caldenius, 1932) to ice terminating close to the present-day shoreline (Coronato et al.,  
1614 878 2009). Given the uncertainty in GLGM positions around most of the margin, the time of  
1615 879 the onset of glacier recession is difficult to pinpoint. However, on both the north and south  
1616 880 sides of the range, radiocarbon ages from bog sediments, as well as a limited number of  
1617 881 exposure ages from erratics, indicate that glaciers had receded to the interior of the  
1618 882 mountains by ~17 ka (Heusser, 1989; Hall et al., 2013; Menounos et al., 2013) (Fig. 20).  
1619 883

#### 1620 884 *4.14 Synthesis*

1621 885 Based on current understanding, glaciers in North and Central America during the GLGM  
1622 886 (Table 1 and Fig. 3) appear to have fluctuated near-synchronously and likely responded  
1623 887 to the same climate drivers. In many sectors, glaciers achieved their LLGM extents  
1624 888 around 26-21 ka. In some cases, glacier fronts remained stable from that time until shortly  
1625 889 after 21 ka, when deglaciation began. This was the case for most of the LIS and for  
1626 890 glaciers in Alaska, the North Cascades, several valleys in the Rocky  
1627 891 Mountain/Yellowstone region, the Sierra Nevada, Central Mexico, and the Cordillera de  
1628 892 Talamanca in Costa Rica.

1629 893 Key climate forcing common to all these regions is the decrease in temperature during  
1630 894 the GLGM. Based on a decrease in ELAs of approximately 900 m, temperatures  
1631 895 decreased by approximately 7-8°C across much of the North American continent.

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896 However, there are some differences. For example, the ELA depression in Alaska was  
897 less than 500 m, and the corresponding summer temperature depression was likewise less  
898 than in the western US. The pattern of precipitation during the GLGM apparently was  
899 even less uniform. Evidence shows a trend towards aridity during the GLGM in the North  
900 Cascades close to the ice sheet and the northern Rocky Mountains, and increased  
901 precipitation to the south in the Sierra Nevada, Basin and Range Province and southern  
902 Rocky Mountains.

903 We note that the behavior of glaciers during the GLGM in North and Central America  
904 was also asynchronous. Several glaciers advanced to their maximum positions several  
905 thousand years after the GLGM, at about the time of the HS-1 period. This is the case for  
906 some sectors of the LIS and CIS, some ranges in southern Alaska, some areas close to  
907 Yellowstone, the Colorado Rocky Mountains, mountains of central Mexico near the  
908 oceans, and some valleys of the Cordillera de Talamanca in Costa Rica. Differences in  
909 glacier activity within the same region could be due to local differences in precipitation  
910 stemming from orographic effects, for example in some areas of the Yellowstone region,  
911 or between oceanic and interior mountains in Mexico. Whether or not the relationship  
912 between precipitation and the glacial local maximum is generally applicable for the entire  
913 continent is a subject for future research.

914 The relative consistency in glacier behavior across North and Central America is not  
915 observed in South America. The lack of synchronicity in glacier growth in the Andes  
916 might possibly be due to the relative scarcity of data in the region or, alternatively, to its  
917 large latitudinal range and complex geography, which lead to large differences in  
918 precipitation. The most arid regions of the southern tropical Andes (southern Bolivia,  
919 northern Chile and Argentina) show the largest temporal variability in the time of the  
920 LLGM, probably due to strong precipitation control. In any case, the maximum local  
921 expansion of the glaciers in most areas in the Andes does not coincide with the GLGM.  
922 One of the few exceptions is in Tierra del Fuego, where glaciers may have reached their  
923 maximum extents between ~25 ka and ~18 ka. Even there, however, future work may  
924 show that moraines down-ice of this limit may also date to the last glaciation. In the rest  
925 of the Andean Cordillera, moraines were built during the GLGM, but the maximum  
926 advance apparently happened up to several thousands of years earlier; in southern  
927 Patagonia the LLGM may have occurred during MIS 3 as few other southern high latitude  
928 regions such as Kerguelen (Jomelli, et 2018). We also note that the moraines that

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1714 929 coincide with the GLGM are not necessarily the largest, as is commonly the case in North  
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1716 930 America where glacier fronts remained in the same position for an extended period of  
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1718 931 time.

1719 932 Glaciers in the central part of the Altiplano, in the vicinity of paleo-lake Tauca, remained  
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1721 933 close to their LLGM positions until the end of H-1 (Martin et al., 2018). Elevated  
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1723 934 precipitation during H-1 apparently sustained glaciers until the end of that period. In  
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1725 935 summary, throughout the Andes, the GLGM seem to be marked by an expansion of  
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1727 936 glaciers, but that advance was not the largest everywhere. Across the Andes, this period  
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1729 937 coincided with a clear drop in temperature of ~3-8°C based on ELA depressions. Those  
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1731 938 values are consistent with temperature reductions inferred from ELA depressions in North  
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1733 939 America. Some local indicators, for example the Sajsi paleo-lake on the Altiplano show  
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1735 940 that the GLGM was characterized by slightly higher precipitation than today (Placzek et  
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1737 941 al., 2006).

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## 1738 943 **5. The Impact of Heinrich-1 Stadial (HS-1) (17.5-14.6 ka) on American Glaciers**

### 1740 944 *5.1 Heinrich-1 Stadial*

1742 945 The second period analyzed is the Heinrich 1 Stadial (HS-1), which is called the ‘Oldest  
1743  
1744 946 Dryas’ in Scandinavia. The term HS-1 comes from records of marine sediments that show  
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1746 947 the massive discharge of icebergs into the North Atlantic during this period (Heinrich,  
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1748 948 1988), mainly from the Hudson Bay/Strait region, the main drainage route for the LIS  
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1750 949 (Hemming, 2004). The use of the term as a chronological unit has been criticized  
1751  
1752 950 (Andrews and Voelker, 2018) from a sedimentological point of view. The term Oldest  
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1754 951 Dryas, although widely used, has also been criticized because it is not clearly delimited  
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1756 952 chronologically (Rasmussen et al., 2014). In this paper, we follow the paleoclimate and  
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1758 953 paleoglaciological criteria of Denton et al. (2006), who delimit HS-1 between the  
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1760 954 Heinrich 1 “event” (17.5 ka) and the beginning of the Bølling-Allerød interstadial (14.6  
1761  
1762 955 ka). They refer to this period as the ‘Mystery Interval’ due to the fact that, although CO<sub>2</sub>  
1763  
1764 956 concentrations in the atmosphere increased during this time, temperature dropped sharply  
1765  
1766 957 in the Northern Hemisphere and in the tropics. In our study, we opt for the term HS-1 for  
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1768 958 the same time period, following the standard differentiation between “event” and  
1769  
1770 959 “stadial” (Rasmussen et al., 2014; Heath et al., 2018).

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960 HS-1 is a climate event that interrupted deglaciation. In the North Atlantic region,  
961 temperatures fell drastically in winter, sea ice expanded, and the ocean cooled (Barker et  
962 al., 2010). Atlantic Meridional Overturning Circulation (AMOC) was sharply reduced or  
963 even collapsed (McManus et al., 2004; Böhm et al., 2015), and many mountain glaciers  
964 advanced in Europe (Gschnitz stadial in the Alps; Ivy-Ochs, 2015), at least at the  
965 beginning of HS-1. Although the European ice sheets decreased in size during this period  
966 (Toucanne et al., 2015), it is clear that climate during HS-1 varied. There were periods  
967 with hot summers that caused massive glacier melting (Thornalley et al., 2010; Williams  
968 et al., 2012). The Asian monsoon disappeared (Wang et al., 2008), the South American  
969 monsoon intensified (Strikis et al., 2015, 2018), and the Southern Hemisphere westerlies  
970 were displaced polewards (Denton et al., 2010). Temperatures in Antarctica increased,  
971 along with atmospheric CO<sub>2</sub> concentrations (Monnin et al., 2001; Ahn et al., 2012), due  
972 to Southern Ocean ventilation (Barker et al., 2009). HS-1 is the period that best  
973 demonstrates the close relationships among AMOC, atmospheric CO<sub>2</sub> and temperatures  
974 in Antarctica (Deaney et al., 2017).

## 975 *5.2 Laurentide Ice Sheet*

976 The extent of glaciers and ice sheets during HS-1 is summarized in Figure 4. Although  
977 explanations of Heinrich events have tended to focus on the Hudson Strait ice stream, it  
978 is clear that there are sedimentological differences both within and between individual  
979 Heinrich ‘layers,’ including variable source areas (Andrews et al., 1998, 2012; Piper and  
980 Skene, 1998; Hemming, 2004; Tripsanas and Piper, 2008; Rashid et al., 2012; Roger et  
981 al., 2013; Andrews and Voelker, 2018). Thus, it is likely that other ice streams along the  
982 eastern margin of the LIS, and possibly even farther afield at its northern margin (Stokes  
983 et al., 2005), may have contributed, at least in part, to some Heinrich-like events  
984 (Andrews et al., 1998, 2012; Piper and Skene, 1998). However, the extent to which these  
985 events were correlative is unclear, as are the wider impacts of Heinrich events on the  
986 dynamics of the LIS. For example, readvances or stillstands elsewhere in the Americas  
987 have been linked to HS-1, and yet evidence from the LIS is comparatively scarce.

988 Clark (1994) was one of the first to propose a link between Heinrich events in the Hudson  
989 Strait and the advance of ice margins/lobes resting on soft deformable sediments along  
990 the southern margin of the ice sheet. Mooers and Lehr (1997) also noted the possibility  
991 that the advance and rapid retreat of lobes in the western Lake Superior region may have  
992 been correlative with Heinrich events 2 and 1, but this idea has since received relatively



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993 little attention and there is little clear evidence for major re-advances of the LIS during  
994 or soon after HS-1 (Heath et al., 2018). Rather, the most likely impact of HS-1 was to  
995 lower the ice surface over Hudson Bay and drive changes in the location of ice dispersal  
996 centers, with subsequent effects on ice-flow patterns (Margold et al., 2018). For example,  
997 Dyke et al. (2002) suggest that the drawdown of ice during HS-1 was likely sufficient to  
998 displace the Labrador ice divide some 900 km eastward from the coast of Hudson Bay  
999 and cause a major flow reorganization (see also Veillette et al., 1999). There is also  
1000 evidence that parts of the ice sheet thinned rapidly in coastal Maine during the latter part  
1001 of HS-1 (Hall et al., 2017b; Koester et al., 2017).

1002 There is also clear evidence from several regions that the ice sheet retreated during HS-  
1003 1, punctuated by brief readvances or stillstands. For example, recalculated  $^{10}\text{Be}$  data  
1004 (Balco and Schaefer, 2006), coupled with the New England varve chronology (Ridge et  
1005 al., 2004), indicate retreat of the ice margin in the northeastern United States.  $^{36}\text{Cl}$   
1006 exposure ages from the Adirondack Mountains (Barth et al., 2019) suggest that the ice  
1007 sheet may have begun to thin around  $19.9\pm 0.5$  ka. Thinning continued throughout HS-1  
1008 and accelerated between  $15.5\pm 0.4$  ka and  $14.3\pm 0.4$  ka (see also Section 5.2). Rapid ice  
1009 sheet thinning has also been inferred in coastal Maine during the latter part of HS-1 (Hall  
1010 et al., 2017b; Koester et al., 2017).

### 1011 5.3 Alaska

1012 Although detailed moraine chronologies needed to fully explain glacier change in Alaska  
1013 during HS-1 do not exist, there is patchy information on ice extent at that time. In most  
1014 locations where recessional moraines have been dated, some stillstands or re-advances  
1015 have been inferred during HS-1. In the Brooks Range, a prominent recessional moraine  
1016 has been dated to  $\sim 17$  ka (Pendleton et al., 2015), and the Elmendorf Moraine in south-  
1017 central Alaska dates to  $\sim 16.5$  ka (Kopczynski et al., 2017). Given the number of  
1018 recessional moraines in most valleys, for example throughout the Alaska Range, the  
1019 Ahklun Mountains, and the Kenai Peninsula, it is difficult to know if these glacial  
1020 stabilizations necessarily relate to cooling triggered in the North Atlantic Ocean. Rather,  
1021 they could be related to a number of factors that could cause glacier recession to be  
1022 interrupted by re-advances or stillstands (e.g. isostatic rebound, solar variability, glacier  
1023 hypsometric effects). Thus, attributing them *per se* to North Atlantic stadial conditions at  
1024 this time is premature. In fact, in spite of some interruptions, there was overall significant

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1025 recession of glaciers throughout HS-1 in Alaska. Most glaciers in Alaska with reasonable  
1026 chronological constraints experienced net retreat during HS-1.

#### 1027 *5.4 Cordilleran Ice Sheet and the North Cascades*

1028 Alpine glaciers receded from maximum positions during the Port Moody interstade,  
1029 which began after 21.4 ka (Riedel et al., 2010). Two glacial events in this region correlate  
1030 with HS-1: construction of alpine glacier end moraines and the advance of the CIS to its  
1031 maximum limit. Deposition of ice-rafted detritus at a deep-sea core site west of  
1032 Vancouver Island began about 17 ka and abruptly terminated at about 16.2 ka, recording  
1033 the rapid advance and retreat of the western margin of the CIS (Cosma et al., 2008).  
1034 Studies west of Haida Gwaii (Blaise et al., 1990) and near the southwestern margin of the  
1035 CIS (Porter and Swanson, 1998; Troost, 2016) also indicate that it reached its maximum  
1036 extent several thousand years after the GLGM. Glaciers in two mountain valleys in the  
1037 southern North Cascades retreated from moraines closely nested inside the GLGM  
1038 moraines. However, <sup>36</sup>Cl ages on the Domerie II (17.9-14.7 ka) and the Leavenworth II  
1039 moraines (17.2-15.0 ka) have large uncertainties, and the moraine ages may or may not  
1040 be associated with HS-1 (Porter, 1976; Kaufman et al., 2004; Porter and Swanson, 2008).

1041 The climate in the North Cascades during HS-1 is not well understood due to a lack of  
1042 age control on landforms, limited paleoecological data, and the large influence of the  
1043 continental ice sheets on climate. However, glacial ELAs associated with potential HS-1  
1044 moraines located well to the south of the CIS terminus were slightly above the GLGM  
1045 maximum (Porter, 1976; Kaufman et al., 2004; Porter and Swanson, 2008). In areas  
1046 inundated by the CIS to the north, alpine glaciers retreated to valley heads, presumably  
1047 due to lower precipitation as the continental ice sheets expanded to cover most of Canada  
1048 and northern Washington. Climate models and pollen data indicate that at 16 ka mean  
1049 annual air temperature was 4-7°C cooler than today (Heusser, 1977; Kutzbach, 1987; Liu  
1050 et al., 2009).

#### 1051 *5.5 Rocky Mountain/Yellowstone region*

1052 In some areas of this region, glacier retreat began toward the end of the GLGM; in other  
1053 areas, glaciers maintained their fronts or re-advanced at ~16.5 ka, although with a great  
1054 degree of local variability, and then immediately retreated. Glaciers in some valleys near  
1055 the margins of the Yellowstone Ice Cap reached their local maximum extent at ~17 ka,  
1056 then rapidly retreated at ca. 15 ka when several external climate forcings coincided

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1057 (Licciardi and Pierce, 2018) (Figs. 11 and 12). Moraines dated to the HS-1 period are  
1058 common in valleys along the eastern slope of the Teton Range (Licciardi and Pierce,  
1059 2018) and in the Wind River Range (Dahms et al., 2018, 2019; Marcott et al., 2019). In  
1060 the Wind River Range, these moraines are ~1-2 km downvalley from cirque headwalls in  
1061 14 valleys (Dahms et al., 2010). Ages from these moraines in Stough Basin, Cirque of the  
1062 Towers and Temple Lake cluster around ~15.5 ka (Fig. 10) (Dahms et al., 2018; Marcott  
1063 et al., 2019). Subsequently, a second period of regional deglaciation was well under way  
1064 after ~15 ka (Larsen et al., 2016; Dahms et al., 2018; Pierce et al., 2018).

1065 Glaciers in some valleys in the Colorado Rocky Mountains receded during HS-1. In  
1066 contrast, many other valleys contain end moraines dating to 17-16 ka. Ages on polished  
1067 bedrock surfaces up-valley of these moraines have yielded ages that show that the glaciers  
1068 retreated shortly thereafter (Young et al., 2011; Shakun et al., 2015a; Leonard et al.,  
1069 2017a, 2017b; Laabs et al., 2020, submitted). Ward et al. (2009) suggest that there was a  
1070 stillstand or possible re-advance around 17-15 ka in the Colorado Front Range,  
1071 interrupting overall post-GLGM recession.

## 1072 *5.6 Sierra Nevada*

1073 There is strong evidence for an advance of glaciers in the Sierra Nevada during HS-1 –  
1074 the Tioga 4 advance in local terminology (Phillips et al., 1996) – but HS-1 was not a time  
1075 of extensive glaciation. As described in Section 3.5, retreat from the GLGM maximum  
1076 began gradually at about 19 ka. It accelerated rapidly after 18 ka, and glaciers receded  
1077 past Tioga 4 glacier margins by about 17 ka (Phillips, 2017). Retreat then reversed and  
1078 glaciers readvanced to Tioga 4 positions by 16.2 ka (Fig. 14). The ELA depression for  
1079 this advance was about 900 m, compared to the GLGM ELA depression of about 1200  
1080 m. The Tioga 4 advance apparently was short-lived; by 15.5 ka, the range was effectively  
1081 deglaciated. It is clear from the simultaneous expansion of Lake Lahontan and the Tioga  
1082 4 glaciers that increased precipitation played a major role in glacier expansion at this time.

1083 The fact that Lake Lahontan was relatively small during Tioga 3 (21-19 ka, the LLGM),  
1084 while glaciers were more extensive, shows that Tioga 3 was colder and drier than Tioga  
1085 4. Plummer (2002) estimated that Tioga 4 precipitation was 160% greater than today and  
1086 temperature was 3°C cooler based on the inferred size of Searles Lake at that time. Had  
1087 he used the extent of Lake Lahontan in his analysis, the increase in precipitation would  
1088 have been even larger. Phillips (2017) suggested that the large extent of sea ice in the  
1089 North Atlantic during HS-1 led to greatly increased precipitation and cooler temperatures

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1090 in California through an atmospheric teleconnection. An impediment to further analysis  
1091 of these topics is the chronological inconsistencies between the dating of the Sierra glacial  
1092 record, nearby marine cores, and lacustrine records (Phillips, 2017). More confidence in  
1093 the chronology could allow researchers to resolve questions of climate leads and lags, and  
1094 determine whether the apparent differences in timing are the result of chronological  
1095 imprecision or latitudinal paleoclimate gradients.

### 1096 *5.7 Mexico and Central America*

1097 Glaciers in central Mexico remained at or near their maximum positions throughout HS-  
1098 1. In the interior mountains (e.g. Iztaccíhuatl), glaciers were slightly smaller during HS-  
1099 1 (ELA = 4040 m asl) than at the LLGM (ELA = 3940 m asl, from 21 ka to 17 ka)  
1100 (Vázquez-Selem and Lachniet, 2017). Recession at this time is not recorded in mountains  
1101 near the Pacific Ocean, where a low ELA persisted until 15-14 ka. Indeed, during HS-1,  
1102 ELAs were ca. 400-650 m lower on mountains near the coast than in the interior, which  
1103 suggests a strong precipitation gradient from the coast to the interior and overall drier  
1104 conditions in the interior during HS-1 (Lachniet et al., 2013). In general, the end of HS-  
1105 1 is coeval with the onset of glacier recession ~14.5 ka in central Mexico. Existing  
1106 evidence at Cerro Chirripó, Costa Rica, indicates moraine formation between 18.5 ka and  
1107 17 ka (Cunningham et al., 2019), potentially during the earlier part of HS-1. If the summit  
1108 area was ice-free by 15.2 ka, as suggested by Cunningham et al. (2019), glacier recession  
1109 prevailed during the second part of HS-1 (as defined by Hodell et al., 2017).

### 1110 *5.8 Northern Andes*

1111 Most of the glacier advances in the northern tropical Andes were dated between the end  
1112 of the GLGM and the end of HS-1 (~15 ka). In the Sierra Nevada of the Venezuelan  
1113 Andes, some valleys were completely deglaciated by ~16.5 ka (Angel et al., 2016). In  
1114 others, glaciers advanced ~17 ka (modified ages of Angel, 2016). In the Sierra Santo  
1115 Domingo, maximum advances are dated to ~17.5 ka (modified ages from Wesnousky et  
1116 al., 2012; Angel, 2016). In the Sierra del Norte they date to between 18 ka and 15.5 ka  
1117 (modified ages from Wesnousky et al., 2012; Angel, 2016), and in the Cordillera de  
1118 Trujillo, to around 18 ka (<sup>10</sup>Be ages modified ages from those of Angel, 2016). Some  
1119 advances in the Colombian Andes may be related to HS-1. This is the case in the Bogotá  
1120 Plain, where a moraine complex has been dated to between 18 ka and 14.5 ka (Helmens,  
1121 1988; Helmens et al., 1997b), and in the Central Cordillera, where peat overlying a  
1122 moraine complex yielded a minimum age of 16-15 ka (Thouret et al., 1996). There are

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1123 moraines in the Ecuadorian Andes that are related to HS-1, for example in Cajas National  
1124 Park, vicinity of Pallcacocha lake above 3700 m asl, where a moraine was radiocarbon  
1125 dated to 17-14.5 cal ka BP (Hansen et al., 2003).

1126 Most glacier advances in the northern tropical Andes have been dated to ~18-15 ka based  
1127 on  $^{10}\text{Be}$  ages. However, the scarcity of paleoclimatic information limits our ability to  
1128 estimate the regional HS-1 climate and to compare it to GLGM conditions. Rull (1998)  
1129 proposed a cold event ( $\sim 7^\circ\text{C}$  cooler than today), locally called as El Caballo Stadial, at  
1130 16.5 ka based on a palynological record from the central Mérida Andes in Venezuela.  
1131 Similarly, Hooghiemstra et al. (1993) proposed the Fúquene Stadial at a similar time in  
1132 the Colombian Andes based on a palynological study in the Bogotá Plain. In contrast,  
1133 Brunschön and Behling (2009) concluded that both temperature and precipitation in the  
1134 southern Ecuadorian Andes were higher during the period 16.2-14.7 cal yr BP than during  
1135 the GLGM.

### 1136 *5.9 Peru and Bolivia*

1137 Three moraines near Lake Junín have cosmogenic ages of ~21 ka to 18 ka (Smith et al.,  
1138 2005), providing evidence of an advance prior to HS-1. In contrast, the Galeno moraines  
1139 in the Cajamarca region have slightly younger ages and have complete inset  
1140 lateral/terminal loops with an average age of 19 ka. The Juellesh and Tuco valleys in the  
1141 Cordillera Blanca have inner and outer moraine loops that date, respectively, to  $\sim 18.8 \pm 2.0$   
1142 ka and  $\sim 18.7 \pm 1.6$  ka (Smith and Rodbell, 2010). Glasser et al. (2009) presented similar  
1143 ages ( $\sim 18.3 \pm 1.4$  ka) for an outer lateral moraine in the Tuco valley. An inner lateral  
1144 moraine (M4 of Smith and Rodbell, 2010) has been dated to  $\sim 18.8 \pm 2.3$  ka, and Glasser  
1145 et al. (2009) reported similar ages on the same moraine ( $\sim 17.9 \pm 0.9$  ka). Revised ages on  
1146 various stages of deglaciation of the Cordillera Huayhuash are centered on  $\sim 17.8$ -16.5 ka  
1147 (Hall et al., 2009). Similarly, dated boulders on the Huara Loma, Coropuna, and Wara  
1148 Wara moraines in Bolivia may record post-GLGM advances between 19.4 ka and 18.2  
1149 ka (Zech et al., 2010; May et al., 2011; Martin et al., 2018).

1150 Many valleys in central Peru and Bolivia contain evidence of glacier advances or  
1151 persistent stillstands during HS-1 ( $\sim 17.5$ -14.6 ka) (syntheses in Mark et al., 2017, and  
1152 Martin et al., 2018). The mean exposure ages of all groups of moraine boulders in this  
1153 region that fall within HS-1 is  $16.1 \pm 1.1$  ka. A stillstand synchronous with HS-1 is also  
1154 indicated by cosmogenic  $^3\text{He}$  ages of moraines on the Coropuna volcano, southern Peru  
1155 (Bromley et al., 2009). Radiocarbon and cosmogenic ages from the Cordillera Vilcanota

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1156 and the HualcaHualca volcano (Fig. 17) provide independent evidence that glaciers in  
1157 southern Peru advanced sometime after ~18.0-16.8 ka (Mercer and Palacios, 1977;  
1158 Alcalá-Reygosa et al., 2017), and radiocarbon ages from the Altiplano indicate an  
1159 advance occurred there from ~17 ka to 15.4 ka (Clapperton et al., 1997b; Clapperton,  
1160 1998).

1161 Ice core records from Huascarán, Peru, suggest that HS-1 was the coldest period of the  
1162 past ~19 ka (Thompson et al., 1995), but researchers have argued recently that the  $\delta^{18}\text{O}$   
1163 signal in tropical ice does not provide a pure temperature signal (Quesada et al., 2015).  
1164 The cooling inferred from reconstructions of paleo-ELAs during HS-1 is around 3°C in  
1165 the central Altiplano (Martin et al., 2018).

1166 The northern equatorial Andes of Peru appear to have been wetter during most of HS-1  
1167 (Mollier-Vogel et al., 2013), whereas speleothem records in central Peru suggest that the  
1168 local climate became abruptly drier at ~16 ka (Kanner et al., 2012; Mollier-Vogel et al.,  
1169 2013). Lake-level fluctuations provide strong evidence for pronounced shifts in  
1170 precipitation across the central Andes during this period (Baker et al., 2001a, 2001b;  
1171 Placzek et al., 2006; Blard et al., 2011). Farther south, over the Altiplano, shoreline  
1172 reconstructions demonstrate that the first part of HS-1 (~18-16.5 ka) was similar to or  
1173 drier than today. However, during the Lake Tauca highstand in the second part of HS-1  
1174 (16.5-14.5 ka) precipitation was ca. 130% higher than today (Placzek et al., 2013; Martin  
1175 et al., 2018). Some of the GLGM and older moraines in this part of the Altiplano may  
1176 have been overridden during this wet phase. Martin et al. (2018) established that the  
1177 downward shift in ELA at this time was amplified in valleys that are near the latitudinal  
1178 center of paleo-lake Tauca, resulting from a significant local increase in precipitation.

### 1179 *5.10 Southern Bolivia and Northern Chile*

1180 During HS-1, there was a sharp spatial gradient in climate between Cerro Tunupa, which  
1181 is located at the geographic center of Lake Tauca, and Cerro Uturuncu (Bolivia) and  
1182 elsewhere north of the Arid Diagonal (Ward et al., 2017; Martin et al., 2018). Blard et al.  
1183 (2014) describe a 900 m gradient in ELAs between Cerro Tunupa and Cerro Uturuncu  
1184 based on the Tauca-phase moraines at each site. The spatial gradient in temperature  
1185 between these sites (Ammann et al., 2001) is not sufficient to explain the ELA difference,  
1186 which implies the existence of a strong spatial gradient in precipitation across the  
1187 southern margin of Lake Tauca. Further work by Martin et al. (2018) quantified this  
1188 precipitation gradient, confirming that it was significantly drier in the southern portion of

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1189 the Lake Tauca basin. The presence of this drying trend to the south and west is supported  
1190 by the lack of a clear Tauca-phase transgression at Pozuelos Basin in the Puna region,  
1191 which is at a similar latitude to Cerro Uturuncu and El Tatio (McGlue et al., 2013). Based  
1192 on the clustering of  $^{10}\text{Be}$  and  $^{36}\text{Cl}$  exposure ages on LLGM moraines (Section 4.10),  
1193 Tauca-phase moraines appear to be either absent or restricted to higher parts of valleys at  
1194 El Tatio, Cerro La Torta, and Chajnantor Plateau (Ward et al., 2017), as well as at several  
1195 sites on the central Puna Plateau (Luna et al., 2018) and the western slope of Nevado  
1196 Chañi (24°S) in Argentina (Martini et al., 2017a). The precipitation gradient is consistent  
1197 with paleo-vegetation proxy records that indicate an approximate doubling of modern  
1198 precipitation, from ~300 to ~600 mm/yr (Grosjean et al., 2001; Maldonado et al., 2005;  
1199 Gayo et al., 2012), in the northern Arid Diagonal and adjacent Andes during the Tauca  
1200 highstand. South of the Arid Diagonal, at Valle de Encierro and Cordón de la Rosa, ages  
1201 of 17 ka from highly recessed locations indicate a stillstand or minor advance during HS-  
1202 1, followed by full deglaciation (Ward et al., 2017).

### 1203 *5.11. Central Andes of Argentina*

1204 Initial deglaciation in the Central Andes after the LLGM was followed by renewed glacier  
1205 expansion during HS-1. Moraines that mark the HS-1 limit are found up-valley of those  
1206 constructed during the GLGM. North of the Arid Diagonal, glacier expansion during HS-  
1207 1 coincided with the Tauca paleo-lake (Blard et al., 2011; Placzek et al., 2013). Glaciers  
1208 advanced in the Laguna Grande valley in the Tres Lagunas area between ~17 ka and ~15  
1209 ka (Zech et al., 2017), the east and west sides of Nevado de Chañi ~15 ka (Fig. 19)  
1210 (Martini et al., 2017a), and in the Sierra de Quilmes, between ~18 ka and ~15 ka (Zech  
1211 et al., 2017). An exception to these findings comes from Sierra de Aconquija where  
1212 renewed glacier growth appears to have occurred after the HS-1 stadial (D'Arcy et al.,  
1213 2019). South of the Arid Diagonal, there is almost no evidence of glacial limits dating to  
1214 HS-1. Just one sample from the La Angostura I moraine in the Cordon del Plata has been  
1215 dated to ~15 ka (Moreiras et al., 2017). Moraines up-valley of the GLGM limit in the Las  
1216 Leñas valley and Ansilta Range have not yet been dated (Terrizzano et al., 2017; Zech et  
1217 al., 2017).

### 1218 *5.12 Patagonia*

1219 At the time of the HS-1 stadial, the Patagonian region was experiencing widespread  
1220 warming and deglaciation (Moreno et al., 2015; Bertrand et al., 2008). Rapid warming  
1221 began at 17.8 ka in northwestern Patagonia and approached average interglacial

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1222 temperatures by 16.8 ka (Moreno et al., 2015). Glaciers in northwestern Patagonia  
1223 retreated out of the lowlands shortly before ~17.8 ka and into high mountain cirques  
1224 above 800 m asl by 16.7 ka (Denton et al., 1999a; Moreno et al., 2015). The abrupt and  
1225 synchronous withdrawal of many glacier lobes in northwestern Patagonia was  
1226 contemporaneous with the rapid expansion of temperate rainforests (Heusser et al., 1999;  
1227 Moreno et al., 1999), suggesting pronounced warming at 17.8 ka coupled with a poleward  
1228 shift of the southern westerlies between 17.8 ka and 16.8 ka (Pesce and Moreno, 2014;  
1229 Moreno et al., 2018). However, on the east flank of the Andes (Cisnes valley, 44°S), it  
1230 has been suggested that glaciers started retreating somewhat earlier, at ~19 ka. At this  
1231 site, it has been estimated that the ice had diminished to 40% of its local maximum extent  
1232 by ~16.9 ka (Weller et al., 2017; Garcia et al., 2019).

1233 Farther south, in central Patagonia, lake cores from two small basin (Villa-Martínez et  
1234 al., 2012; Henríquez et al., 2017) show that the Lago Cochrane/Pueyrredón ice lobe  
1235 (47.5°S) retreated over 90 km into the Chacabuco Valley between ~21 ka (Río Blanco  
1236 moraines; Hein et al., 2010a) and 19.4 ka. Ice receded an additional ~60 km to reach a  
1237 position close to modern glacier limits by around 16-15 ka (Turner et al., 2005; Hein et  
1238 al., 2010; Boex et al., 2013; Mendelova et al., 2017; Davies et al., 2018; Thorndycraft et  
1239 al., 2019). Retreat east of the shrinking ice sheet in the Lago Cochrane sector of central  
1240 Patagonia occurred without discernable warming (Henríquez et al., 2017). Almost  
1241 certainly, however, this retreat was facilitated by calving in deep proglacial lakes that  
1242 formed in the over-deepened Cochrane/Pueyrredón and General Carrera/Buenos Aires  
1243 basins as the glaciers withdrew (Turner et al., 2005; Bell, 2008; Hein et al., 2010; Borgois  
1244 et al., 2016; Glasser et al., 2016; Davies et al., 2018; Thorndycraft et al., 2019). At Lago  
1245 General Carrera/Buenos Aires (46.5°S), glacier retreat from the Fenix I moraine  
1246 commenced ~18 ka, but was interrupted by a readvance to the Menucos moraines at ~17.7  
1247 ka. An annually resolved lake sediment record, tied to a calendar-year timescale by the  
1248 presence of the well dated Ho tephra erupted from Volcán Hudson (17,378±118 cal yr  
1249 BP), indicates that ice remained close to the east end of the lake until after 16.9 ka, before  
1250 retreating back into the mountains (Kaplan et al., 2004; Douglass et al., 2006; Bendle et  
1251 al., 2017, 2019). Bendle et al. (2019) suggest that the onset of deglaciation in central  
1252 Patagonia was a direct result of the HS-1 event. They hypothesize that warming at the  
1253 start of HS-1 occurred due to rapid poleward migration of southern westerly winds, which  
1254 increased solar radiation and ablation at the ice sheet surface. They linked warming and



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1255 accelerated deglaciation to the oceanic bipolar seesaw, which delayed Southern  
1256 Hemisphere warming following the slowdown of the Atlantic meridional overturning at  
1257 the start of HS-1 (Bendle et al., 2019).

1258 Determining whether “early LGM” and “early deglaciation” are correct interpretations of  
1259 glacier activity in central Patagonia (44°-49°S) (Van Daele et al., 2016; García et al.,  
1260 2019) is important for determining whether local (glaciological, reworking of old organic  
1261 matter) or regional (climatic) mechanisms are responsible for apparent differences in  
1262 timing, rate, and magnitude of glacier fluctuations prior to and during the LLGM and  
1263 Termination I (Vilanova et al., 2019). Another problem emerges from studies of lake  
1264 sediments from the eastern slopes of the Andes in central Patagonia. Based on an analysis  
1265 of seismic data and lake sediment cores from Lago Castor (Fig. 1), Van Daele et al. (2016)  
1266 concluded that the Coyhaique glacier lobe achieved its maximum extent and retreated  
1267 before the LLGM. The concepts of ‘early LGM’ and ‘early deglaciation’ rely heavily on  
1268 the interpretation and selective rejection of anomalously old radiocarbon ages, which  
1269 include results as old as  $43,100 \pm 3600$   $^{14}\text{C}$  yr BP in the clastic-dominated and intensely  
1270 reworked portion of the Lago Castor cores beneath the H0 tephra, which has been  
1271 radiocarbon dated to 17,300 cal yr BP (Weller et al., 2014). This enigmatic radiocarbon  
1272 chronology has not been corroborated by more recent studies in the Río Pollux valley,  
1273 where Moreno et al. (2019) and Vilanova et al. (2019) have reported stratigraphic,  
1274 geochronologic, and palynological results from small, closed-basin lakes to constrain the  
1275 timing and extent of the Coyhaique glacier lobe during Termination I. These studies point  
1276 to the abandonment of the final LLGM margins at ~17.9 ka, ~600 years before the  
1277 reported age of the H0 tephra. The similarities between northern and southern Patagonia  
1278 (see below), and contrasts with the Río Cisnes and Lago Cochrane/Pueyrredón glacier  
1279 lobes, suggest that the different behavior of the latter might arise from differences in their  
1280 topographic setting, ice divide migration (Mendelova et al., 2019), or differential calving  
1281 in large proglacial lakes in the Central Patagonian Andes during the final stage of the  
1282 LLGM.

1283 In southern Patagonia, the Lago Argentino lobe (50°S) retreated at least 60 km from its  
1284 LLGM by 16.2 ka (Strelin et al., 2011). A nearby mountain glacier at Río Guanaco (50°S)  
1285 retreated to half its extent between 18.9 ka and 17 ka, suggesting a temperature increase  
1286 of ~1.5°C, or about one-third of the total deglacial warming relative to today (Murray et

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1287 [al., 2012](#)). Similarly, the Última Esperanza ice lobe retreated after 17.5 ka, but with a  
1288 short period of stabilization at ~16.9-16.2 ka ([Sagredo et al., 2011](#)).

### 1289 *5.13 Tierra del Fuego*

1290 HS-1 in the Cordillera Darwin was characterized by very rapid glacier recession with no  
1291 evidence of stillstands ([Hall et al., 2013, 2017a](#)). Surface exposure ages on boulders  
1292 indicate that ice was at the innermost GLGM moraine at the shore of Bahía Inútil at ~18  
1293 ka ([McCulloch et al., 2005b; Kaplan et al., 2008; Hall et al., 2013](#)), but retreated shortly  
1294 thereafter ([McCulloch et al., 2005b](#)). Radiocarbon ages from peat bogs near present-day  
1295 sea level indicate that the Cordillera Darwin icefield had retreated inside fjords by 16.8  
1296 ka ([Hall et al., 2013, 2017b](#)). On the north side of the Cordillera Darwin, this recession  
1297 was ~130 km from its LLGL. In the Fuegian Andes, two <sup>10</sup>Be ages from glacially eroded  
1298 bedrock in front of an alpine glacier indicate that recession was well underway by ~17.8  
1299 ka and had reached the late-glacial position as early as ~16.7 ka ([Menounos et al., 2013](#)).  
1300 Whether this glacier was part of the Cordillera Darwin icefield or a separate ice mass at  
1301 the GLGM remains uncertain ([Coronato, 1995; Menounos et al., 2013](#)). In any case,  
1302 glaciers in the region responded to HS-1 by rapidly retreating, as was the case at some  
1303 other Southern Hemisphere locations ([Putnam et al., 2013](#)).

### 1304 *5.14 Synthesis*

1305 Glaciers in most of North and Central America began to retreat from their GLGM  
1306 positions by about 21 ka (Table 2 and Fig. 4). In some areas (e.g. Wind River Range),  
1307 they suffered the same mass losses after ~21 ka as other glaciers, but apparently re-  
1308 advanced during HS-1. In other regions (e.g. Yellowstone Ice Cap, the Colorado Rocky  
1309 Mountains and on some Mexican volcanoes), glaciers reached their maximum extents  
1310 during HS-1. Some of these glaciers may have advanced from the GLGM to HS-1 and  
1311 surpassed their GLGM limits. This possibility, however, must be considered hypothetical,  
1312 as it is inherently difficult to verify.

1313 Interestingly, one of the Northern Hemisphere regions that appears to have been least  
1314 affected by the HS-1 event, at least in terms of the ice-marginal fluctuations, is the LIS.  
1315 Rather, the ice sheet thinned and retreated during this period. It is likely that internal flow  
1316 patterns and ice divides were impacted by drawdown induced by the Hudson Strait ice  
1317 stream. There are few data from Alaska to evaluate the effects of HS-1 on glaciers, but  
1318 there is some evidence of advances interrupting overall retreat during this interval. The

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1319 southern sector of the CIS and a number of glaciers in Colorado and those proximal to  
1320 the Yellowstone Ice Cap area reached their maximum extents during HS-1. In a few  
1321 valleys in the North Cascades south of the CIS limit, possible HS-1 moraines lie upvalley  
1322 of GLGM moraines, although data are sparse. A clear advance immediately following  
1323 HS-1 has been documented in the Sierra Nevada and the Wind River Range. In the Sierra,  
1324 HS-1 moraines, locally termed Tioga 4, lie well inside GLGM moraines. These moraines  
1325 record an ELA depression of 900 m, which is 300 m less than during the GLGM. In the  
1326 Wind River Range, the Older Dryas/HS-1 moraines lie 19-27 km upvalley of  
1327 LLGM/GLGM moraines. In the interior mountains of Central Mexico and Costa Rica,  
1328 moraines dating to near HS-1 lie inside GLGM moraines. However, glaciers in mountains  
1329 close to the oceans remained at, or advanced past, their GLGM limits until the end of HS-  
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1331 In the Sierra Nevada, temperatures were 3°C lower than today during HS-1, but clearly  
1332 precipitation was increased. In other regions, data appear to confirm the decrease in  
1333 temperature in the Sierra Nevada, but there is little information on precipitation.

1334 Glaciers in the tropical Andes built significant moraine complexes during HS-1, attesting  
1335 to a significant stillstand or readvance. In the northern Andes, numerous moraines have  
1336 been dated to this period, reflecting an interruption of the longer-term of trend glacier  
1337 retreat. HS-1 advances are widespread and significant in central and southern Peru and in  
1338 Bolivia. Although the first part of the HS-1 stadial in these areas was dry, the second part  
1339 was wet, with, on average, a two-fold increase in precipitation above modern values. The  
1340 precipitation increase may have been five-fold around the Altiplano paleo-lakes (Tauca  
1341 highstand from 16.5 ka to 14.5 ka). This precipitation control on glacier mass balance is  
1342 a strong driver of the spatial variability of ELA reductions during HS-1. Several of the  
1343 HS-1 moraines in the region appear to have been constructed by glaciers that were very  
1344 close to LLGM moraines. HS-1 moraines are also present in the Arid Diagonal, although  
1345 aridity increased towards the south, resulting in a more limited glacier extent in that area.  
1346 In some cases, glaciers in the Arid Diagonal disappeared during HS-1. Glaciers advanced  
1347 during HS-1 in the Central Andes of Argentina after a long period of retreat, and at the  
1348 same time as the Tauca highstand.

1349 In contrast, glaciers in the temperate and subpolar Andes abandoned their LGM positions  
1350 and underwent sustained or step-wise recession during HS-1. In northwestern Patagonia,  
1351 climate warmed rapidly and experienced a significant decline in precipitation, driven by

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1352 a southward shift of the southern westerly winds (Pesce and Moreno, 2014; Moreno et  
1353 al., 2015, 2018; Henríquez et al., 2017; Vilanova et al., 2019). The magnitude of these  
1354 changes appears to decline south of 45°S, modulated by the regional cooling effect of  
1355 residual ice masses in sectors adjacent to the eastern margins of the Patagonian ice sheet  
1356 (Henríquez et al., 2017). The difference in glacier behavior between the tropical Andes  
1357 and Patagonia and Tierra del Fuego during HS-1 could be due to two causes. First, the  
1358 significant increase in precipitation in the tropical Andes during HS-1 could be the main  
1359 cause of the glacier advances in that region. Second, Patagonia and Tierra del Fuego may  
1360 have been too distant from the events responsible for HS-1, which are closely related to  
1361 North Atlantic circulation; rather they may have been more affected by Antarctica and  
1362 southern westerly winds. The two effects may have even converged, dividing the  
1363 continent into two different glacial regimes during HS-1 (Sugden et al., 2005).

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## 1365 **6. Evolution of American Glaciers during the Bølling-Allerød Interstadial (B-A) and** 1366 **the Antarctic Cold Reversal (ACR) (14.6-12.9 ka)**

### 1367 *6.1 Bølling-Allerød Interstadial and the Antarctic Cold Reversal*

1368 The term 'Bølling-Allerød' (B-A) is derived from recognition of two warm Late Glacial  
1369 palynological zones (the Bølling and the Allerød) between the HS-1 and Younger Dryas.  
1370 The use of this term for a chronological period has been criticized from a palynological  
1371 point of view (De Klerk, 2004). Nevertheless, warming during this period has been  
1372 identified (Lowe et al., 2001) and firmly dated in the GI-1 Greenland ice core to 14.6 ka  
1373 to 12.9 ka (Rasmussen et al., 2014), and the term Bølling-Allerød interstadial  
1374 (abbreviated 'B-A') is customarily applied to this period.

1375 The B-A period began with reinforcement of the AMOC (McManus et al., 2004) and a  
1376 marked increase in atmospheric CO<sub>2</sub> (Chen et al., 2015) and methane (Rosen et al., 2014);  
1377 these conditions persisted through this period (Monnin et al., 2001). Climate rapidly  
1378 warmed, at least around the North Atlantic (Clark et al., 2012). The AMOC remained  
1379 vigorous throughout the B-A period (Deaney et al., 2017), and only a few cold events  
1380 interrupted it in the Northern Hemisphere (Rasmussen et al., 2014). Sea ice retreated to  
1381 the north (Denton et al., 2005), and glaciers in Europe thinned and retreated (for example  
1382 in the Alps; Ivy-Ochs, 2015). The Asian monsoon strengthened to a level similar to the  
1383 present (Sinha et al., 2005; Wang et al., 2008). It seems that the changes in the oceans

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1384 preceded changes in the atmosphere, and the oceans had a decisive influence on Northern  
1385 Hemisphere warming (Thiagarajan et al., 2014). The changes in the oceans were possibly  
1386 caused by a period of intense melt in Antarctica just before the B-A (Weaver et al., 2003;  
1387 Weber et al., 2014). The process that drove the B-A would then be the opposite of that  
1388 which caused HS-1, when the melting of the northern ice sheets led to warming in the  
1389 Southern Hemisphere (Zhang et al., 2016). During the B-A, cooling in Antarctica caused  
1390 increased sea ice cover in the surrounding ocean, causing the southern westerlies and the  
1391 Intertropical Convergence Zone (ITCZ) to migrate northward, and strengthening the  
1392 AMOC, which in turn caused warming in the Northern Hemisphere (Pedro et al., 2015;  
1393 Zhang et al., 2016).

1394 The cold period in the south has been called the Antarctic Cold Reversal (ACR). We  
1395 analyze the B-A and ACR together because they occurred around the same time, although  
1396 the boundary between cooling in the south and the warming in the north is not well  
1397 defined (Pedro et al., 2015). The ACR has been well documented in Antarctic ice cores,  
1398 and a clear bipolar seesaw is observed in relation to Greenland ice cores (Blunier et al.,  
1399 1997, 1998; Pedro et al., 2011). Cooling in the Southern Hemisphere is apparent up to  
1400 40° S (Pedro et al., 2015), resulting in widespread glacier advance (Putnam et al., 2010;  
1401 Shulmeister et al., 2019). There is also a clear cooling signal in tropical areas, at least in  
1402 high Andean regions (Jomelli et al., 2014, 2016).

## 1403 *6.2 Laurentide Ice Sheet*

1404 The hemispheric extent of glaciation during the B-A is summarized in Figure 5. The  
1405 Bølling-Allerød interstadial is characterized by enhanced ablation in marginal areas of  
1406 the LIS (Ullman et al., 2015b) and a marked acceleration in the rate of retreat, most  
1407 notably along the southern and western margins, but with minimal retreat along its  
1408 northern margin (Dyke and Prest, 1987; Dyke, 2004; Stokes, 2017). As a result, the LIS  
1409 is likely to have fully separated from the CIS by the end of the interstadial, although  
1410 precise dating of the opening of the ‘ice-free corridor’ remains a challenge (Dyke and  
1411 Prest, 1987; Gowan, 2013; Dixon, 2015; Pedersen et al., 2016). It is worth noting,  
1412 however, that positive feedback mechanisms related to ice surface lowering and surface  
1413 mass balance are likely to have resulted in the rapid ‘collapse’ of the saddle between the  
1414 LIS and the CIS, which some have hypothesized was the source of Meltwater Pulse 1A  
1415 (Gregoire et al., 2012).

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1416 The rapid retreat of the southern and western margins of the LIS was also likely aided by  
1417 the development of proglacial lakes that facilitated calving and the draw-down of ice,  
1418 particularly at the southern margin (Andrews, 1973; Dyke and Prest, 1987; Cutler et al.,  
1419 2001). Moreover, the rapid retreat of the LIS during this time period led to major changes  
1420 in the trajectory of ice streams at the western and southern margins, with associated  
1421 changes in the location of the major ice divide in Keewatin, which migrated several  
1422 hundred kilometers east towards Hudson Bay (Dyke and Prest, 1987; Margold et al.,  
1423 2018).

1424 There is also clear evidence for an overall acceleration in the rate of retreat and thinning  
1425 of the ice sheet in the southeastern sector. This has been characterized as a two-phase  
1426 pattern of deglaciation (Barth et al., 2019), with steady retreat starting ~20 ka and then  
1427 increasing around 14.5 ka, coincident with the B-A warming. A clear example of this is  
1428 seen in an extensive suite of 21 <sup>36</sup>Cl ages from boulder and bedrock samples along vertical  
1429 transects spanning ~1000 m of relief in the Adirondack Mountains of the northeastern  
1430 USA (Barth et al., 2019). These data suggest gradual ice sheet thinning of 200 m initiated  
1431 around 20 ka, followed by a rapid surface lowering of 1000 m, coincident with the onset  
1432 of the B-A warming (Barth et al., 2019). Similarly high rates of thinning are also recorded  
1433 on Mt. Mansfield, Vermont's highest peak, although they appear to have initiated around  
1434 13.9±0.6 ka, which slightly post-dates the abrupt onset of the B-A (Corbett et al., 2019).

1435 Despite an acceleration in the overall rate of recession, there appears to have been  
1436 minimal recession of the LIS along its northern margin (Dyke, 2004). Also, there is  
1437 evidence for readvances/oscillations of some of the lobes in the vicinity of the Great  
1438 Lakes (Dyke, 2004), perhaps related to internal 'surge' dynamics and short-lived ice  
1439 stream activity, rather than any external climatic forcing (Clayton et al., 1985; Patterson,  
1440 1997; Cutler et al., 2001; Margold et al., 2015, 2018; Stokes et al., 2016). There is also  
1441 some evidence of climatically induced readvances of parts of the LIS during the B-A. For  
1442 example, recession of the ice margin in northern New Hampshire was interrupted by the  
1443 Littleton-Bethlehem readvance and deposition of the extensive White Mountain moraine  
1444 system (Thompson et al., 2017). Based on a suite of approaches (glacial stratigraphy and  
1445 sedimentology, radiocarbon dating, varve chronology, and cosmogenic-nuclide exposure  
1446 dating), Thompson et al. (2017) constrained the age of this readvance to ~14.0-13.8 ka,  
1447 coincident with Older Dryas cooling.

### 1448 6.3 Alaska

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1449      Glaciers in the Brooks Range were smaller than today by 15 ka in some valleys and ~14  
1450      ka in others (Badding et al., 2013; Pendleton et al., 2015), suggesting widespread glacier  
1451      retreat around the time of the B-A onset. In southeast Alaska, there was widespread  
1452      glacier collapse throughout fjords and sounds during this period (Baichtal, 2010; Carlson  
1453      and Baichtal, 2015; J. Baichtal, unpublished data). Whether this recession was related to  
1454      an abrupt increase in temperature or to a steady temperature increase during this broader  
1455      time period is unknown. However, rising lake levels and decreasing aridity at ~15 ka  
1456      (Abbott et al., 2000; Finkenbinder et al., 2014; Dorfman et al., 2015) suggest that there  
1457      was a major climate shift in Alaska at this time.

#### 1458      6.4 Cordilleran Ice Sheet and North Cascades

1459      The B-A interstadial began with the rapid disintegration of the CIS and deglaciation in  
1460      the North Cascades from 14.5 ka to 13.5 ka (Clague, 2017; Menounos et al., 2017; Riedel,  
1461      2017). Recent glacio-isostatic adjustment models supported by data calibration from  
1462      records of sea level, paleo-lake shorelines, and present-day geodetic measurements  
1463      confirm more than 500 m of thinning of the CIS between 14.5 ka and 14.0 ka (Peltier et  
1464      al., 2015; Lambeck et al., 2017). The pattern of CIS deglaciation was complex due to the  
1465      influences of mountain topography, marine waters and regional climate variability. Early  
1466      deglaciation was marked by rapid eastward frontal retreat across the British Columbia  
1467      continental shelf and northward retreat up Puget Sound. Rapid down-wasting exposed  
1468      high-elevation hydrologic divides and led to the isolation of large ice masses in mountain  
1469      valleys (Riedel, 2017). Lakeman et al. (2008) presented evidence that the CIS in north-  
1470      central British Columbia thinned and in some areas transformed into a labyrinth of dead  
1471      or dying ice tongues in valleys. The presence of ice-marginal landforms in most North  
1472      Cascade valleys is likely related to temporary stillstands of the wasting remnants of the  
1473      CIS, but the ages of most of these landforms are unknown (Riedel, 2017).

1474      Ice sheet deglaciation temporarily rearranged regional drainage patterns. Frontal retreat  
1475      of ice back to the north from hydrologic divides led to the formation of proglacial lakes  
1476      in southern British Columbia and northern Washington (Fulton, 1967; Riedel, 2007). The  
1477      lakes generally drained to the south, and several major valleys carried Late Glacial  
1478      outburst floods that crossed low hydrologic divides, connecting rivers and fish migration  
1479      pathways that later became isolated. The Sumas advances of the CIS diverted Chilliwack  
1480      and Nooksack rivers to the south into lower Skagit valley (Clague et al., 1997). Fish  
1481      genetics and geomorphic evidence, including perched deltas and boulder gravel deposits,

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2717 1482 indicate that the lower Fraser River may have been diverted through Skagit valley at this  
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2719 1483 time.  
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2721 1484 CIS deglaciation during the B-A was interrupted by minor advances of the CIS, and some  
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2723 1485 alpine glaciers also advanced. The Sumas I advance of the CIS across Fraser Lowland  
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2725 1486 occurred between 13.6 ka and 13.3 ka (Clague et al., 1997; Kovanen and Easterbrook,  
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2727 1487 2002). Top-down deglaciation of the ice sheet from mountain divides led to exposure of  
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2729 1488 valley heads and cirques before adjacent valley floors. This set the stage for the formation  
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2731 1490 of new cirque and valley moraines from Yukon Territory to the North Cascades during  
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2733 1491 the B-A (Clague, 2017; Riedel, 2017). Menounos et al. (2017) report 76 <sup>10</sup>Be surface  
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2735 1492 exposure ages on bedrock and boulders associated with lateral and end moraines at 26  
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2737 1493 locations in high mountains of British Columbia and Yukon Territory. At some of these  
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2739 1494 sites, they also obtained radiocarbon ages from lakes impounded by moraines or till.  
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2741 1495 Three older moraines have a combined median age of 13.9 ka, which the authors assigned  
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2743 1496 to the B-A. A moraine near Rocky Creek at Mount Baker was built before 13.4-13.3 ka  
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2745 1497 based on the age of volcanic ash and charcoal on the moraine surface. The Hyak I and  
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2747 1498 Rat Creek I moraines have <sup>36</sup>Cl surface exposure ages of 14.6–12.8 ka, but uncertainty in  
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2749 1499 the <sup>36</sup>Cl surface exposure ages precludes a definitive correlation with this event (Weaver  
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2751 1500 et al., 2003).

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2753 1501 There is sparse geological and paleoecological data on climate during the B-A interval  
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2755 1502 from North Cascades and CIS region. In the North Cascades, the tentatively dated Rat  
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2757 1503 Creek and Hyak alpine glacial moraines had ELAs ~500-700 m below those of modern  
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2759 1504 glaciers or about 200 m above the GLGM advances (Porter et al., 1983). The lower ELAs  
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2761 1505 were caused, in part, by mean July temperatures about 4-6°C below modern values  
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2763 1506 (Heusser, 1977; Kutzbach, 1987; Liu et al., 2009). Rapid loss of the CIS was driven by a  
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2765 1507 positive temperature anomaly of 1-2°C early in the B-A, while a regional increase in mean  
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2767 1508 annual precipitation of 250 mm and brief cold periods with temperature reductions of 1.5  
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2769 1509 °C caused the small glacier advances later in the B-A (Liu et al., 2009).

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2771 1509 *6.5 Rocky Mountain/Yellowstone region*  
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2773 1510 Although glaciers in some southwestern valleys continued to advance after 16 ka due to  
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2775 1511 their exposure to greater orographic precipitation, the Yellowstone ice cap experienced  
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2777 1512 intense deglaciation from 15 ka to 14 ka in response to a warming climate (Licciardi and  
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2779 1513 Pierce, 2018; Pierce et al., 2018). Glaciers in the Wind River Range retreated behind their  
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2781 1514 HS-1 moraines at this time, possibly as far as cirque headwalls (Dahms et al., 2018;



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1515 [Marcott et al., 2019](#)) before they began to readvance during the YD (see below).  
1516 Deglaciation occurred in all ranges in the Colorado Rocky Mountains after about 16 ka,  
1517 and by 13 ka most glaciers had disappeared ([Laabs et al., 2009](#); [Young et al., 2011](#);  
1518 [Shakun et al., 2015a](#); [Leonard et al., 2017a, 2017b](#)).

#### 1519 *6.6 Sierra Nevada*

1520 Glaciers in the Sierra Nevada retreated to cirque headwalls by about 15.5 ka, well before  
1521 the start of the B-A ([Phillips, 2016, 2017](#)). This relatively early disappearance is  
1522 attributable to the southerly latitude and summer-warm, high-insolation Mediterranean  
1523 climate of the Sierra Nevada. Following the B-A transition, glaciers reappeared for a very  
1524 short interval prior to the Holocene. This event, named the ‘Recess Peak advance’,  
1525 resulted from an approximate 150 m decrease in the ELA, in comparison to a 1200 m  
1526 decrease during the GLGM maximum advance ([Clark and Gillespie, 1997](#)).  
1527 Unfortunately, the chronological control for the time of this advance is imprecise. Three  
1528 radiocarbon ages from bulk organic matter in lake cores from two different lake basins  
1529 that overlie Recess Peak till fall between 14 ka and 13 ka, suggesting correlation with  
1530 both the Inter-Allerød Cold Period and the ACR ([Bowerman and Clark, 2011](#)). However,  
1531 cosmogenic ages (both  $^{10}\text{Be}$  and  $^{36}\text{Cl}$ ), although somewhat scattered and imprecise, tend  
1532 to cluster in the 12.7-11.3 ka range, which would be correlative with the Younger Dryas.  
1533 More recently, [Marcott et al. \(2019\)](#) averaged six new  $^{10}\text{Be}$  ages to obtain a date of  
1534  $12.4\pm 0.8$  ka for the Recess Peak advance, which is consistent with the previous  
1535 cosmogenic ages but does not definitively establish whether it was a YD or ACR event.  
1536 Most indirect regional indicators of cooling also fall within the Younger Dryas age range.  
1537 [Phillips \(2016\)](#) performed an in-depth study of this issue, but was unable to arrive at any  
1538 definitive conclusion. In summary, there is no unequivocal evidence for any glacier  
1539 presence in the Sierra Nevada during the B-A. It is possible that there was a brief minor  
1540 advance toward the end of the B-A, but the dating of this event has yet to establish this  
1541 with any certainty.

#### 1542 *6.7 Mexico and Central America*

1543 Data from central Mexico, and to some extent Costa Rica, indicate that glaciers receded  
1544 during the B-A, consistent with warming in the American tropics ([Vázquez-Selem and](#)  
1545 [Lachniet, 2017](#)). In central Mexico, slow initial deglaciation from 15 ka to 14 ka was  
1546 accompanied by the formation of small recessional moraines close to those of the  
1547 maximum advance ([Vázquez-Selem and Lachniet, 2017](#)). Subsequently, glacier recession

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1548 accelerated, as evidenced by exposure ages on glacially abraded surfaces from 14 to 13  
1549 ka. The ELA increased by at least 200 m during that period (Vázquez-Selem and  
1550 Lachniet, 2017). According to Cunningham et al. (2019), Cerro Chirripó, in Costa Rica,  
1551 was ice-free by 15.2 ka, before the onset of the B-A. However, also in Cerro Chirripó,  
1552 Potter et al. (2019) proposed periods of glacier retreat and stillstand from 15 ka to 10 ka.

### 1553 *6.8 Northern Andes*

1554 An advance of Ritacuba Negro Glacier in the Sierra Nevada de Cocuy, Colombia, has  
1555 been linked to the ACR and an ELA decrease of about 500 m (Jomelli et al., 2014). A  
1556 model simulation of the last deglaciation in Colombia (Liu et al., 2009; He et al., 2013)  
1557 suggests a temperature  $2.9^{\circ}\pm 0.8^{\circ}\text{C}$  lower than today during the ACR, with a 10% increase  
1558 in annual precipitation (Jomelli et al., 2016). Bracketing radiocarbon ages on laminated  
1559 proglacial lake sediments indicate that glaciers retreated in the central Mérida Andes of  
1560 Venezuela under warmer and wetter conditions at the start of the Bølling (14.6 ka) (Rull  
1561 et al., 2010). Glaciers then briefly advanced under colder conditions from 14.1 ka to 13.9  
1562 ka), followed by warm and dry conditions during the Allerød (13.9-12.9 ka) (Stansell et  
1563 al., 2010).

### 1564 *6.9 Peru and Bolivia*

1565 There is evidence for glacier advance at many sites in Peru and Bolivia during the ACR  
1566 (Jomelli et al., 2014). Mean surface exposure ages on moraines built during this advance  
1567 are 14.4-12.7 ka; at some sites there is an apparent bimodal distribution of ages (Jomelli  
1568 et al., 2014). A glacier advance at Nevado Huaguruncho in the Eastern Cordillera of the  
1569 Peruvian Andes has been dated to  $14.1\pm 0.4$  ka, based on both exposure ages on moraines  
1570 and radiocarbon ages on lake sediments, and was followed by retreat by  $13.7\pm 0.4$  ka  
1571 (Stansell et al., 2015). Two sets of moraine ridges in valleys within the Cordillera  
1572 Huayhuash date to the ACR (Hall et al., 2009). However, moraine ages from the  
1573 Queshque valley in the nearby Cordillera Blanca are at the end of the ACR (Stansell et  
1574 al., 2017). In Bolivia, the two moraines from Wara Wara and Tres Lagunas (Zech et al.,  
1575 2009, 2010) may have been constructed during the ACR, but could be older (Jomelli et  
1576 al., 2014). A moraine of Telata Glacier in Zongo Valley formed during either the ACR or  
1577 YD (Jomelli et al., 2014). The ACR advance exceeded all subsequent Holocene advances,  
1578 with an ELA estimated to be 450-550 m below its current level based on glaciological  
1579 modeling (Jomelli et al., 2014, 2016, 2017). Some glacial valleys contain at least two sets  
1580 of moraines attributed to the ACR (Jomelli et al., 2014), suggesting multiple advances

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1581 related to possible centennial-scale climate fluctuations during this period. However, such  
1582 patterns must be better documented in other mountain ranges to establish a robust climate  
1583 interpretation (Figs. 21 and 22).

1584 Paleoclimate records suggest that the central tropical Andes were cold during the ACR  
1585 (Jomelli et al., 2014), although some contradictory evidence exists. Moreover, fluvial  
1586 sediment records suggest that northern Peru was wet at the start of the ACR but  
1587 subsequently became drier (Mollier-Vogel et al., 2013); and speleothem records from  
1588 Brazil suggest that the ACR was a period of drier monsoon conditions (Novello et al.,  
1589 2017). Farther south on the Altiplano, lake sediment records also indicate that the ACR  
1590 was likely a drier interval (Sylvestre et al., 1999; Baker et al., 2001b), as does the  
1591 shoreline stratigraphy, indicating that Lake Tauca had vanished (Placzek et al., 2006;  
1592 Blard et al., 2011).

1593 Climate forcings responsible for such glacier trends during the ACR were analyzed using  
1594 transient simulations with a coupled global climate model (Jomelli et al., 2014). Results  
1595 suggest that glacial behavior in the tropical Andes was mostly driven by temperature  
1596 changes related to the AMOC variability superimposed on a deglacial CO<sub>2</sub> rise. During  
1597 the ACR, temperature fluctuations in the tropical Andes are significantly correlated with  
1598 other Southern Hemisphere regions (Jomelli et al., 2014), in particular with the southern  
1599 high-latitudes and the eastern equatorial Pacific. Cold SSTs in the eastern equatorial  
1600 Pacific were associated with glacier advance.

#### 1601 *6.10 Southern Bolivia and Northern Chile*

1602 There are no glacial landforms in the Arid Diagonal that have been dated with sufficient  
1603 precision to permit an ACR age assignment (Ward et al., 2015). There are, however, small  
1604 undated moraines in the upper headwaters at El Tatio that may date to this period, or  
1605 perhaps to the Younger Dryas (Ward et al., 2017). Sites to the south and west, even those  
1606 north of the Arid Diagonal, appear to have been fully deglaciaded by this time.

#### 1607 *6.11 Central Andes of Argentina*

1608 As of yet, there are no firmly documented glacier advances in the Argentine Andes after  
1609 HS-1. In the Sierra de Aconquija, however, D’Arcy et al. (2019) obtained two ages on a  
1610 moraine (M3a) that fall within the B-A/ACR. At Tres Lagunas, there are no moraines  
1611 younger than HS-1 (Zech et al., 2009). Possible B-A/AC moraines at other locations

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1612 (Sierra de Quilmes, Ansilta Range and Las Leñas) have not yet been dated (Terrizzano et  
1613 al., 2017; Zech et al., 2017).

### 1614 6.12 Patagonia

1615 Many researchers have identified B-A/ACR glacier advances in central and southern  
1616 Patagonia (Turner et al., 2005; Ackert et al., 2008; Kaplan et al., 2008; Moreno et al.,  
1617 2009; Glasser et al., 2011; Sagredo et al., 2011, 2018; Strelin et al., 2011; García et al.,  
1618 2012; Nimick et al., 2016; Davies et al., 2018; Mendelova et al., 2020). Past research on  
1619 glacier fluctuations in northwestern Patagonia did not focus on the last termination,  
1620 consequently no evidence of an advance of ACR age has yet been reported. However,  
1621 paleoecological records from sectors as far north as 41°S suggest cooling during this  
1622 interval (Hajdas et al., 2003). For example, records from northwestern Patagonia (40°-  
1623 44°S) show declines in relatively thermophilous trees and increases in the cold-  
1624 tolerant/hygrophilous conifer *Podocarpus nubigena* during ACR time, suggesting a shift  
1625 to cold/wet conditions (Jara and Moreno, 2014; Pesce and Moreno, 2014; Moreno and  
1626 Videla, 2016; Moreno et al., 2018). There is a gap in well-dated glacial geologic studies  
1627 along a ~600 km length of the Andes between 40°S and 47°S (Fig. 23) covering the time  
1628 span of the ACR. The only existing study reports a glacial advance in the Cisnes valley  
1629 (44°S) sometime between 16.9 and 12.3 ka (Garcia et al. 2019), however, the  
1630 chronological constrains are too broad to reach further conclusions.

1631 Detailed geomorphic studies suggest that glaciers in central and southwestern Patagonia  
1632 experienced repeated expansion or marginal fluctuations during the ACR period (Strelin  
1633 et al., 2011; García et al., 2012; Sagredo et al., 2018; Reynhout et al., 2019; Thorndycraft  
1634 et al., 2019). Multiple <sup>10</sup>Be ages from moraines deposited by glaciers on the Mt. San  
1635 Lorenzo massif (47°S) indicate that glaciers there reached their maximum Late Glacial  
1636 extents at 13.8±0.5 ka (Tranquilo Glacier; Sagredo et al., 2018), 13.2±0.2 ka (Calluqueo  
1637 Glacier; Davies et al., 2018), and 13.1±0.6 ka (Lacteo and Belgrano glaciers; Mendelova  
1638 et al., 2020). An ELA reconstruction based on the data from Tranquilo valley suggests  
1639 that temperatures were 1.6-1.8°C lower than at present at the peak of the ACR (Sagredo  
1640 et al., 2018). García et al. (2012) report a mean age of 14.2±0.6 ka for a sequence of  
1641 moraines farther south, in the Torres del Paine area (51°S). The latter findings support  
1642 the conclusions of Moreno et al. (2009), based on radiocarbon-dated ice-dammed lake  
1643 records, that the Río Paine Glacier was near its maximum extent during the ACR.

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3012 1644 *6.13 Tierra del Fuego*  
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3014 1645 Relatively little work has been done on ACR ice extent on Tierra del Fuego. [McCulloch](#)  
3015 [et al. \(2005a\)](#) propose extensive ice in the Cordillera Darwin as far north as the Isla  
3016 1646 Dawson adjacent to the Strait of Magellan during the ACR, but subsequent work has  
3017 1647 failed to support this hypothesis. Rather, evidence from bogs located near sea level up-  
3018 1648 ice of Isla Dawson suggests that there has not been any major re-expansion of Cordillera  
3019 1649 Darwin ice towards the Strait of Magellan since initial deglaciation during HS-1 ([Hall et](#)  
3020 1650 [al., 2013](#)). Similarly, a radiocarbon age from a bog on the south side of the mountains in  
3021 1651 front of Ventisquero Holanda indicates that the glacier has not reached more than 2 km  
3022 1652 beyond its present limit in the past ~15 ka ([Hall et al., 2013](#)). In the only confirmed case  
3023 1653 of ACR moraines in the region, [Menounos et al. \(2013\)](#) used <sup>10</sup>Be surface exposure ages  
3024 1654 of boulders to document an age of ~14 ka for a cirque moraine in the nearby Fuegian  
3025 1655 Andes. Other moraines in the Cordillera Darwin may date to the same period ([Hall,](#)  
3026 1656 [unpublished data](#)), but none has yet been dated adequately.  
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3029 1658 *6.14 Synthesis*  
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3036 1659 Glacier activity in North and Central America was very different from that in South  
3037 1660 America during the B-A interstadial (Table 3 and Fig. 5). This period was generally a  
3038 1661 time of rapid glacier retreat throughout North and Central America. Indeed, in many  
3039 1662 regions, glaciers completely disappeared during the B-A interstadial. Although evidence  
3040 1663 has been presented in some areas for minor advances during the B-A, uncertainties in  
3041 1664 numeric ages on which the conclusions are based do not preclude the possibility that the  
3042 1665 advances happened during the ACR.  
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3049 1666 The LIS experienced rapid retreat along much of its margin during the B-A. Documented  
3050 1667 local advances may be related more to surge processes than to climate, although there  
3051 1668 may be exceptions related to cooling during the Older Dryas (e.g. [Thompson et al., 2017](#)).  
3052 1669 Glaciers in Alaska retreated significantly, even beyond the limits they achieved in the late  
3053 1670 Holocene. In western Canada and in Washington State, the CIS retreated rapidly,  
3054 1671 especially from 14.5 ka to 13.5 ka. During this period of general retreat, however, the CIS  
3055 1672 and many alpine glaciers advanced between 13.9 ka and 13.3 ka. In the Central and  
3056 1673 Southern Rocky Mountains of Wyoming and Colorado, deglaciation had begun by 14.5  
3057 1674 ka, and most glaciers had disappeared by 13.5 ka. Although single-boulder <sup>10</sup>Be ages  
3058 1675 associated with moraines fall between 14.5 ka and 13.3 ka, no evidence of synchronous  
3059 1676 glacier advances within the B-A have been reported from these areas. In the Sierra  
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3071 1677 Nevada, glaciers retreated to cirque headwalls by about 15.5 ka. Some moraines within  
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3073 1678 Sierra cirques indicate that there were relatively minor advances, but again, it is not  
3074 1679 known if they date to the Inter-Allerød Cold Period, the ACR, or even the YD. In central  
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3076 1680 Mexico, recessional moraines close to moraines of the maximum advance date to 15-14  
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3078 1681 ka; after 14 ka, there was rapid glacier recession. Glaciers in Costa Rica disappeared by  
3079 1682 15.2 ka.  
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3081 1683 Glaciers in the Venezuelan Andes retreated during the B-A, whereas glaciers in several  
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3083 1684 regions of Central and Southern South America advanced during this period. In most  
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3085 1685 cases, these advances have been assigned to the ACR. For example, ACR advances have  
3086 1686 been proposed in the Colombian Andes, with temperatures about 3°C lower than today.  
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3088 1687 Multiple ACR advances have also been reported in the Peruvian and Bolivian Andes  
3089 1688 under a cold and relatively dry climate. There are no conclusive data from northern Chile  
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3091 1689 or the central Andes of Argentina, but it appears that there was a trend towards  
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3093 1690 deglaciation during the B-A. Existing data do not resolve whether minor glacier advances  
3094 1691 that have been recognized occurred during the ACR or the YD. There were several ACR-  
3095 1692 related glacier advances in Patagonia, with temperatures almost 2°C below current levels.  
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3097 1693 Only limited evidence of the ACR has been found in Tierra del Fuego.  
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3101 1695 **7. The Impact of the Younger Dryas (YD) (12.9-11.7 ka) and the Final Stages of**  
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3103 1696 **Deglaciation**

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3105 1697 *7.1 Younger Dryas concept*  
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3107 1698 The last period we consider in our review extends from the end of the B-A (12.9 ka) to  
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3109 1699 the beginning of the Holocene (11.7 ka). Again, the name coined by palynologists –  
3110 1700 Younger Dryas (YD) – is now widely used. Although the chronological limits derived  
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3112 1701 from palynology are controversial, this cold interval has now been defined in Greenland  
3113 1702 ice cores (Rasmussen et al., 2014). Undoubtedly, it is the most widely studied deglacial  
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3115 1703 period. Although climate varied extraordinarily during this period (Naughton et al.,  
3116 1704 2019), its effects in the Northern Hemisphere are clear – the AMOC weakened (Meissner,  
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3118 1705 2007; Muschitiello et al., 2019), sea ice expanded, and winter and spring temperatures  
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3120 1706 dropped drastically (Steffensen et al., 2008; Mangerud et al., 2016); summers remained  
3121 1707 relatively warm (Schenk et al., 2018). Glaciers in Europe advanced (Ivy-Ochs, 2015;  
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3123 1708 Mangerud et al., 2016), and the Asian monsoon weakened (Wang et al., 2008). Although  
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1709 the ITCZ migrated southward, precipitation changes in the tropics during the YD were  
1710 complex (Partin et al., 2015). Like HS-1, the YD was accompanied by warming in  
1711 Antarctica and an increase in atmospheric CO<sub>2</sub> (Broecker et al., 2010; Beeman et al.,  
1712 2019). The southern continents appear to have cooled slightly (Renssen et al., 2018),  
1713 although glaciers in New Zealand and Patagonia clearly retreated, an apparent  
1714 contradiction that has not been resolved (Kaplan et al. 2008, 2011; Martin et al., 2019;  
1715 Shulmeister et al., 2019).

1716 The causes of the abrupt YD anomaly continue to be a topic of debate. Changes in deep-  
1717 water circulation in the Nordic seas, weakening of the AMOC (Muschitiello et al., 2019),  
1718 moderate negative radiative forcing and altered atmospheric circulation (Renseen et al.,  
1719 2015; Naughton et al., 2019) likely played a role. Draining of Glacial Lake Agassiz after  
1720 intense melting of the Laurentide Ice Sheet during the B-A would have weakened the  
1721 AMOC and is supported by geomorphic evidence of this lake draining into the Gulf of  
1722 St. Lawrence and the North Atlantic at the end of the B-A (Leydet et al., 2018).  
1723 Additionally or alternatively, Glacial Lake Agassiz may have drained via the Mackenzie  
1724 River into the Arctic Ocean, also weakening the AMOC (Keigwin et al., 2018). The  
1725 hypothesis that the cause was external to the planet has recently attracted renewed interest  
1726 (Wolbach et al., 2018). In any case, the YD ended abruptly, with a 7 °C warming of some  
1727 regions in the Northern Hemisphere in only 50 years (Dansgaard et al., 1989; Steffensen  
1728 et al., 2008).

## 1729 *7.2 Laurentide Ice Sheet*

1730 The hemispheric extent of glaciation during the YD is summarized in Figure 6, and that  
1731 of the early Holocene is shown in Figure 7. The response of the LIS to the abrupt cooling  
1732 of the YD is complex and difficult to generalize, but most records appear to indicate that  
1733 recession slowed and that some major moraine systems were built, likely as a result of  
1734 marginal readvances (Dyke, 2004). For example, the largest end moraine belt along the  
1735 northwestern margin of the ice sheet, encompassing the Bluenose Lake moraine system  
1736 on the Arctic mainland and its correlative on Victoria Island, is now thought to have  
1737 formed due to YD cooling (Dyke and Savelle, 2000; Dyke et al., 2003). Similarly, there  
1738 are examples of readvances on Baffin Island, most notably in Cumberland Sound  
1739 (Jennings et al. 1996; Andrews et al. 1998). The large Gold Cove readvance of Labrador  
1740 ice across the mouth of Hudson Strait has also been assigned to the late stage of the YD,

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3189 1741 possibly in response to the rapid retreat of ice along the Hudson Strait (Miller and  
3190 1742 Kaufmann, 1990; Miller et al., 1999).  
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3193 1743 It has also been noted that several ice streams switched on during the YD, perhaps in  
3194 1744 response to a more positive ice sheet mass balance in some sectors (Stokes et al., 2016;  
3195 1745 Margold et al., 2018). Examples are two large lobes southwest of Hudson Bay (the Hayes  
3196 1746 and Rainy lobes), which readvanced towards the end of the YD. However, the precise  
3199 1747 trigger is uncertain; climatic forcing and dynamic instabilities related to meltwater  
3200 1748 lubrication and/or proglacial lake-level fluctuations are possibilities (Margold et al., 2018).  
3201 1749 Elsewhere, the M'Clintock Channel ice stream in the Canadian Arctic Archipelago (Clark  
3204 1750 and Stokes, 2001) is thought to have been activated during the early part of the YD and  
3205 1751 may have generated a large (60,000 km<sup>2</sup>) ice shelf that occupied Viscount Melville Sound  
3206 1752 (Hodgson, 1994; Dyke, 2004; Stokes et al., 2009). In contrast, the nearby Amundsen Gulf  
3207 1753 ice stream appears to have retreated rapidly during the early part of the YD, perhaps  
3208 1754 triggered by glacier retreat from a bathymetric pinning point into a wider and deeper  
3209 1755 channel (Lakeman et al., 2018).  
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3211 1756 The above examples highlight the difficulty of attempting to relate ice stream activity to  
3212 1757 external climate forcing. Overall, it appears that the LIS receded throughout the YD, but  
3213 1758 that the pace of recession slowed and there were notable readvances at the scale of  
3214 1759 individual lobes or ice streams. It should also be noted that while several moraine systems  
3215 1760 have been robustly linked to YD advances or stillstands, many others might also be  
3216 1761 correlative but have not yet been precisely dated (Dyke, 2004).  
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3218 1762 Following the YD, the LIS retreated rapidly in response to both increased summer  
3219 1763 insolation and increasing levels of carbon dioxide (Carlson et al., 2007, 2008; Marcott et  
3220 1764 al., 2013). Retreat proceeded back towards the positions of the major ice dispersal centers  
3221 1765 in the Foxe-Baffin sector, Labrador and Keewatin (Dyke and Prest, 1987; Dyke, 2004;  
3222 1766 Stokes, 2017). The final retreat of the Labrador Dome has recently been constrained by  
3223 1767 Ullman et al. (2016) using <sup>10</sup>Be surface exposure dating of a series of end moraines that  
3224 1768 likely relate to North Atlantic cooling (Bond et al., 1997; Rasmussen et al., 2006).  
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3226 1769 Following the last of these cold events at 8.2 ka (Alley et al., 1997; Barber et al., 1999),  
3227 1770 Hudson Bay became seasonally ice-free and deglaciation was completed by 6.7±0.4 ka  
3228 1771 (Ullman et al., 2016).  
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3230 1772 *7.3 Alaska*  
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1773 The existing literature offers limited evidence for glacier readvances in Alaska during the  
1774 YD. There may be many moraines that were deposited during or at the culmination of the  
1775 YD, but they have not been dated. One way to assess the possibility of there being YD  
1776 moraines in Alaska is to consider whether or not glaciers extended beyond their present  
1777 limits during the YD. Of the 14 glaciers throughout Alaska discussed by [Briner et al.](#)  
1778 [\(2017\)](#), nine had retreated up-valley of their late Holocene positions prior to the YD.  
1779 Thus, in some cases, it appears that glaciers did indeed extend down-valley of modern  
1780 limits during the YD. This was the case in Denali National Park and several sites in  
1781 southern Alaska. A notable site that provides the best evidence to date of YD glaciation  
1782 in the state is at Waskey Mountain in the Ahklun Mountains. The chronology of the  
1783 moraines at this locality has been updated since the work of [Briner et al. \(2002\)](#). [Young](#)  
1784 [et al. \(2019\)](#) report evidence for an early YD glacier culmination, followed by minor  
1785 retreat through the remainder of the interval.

1786 In terms of climate, [Kokorowski et al. \(2008\)](#) conclude that evidence for YD cooling is  
1787 mainly restricted to southern Alaska. [Kaufman et al. \(2010\)](#) argue that the coldest  
1788 temperatures in southern Alaska were at the beginning of the YD and that warming  
1789 occurred subsequently. This climatic pattern is consistent with the revised glacier  
1790 chronology of the Waskey Mountain moraines. [Denton et al. \(2005\)](#) hypothesized that  
1791 YD cooling was mostly a wintertime phenomenon and hence may have had limited effect  
1792 on glacier mass balance. This hypothesis is supported in Arctic Alaska with the  
1793 documentation of extreme winter temperature depression during the YD ([Meyer et al.,](#)  
1794 [2010](#)). Most of the pollen records summarized by [Kokorowski et al. \(2008\)](#) show no  
1795 significant cooling during the YD. In addition to the climate forcing transmitted from the  
1796 North Atlantic region, the Bering Land Bridge was flooded around the time of the YD  
1797 ([England and Furze, 2008](#)), although it may not have been completely covered by the sea  
1798 until about 11 ka ([Jakobsson, 2017](#)). This flooding event may have led to an increase in  
1799 precipitation due to more northerly storm tracks ([Kaufman et al., 2010](#)), which may have  
1800 influenced glacier mass balance. Additionally, the decreasing influence of LIS-induced  
1801 atmospheric reorganization may have affected summer temperature in Beringia during  
1802 the Late Pleistocene-Holocene transition. Of course, there may have been more glacier  
1803 fluctuations during the YD than is currently envisioned, because they may have occurred  
1804 under a climate that was similar to, or warmer than, that of the late Holocene ([Kurek et](#)

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1805 al., 2009; Kaufman et al., 2016), in which case moraines may have been destroyed by  
1806 Holocene glacier advances.

#### 1807 *7.4 Cordilleran Ice Sheet and the North Cascades*

1808 Many alpine glaciers and at least two remnant lobes of the CIS advanced during the YD.  
1809 In all cases, the advances were much smaller than those during the LGM and HS-1. At  
1810 alpine sites, most glaciers reached only several hundred meters beyond late Holocene  
1811 maximum positions attained during the Little Ice Age (Osborn et al., 2012; Menounos et  
1812 al., 2017). Other glaciers advanced and came into contact with stagnant CIS ice at lower  
1813 elevations (Lakeman et al., 2008). In the western North Cascades, there are multiple,  
1814 closely spaced moraines constructed during the YD (Riedel 2017). Radiocarbon dating  
1815 constrains the time of an advance on Mount Baker in the North Cascades to 13.0-12.3 ka  
1816 (K. Scott, written communication; Kovanen and Easterbrook, 2001). The Hyak II advance  
1817 in the southernmost North Cascades near Snoqualmie Pass occurred after 13 ka (Porter,  
1818 1976). Menounos et al. (2017) established <sup>10</sup>Be ages on 12 high-elevation moraines in  
1819 western Canada with a median age of 11.4 ka. A lobe of the CIS advanced across central  
1820 Fraser Lowland one or two times after 12.9 ka (Saunders et al., 1987; Clague et al., 1997;  
1821 Kovanen and Easterbrook, 2001; Kovanen, 2002), and the final advance of the glacier in  
1822 the Squamish River valley in the southern Coast Mountains north of Vancouver has been  
1823 dated to about 12.5 ka (Friele and Clague, 2002). It is not clear how long the CIS persisted  
1824 in each North Cascade mountain valley, but the middle reaches of Silver Creek were ice-  
1825 free by 11.6 ka, as were many sites in western Canada (Clague, 2017; Riedel, 2017). By  
1826 the beginning of the Holocene or shortly thereafter, ice cover in British Columbia was no  
1827 more extensive than it is today. A radiocarbon age from basal sediments in a pond  
1828 adjacent to the outermost Holocene moraine at Tiedemann Glacier in the southern Coast  
1829 Mountains shows that ice cover in one of the highest mountain areas in British Columbia  
1830 was, at most, only slightly more extensive at 11 ka than today (Clague, 1981; Arsenault  
1831 et al., 2007). This conclusion is supported by an age of 11.8-11.3 ka on a piece of wood  
1832 recovered from a placer gold mine near Quesnel, British Columbia, which is located near  
1833 the center of the former CIS (Lowdon and Blake, 1980).

1834 Alpine glacial ELAs associated with YD advances were 200-400 m below modern values  
1835 in the North Cascades, but fluctuated 100-200 m (Riedel, 2007). The colder YD climate  
1836 is also recorded in changes in loss-on-ignition carbon in lake bed sediments in the eastern  
1837 North Cascades (Riedel, 2017). Changes in pollen zone boundaries led Heusser (1977) to

3364  
3365  
3366 1838 conclude that YD mean July air temperature was 2-3°C cooler than today. [Liu et al. \(2009\)](#)  
3367  
3368 1839 suggested that annual precipitation increased by 250 mm, while mean annual air  
3369  
3370 1840 temperature was 4°C colder compared to the 1960-1990 average, and fluctuated by  
3371  
3372 1841  $\pm 0.5^\circ\text{C}$  during the YD interval.

3373 1842 *7.5 Rocky Mountain/Yellowstone region*  
3374

3375 1843 A YD glacier advance or stillstand has been documented in the Lake Solitude cirque in  
3376  
3377 1844 the Teton Range. Boulders perched on the small cirque lip date to  $12.9 \pm 0.7$  ka ([Licciardi](#)  
3378  
3379 1845 [and Pierce, 2008](#)). Glaciers in cirques in the Wind River Range advanced to form  
3380  
3381 1846 moraines or rock glaciers 50-300 m upvalley of Older Dryas/ HS-1 deposits (Fig. 10).  
3382  
3383 1847 Ages on these moraines in Stough Basin, Cirque of the Towers and Titcomb Basin are  
3384  
3385 1848 between 13.3 ka and 11.4 ka ([Shakun et al., 2015a](#); [Dahms et al., 2018](#); [Marcott et al.,](#)  
3386  
3387 1849 [2019](#)) and provide clear evidence of a glacier advance during the YD period. It is  
3388  
3389 1850 uncertain whether or not these glaciers disappeared prior to re-advancing to their YD  
3390  
3391 1851 positions.

3390 1852 There is clear evidence of a significant glacier advance in the Colorado Mountains during  
3391  
3392 1853 the YD ([Marcott et al., 2019](#)), confirming previous age assignments ([Menounos and](#)  
3393  
3394 1854 [Reasoner, 1997](#); [Benson et al., 2007](#)). Pollen studies ([Jiménez-Moreno et al., 2011](#); [Briles](#)  
3395  
3396 1855 [et al., 2012](#)) also indicate Younger Dryas cooling in the Colorado Rocky Mountains, as  
3397  
3398 1856 does a study of lacustrine sediment ([Yuan et al., 2013](#)) in the San Luis Valley of southern  
3399  
3400 1857 Colorado ([Leonard et al., 2017a](#)).

3400 1858 *7.6 Sierra Nevada*  
3401

3402 1859 The only known glacier advance between the retreat of the Tioga 4 glaciers at  $\sim 15.5$  ka  
3403  
3404 1860 and the late Holocene Matthes ('Little Ice Age') advance in the Sierra Nevada is the  
3405  
3406 1861 Recess Peak advance ([Bowerman and Clark, 2011](#)). Both cosmogenic surface exposure  
3407  
3408 1862 ages and independent regional climate records favor a YD age for this minor advance.  
3409  
3410 1863 However, limiting radiocarbon ages on bulk organic matter just above the Recess Peak  
3411  
3412 1864 till in lacustrine cores are between 14 ka and 13 ka ([Philips, 2017](#)), suggesting that the  
3413  
3414 1865 advance may be older than the YD. The weight of the evidence appears to still favor the  
3415  
3416 1866 YD age assignment, but the replicated direct radiocarbon measurements are difficult to  
3417  
3418 1867 dismiss. Confirmation of a YD age would support the model that the YD cooling had a  
3419  
3420 1868 detectable, although not major, impact on the deglacial climate of the west coast of North  
3421  
3422 1869 America. Confirmation of a slightly older age would suggest that there was a brief, but

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1870 significant episode of cooling there late during the B-A. In either case, the climate signal  
1871 is small compared to that of the GLGM. The linked glacial/lacustrine modeling of  
1872 [Plummer \(2002\)](#) yields a match to Recess Peak glacier extent and lake surface area in the  
1873 paleo-Owens River watershed, with a temperature reduction of 1°C and 140% of modern  
1874 precipitation. This local combination of glacial and closed-basin lacustrine records offers  
1875 an unusual opportunity to assess the paleoclimatic drivers of Recess Peak event, but the  
1876 significance of the event cannot be understood until the chronology is secure. Clearly,  
1877 additional radiocarbon and high-precision cosmogenic dating of the Recess Peak deposits  
1878 is a priority.

### 1879 *7.7 Mexico and Central America*

1880 Glaciers constructed a distinctive group of closely spaced end moraines in the mountains  
1881 of central Mexico at 3800-3900 m asl from 13-12 ka to ~10.5 ka ([Vázquez-Selem and](#)  
1882 [Lachniet, 2017](#)). ELAs were 4100-4250 m asl, which is 650-800 m below the modern  
1883 ELA, suggesting temperatures ~4-5°C below modern values. Considering that other  
1884 proxies generally show relatively dry conditions ([Lachniet et al., 2013](#)), the relatively low  
1885 ELAs were likely controlled by temperature.

1886 The terminal Pleistocene moraines of central Mexico provide clear evidence for Younger  
1887 Dryas glaciation in the northern tropics. The moraines are closely spaced and relatively  
1888 small (in general <6 m high near their front), but are well preserved in most mountain  
1889 valleys at elevations of 3800-3900 m asl. They suggest that glaciers remained near 3800-  
1890 3900 m asl for 1000-2000 years at the close of the Pleistocene, forming several small  
1891 ridges only tens of meters apart from one another ([Vázquez-Selem and Lachniet, 2017](#)).  
1892 Cosmogenic ages on glacially abraded surfaces indicate that mountains <4000 m asl in  
1893 central Mexico were ice-free by 11.5 ka, and mountains <4200 m asl became ice-free  
1894 between 10.5 ka and 10 ka. South-facing valleys were ice-free even earlier (12 ka)  
1895 ([Vázquez-Selem and Lachniet, 2017](#)). Glaciers on high peaks (Iztaccíhuatl, Nevado de  
1896 Toluca, La Malinche) receded, exposing polished bedrock surfaces below 4100 m asl,  
1897 from 10.5 ka to 9 ka. A brief, but distinctive glacier advance is recorded later, from ca.  
1898 8.5 ka to 7.5 ka, on the highest peaks of central Mexico (>4400 m asl) ([Vázquez-Selem](#)  
1899 [and Lachniet, 2017](#)) (Fig. 24).

1900 Cosmogenic exposure ages from the summit of Cerro Chirripó, Costa Rica, indicate that  
1901 the mountain was ice-free by 15.2 ka ([Cunningham et al., 2019](#)). However, other ages  
1902 suggest moraine formation around YD time ([Potter et al., 2019](#)) and complete

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1903 deglaciation thereafter (Orvis and Horn, 2000). The mountains of Costa Rica and likely  
1904 Guatemala were ice-free before 9.7 ka (Orvis and Horn, 2000).

### 1905 *7.8 Northern Andes*

1906 The evidence of possible YD glacier advances in the northern Andes is limited and mainly  
1907 restricted to elevations above 3800 m asl (Angel et al., 2017). Glacier advances in some  
1908 valleys in the Venezuelan Andes seem to be related to cooling during the YD. In the  
1909 Sierra Nevada, climate was dry, but temperatures were 2.2-3.8°C colder than today  
1910 between 12.9 ka and 11.6 ka (Salgado-Labouriau et al., 1977; Carrillo et al., 2008; Rull  
1911 et al., 2010; Stansell et al., 2010). Mahaney et al. (2008) suggest glaciers advanced in the  
1912 Humboldt Massif of this mountain range at 12.4 ka. In the Mucubají valley, also in this  
1913 range, small moraines located at elevations higher than 3800 m asl have yielded <sup>10</sup>Be ages  
1914 of 12.22±0.60 ka and 12.42±1.05 ka (modified ages from Angel, 2016) and may be  
1915 related to the YD. Glacier advances have been linked to the YD in the Sierra Nevada del  
1916 Cocuy, Colombia, based on <sup>10</sup>Be dating (Jomelli et al., 2014), and on the Bogota Plain  
1917 based on ages on lacustrine sediments behind the moraines (Helmens, 1988). There are  
1918 also moraines that might date to the YD in the Ecuadorian Andes. For example, in the  
1919 Chimborazo-Carihuairazo Massif, two moraine complexes have been radiocarbon-dated  
1920 to 13.4-12.7 cal ka BP (Clapperton and McEwan, 1985).

### 1921 *7.9 Peru and Bolivia*

1922 The weight of evidence suggests that glaciers were generally in retreat during the YD in  
1923 Peru and Bolivia. In the Cordillera Oriental of northern Peru, lake sediment records show  
1924 some evidence of readvance and reoccupation of higher cirques by glaciers, but no  
1925 moraines have been dated (Rodbell, 1993). Similar evidence from Vilcabamba in  
1926 southern Peru suggests glaciers advanced at the beginning of the YD, but then retreated  
1927 (Licciardi et al., 2009). Mercer and Palacios (1977) present evidence that glaciers  
1928 advanced near Quelccaya near the beginning and end of the YD. Similarly, Rodbell and  
1929 Seltzer (2000) and Kelly et al. (2012) provide radiocarbon-based evidence that sites in  
1930 the Cordillera Blanca and the Quelccaya Ice Cap advanced either just prior to or at the  
1931 start of the YD, followed by retreat. A cirque lake in Bolivia (16°S, headwall 5650 m asl)  
1932 formed before 12.7 ka, suggesting that ice had retreated by that time (Abbott et al., 1997).  
1933 According to Bromley et al. (2011), the ice cap on the Coropuna volcano experienced a  
1934 strong advance at ~13 ka. Similar glacier activity has been reported at Hualca Hualca  
1935 volcano (Alcalá-Reygosa et al., 2017) and Sajama (Smith et al., 2009) volcanoes. These

3541  
3542  
3543 1936 advances coincide with the highest level of the Coipasa paleo-lake cycle, confirming the  
3544  
3545 1937 high sensitivity of the glaciers in this region to shifts in humidity (Blard et al., 2009;  
3546 1938 Placzek et al., 2013) (Fig. 17).  
3548  
3549 1939 Many glaciers advanced or experienced stillstands in the Central Andes during the early  
3550 1940 Holocene. The mean age of all Holocene moraine boulders is  $11.0 \pm 0.4$  ka (Mark et al.,  
3551 1941 2017). In the Cordillera Huayhuash,  $^{10}\text{Be}$  samples from moraine boulders date from 11.4  
3552 1942 ka to 10.5 ka (Hall et al., 2009). Early Holocene (11.6-10.5 ka) moraines are also present  
3553 1943 on Nevado Huaguruncho (Stansell et al., 2015), and a moraine in the Cordillera  
3554 1944 Vilcabamba in southern Peru has been dated to  $\sim 10.5$  ka (Licciardi et al., 2009). Basal  
3555 1945 radiocarbon ages from lake sediments in the Cordillera Raura suggest ice-free conditions  
3556 1946 after 9.4 ka (Stansell et al., 2013). At Quelccaya in the Cordillera Vilcanota, peat overlain  
3557 1947 by till has been dated to 11.1 ka and 10.9 ka (Mercer and Palacios, 1977). Similarly, the  
3558 1948 Taptapa moraine on the Junin plain has been radiocarbon-dated to  $\sim 10.1$  cal ka BP  
3559 1949 (Wright, 1984). The Quelccaya Ice Cap reached its present extent by 10 cal ka BP, based  
3560 1950 on dated peat at its margin (Mercer, 1984). Glaciers in Peru seem not to have advanced  
3561 1951 throughout the remainder of the early Holocene. In Bolivia, however,  $^{10}\text{Be}$  ages suggest  
3562 1952 several advances during the early Holocene period (Jomelli et al., 2011, 2014).  
3570  
3571 1953 *7.10 Southern Bolivia and Northern Chile*  
3572  
3573 1954 As in the case of the ACR, there are no confirmed YD glacial landforms in the Arid  
3574 1955 Diagonal, but the chronology is insufficient to exclude minor glacial fluctuations in the  
3575 1956 high headwaters at this time (Ward et al., 2017).  
3577  
3578 1957 *7.11 Central Andes of Argentina*  
3579  
3580 1958 Published evidence of YD glacier activity exists at only two sites in the central Andes of  
3581 1959 Argentina. In the Nevado de Chañi, glaciers retreated after HS-1, followed by an advance  
3582 1960 during the YD (Martini et al., 2017a). Four  $^{10}\text{Be}$  ages from lateral and frontal moraines  
3583 1961 in the Chañi Chico valley average  $12.1 \pm 0.6$  ka (Fig. 19) (Martini et al., 2017a). The ELA  
3584 1962 during the YD advance was at  $\sim 5023$  m asl, which is 315 m above the GLGM ELA  
3585 1963 (Martini et al., 2017a). Moraines assigned to the YD have been found in two valleys in  
3586 1964 the Sierra de Aconquija (D’Arcy et al., 2019). According to D’Arcy et al. (2019), one  
3587 1965 moraine (M3a) was deposited at 12.5 ka and a second (M3b) at 12.3 ka. YD glacier  
3588 1966 advances coincided with a period of higher-than-present precipitation at paleo-lake  
3589 1967 Coipasa on the Altiplano (Blard et al., 2011; Placzek et al., 2013). No general early

3600  
3601  
3602 1968 Holocene glacier activity has been reported for the region, although two moraine boulders  
3603  
3604 1969 from Sierra de Aconquija yielded  $^{10}\text{Be}$  ages of 8.5 ka and 7.9 ka (D'Arcy et al., 2019). In  
3605  
3606 1970 northwestern Argentina, the presence of relict rock glaciers in cirques suggests that YD  
3607  
3608 1971 or early Holocene cooling may have activated rock glaciers instead of causing glaciers to  
3609  
3610 1972 re-form (Martini et al., 2013, 2017b).

3611 1973 *7.12 Patagonia*

3612  
3613 1974 After reaching their maximum Late Glacial extents during the ACR, Patagonian glaciers  
3614  
3615 1975 receded during the YD period. In some regions (47°-52°S), this general trend was  
3616  
3617 1976 interrupted by stillstands or minor readvances that deposited small moraines upvalley  
3618  
3619 1977 from the much larger ACR moraines (Moreno et al., 2009; Sagredo et al., 2011, 2018;  
3620  
3621 1978 Strelin et al., 2011; Glasser et al., 2012; Mendelova et al., 2020). Some of these advances  
3622  
3623 1979 may relate to the end of terminal calving following the draining of paleo-lakes in the  
3624  
3625 1980 region (Davies et al., 2018; Thorndycraft et al., 2019). Again, no evidence of glacier  
3626  
3627 1981 advances during the YD has been reported north of 47°S.

3628  
3629 1982 Paleo-vegetation records indicate a decline in precipitation during the YD in northwestern  
3630  
3631 1983 Patagonia (Jara and Moreno, 2014; Pesce and Moreno, 2014; Moreno et al., 2018),  
3632  
3633 1984 warm/wet conditions in central-western sectors (44°-48°S) (Villa-Martínez et al., 2012;  
3634  
3635 1985 Henríquez et al., 2017), and increased precipitation in southwestern sectors (48°-54°S)  
3636  
3637 1986 (Moreno et al., 2012, 2018). A widespread warm/dry interval is evident between 11 ka  
3638  
3639 1987 and 8 ka (Moreno et al., 2010). Although, most studies suggest that Patagonian glaciers  
3640  
3641 1988 retreated through the early Holocene, approaching their present-day configurations  
3642  
3643 1989 (Strelin et al., 2011; Kaplan et al., 2016), recent finding by Reynhout et al. (2019) at Torre  
3644  
3645 1990 glacier (49°S) show robust evidence of early renewed glacial activity during the early  
3646  
3647 1991 Holocene.

3648  
3649 1992 *7.13 Tierra del Fuego*

3650  
3651 1993 To our knowledge, there are no published data on glacier behavior during the Northern  
3652  
3653 1994 Hemisphere YD or at the start of the Holocene in the Cordillera Darwin. Glaciers are  
3654  
3655 1995 assumed to be restricted to the inner fjords. In the adjacent Fuegian Andes, an excavation  
3656  
3657 1996 just upvalley of an ACR moraine yielded a calibrated radiocarbon age on peat of ~12.2  
3658  
3659 1997 ka, indicating that the glacier had receded by that time (Menounos et al., 2013), possibly  
3660  
3661 1998 during the YD. In the same cirque, the presence of the Hudson tephra (7.96-7.34 ka)

3659  
3660  
3661 1999 within ~100 m of Little Ice Age moraines suggests the glacier had receded to the Little  
3662  
3663 2000 Ice Age limit by the early Holocene.  
3664  
3665 2001 *7.14 Synthesis*  
3666  
3667 2002 Information on glacier activity in the Americas during the YD is limited, but has been  
3668  
3669 2003 improving in recent years (Table 4 and Figs. 6 and 7).  
3670  
3671 2004 The LIS continued to thin and retreat throughout the YD, although at a lower rate than  
3672  
3673 2005 earlier. Some major moraine systems were built during YD stillstands or re-advances, but  
3674  
3675 2006 it is uncertain if they are a consequence of climate forcing or glacier dynamics related to  
3676  
3677 2007 internally driven instabilities. Evidence for YD advances is sparse in Alaska, and it seems  
3678  
3679 2008 that glacier retreat dominated there. In southern Alaska, however, temperatures decreased  
3680  
3681 2009 during the YD. It is possible that glaciers advanced during this period, but if so, the  
3682  
3683 2010 evidence was destroyed by late Holocene advances. There is evidence of YD glacier  
3684  
3685 2011 advances at the southwestern margin of the CIS and in the North Cascades, where a  
3686  
3687 2012 significant reduction in temperature and an increase in precipitation have been detected.  
3688  
3689 2013 In the Wyoming and Colorado Rocky Mountains, moraines in several cirque basins,  
3690  
3691 2014 which once were thought to be mid-Holocene ('Neoglacial') age, are now attributed to the  
3692  
3693 2015 YD. In the Sierra Nevada a minor advance may be attributed to the YD, although the  
3694  
3695 2016 dating is problematic. Many other small moraine complexes in the western mountains of  
3696  
3697 2017 the U.S. have yet to be dated.  
3698  
3699 2018 One of the few regions with obvious YD moraines is central Mexico, where reconstructed  
3700  
3701 2019 ELAs suggest temperatures were ~4-5°C below modern values in an environment that  
3702  
3703 2020 was drier than today. The evidence for YD glaciation in the mountains of Costa Rica is  
3704  
3705 2021 inconclusive, but in the Northern Andes at elevations above 3800 m asl, some glaciers  
3706  
3707 2022 advanced during the YD due to a decrease in temperatures of 2.2-3.8°C below present  
3708  
3709 2023 values under a dry climate.  
3710  
3711 2024 Glaciers continued to retreat in Peru and Bolivia during the YD, except on the Altiplano  
3712  
3713 2025 where the YD coincided with the highest level of the Coipasa paleo-lake cycle and with  
3714  
3715 2026 advances of glaciers in numerous mountain ranges and on high volcanoes. It is  
3716  
3717 2027 questionable whether some late advances in northern Chile occurred during the ACR or  
3718  
3719 2028 the YD, but most of the Arid Diagonal was already ice-free in the YD. Some evidence  
3720  
3721 2029 for YD advances has been found in the central Andes of Argentina, but in Patagonia  
3722  
3723 2030 glacier retreat continued throughout the YD and was interrupted only by stillstands or



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3719  
3720 2031 minor readvances that deposited small moraines. Glacier retreat also dominated during  
3721  
3722 2032 the YD on Tierra del Fuego, where there is no evidence for advances during this period.  
3723  
3724 2033 Deglaciation accelerated after the YD in nearly all of North, Central and South America,  
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3726 2034 and most small glaciers reached their current size or disappeared during the early  
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3728 2035 Holocene. In the area of the LIS, deglaciation occurred rapidly following the YD and was  
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3730 2036 largely complete by 7 ka. In Alaska, glaciers reached sizes similar to today in the early  
3731  
3732 2037 Holocene. The CIS had disappeared by the beginning of the Holocene. Most glaciers in  
3733  
3734 2038 the Yellowstone region and the Colorado Rocky Mountains disappeared before the  
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3736 2039 Holocene, and in the Sierra Nevada glaciers were about their current size at that time. In  
3737  
3738 2040 central Mexico, glaciers probably reached their current size or disappeared by the  
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3740 2041 beginning of the Holocene, although a minor advance, probably related to the 8.2 ka  
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3742 2042 event, is recorded on the highest volcanoes. Many glaciers advanced or experienced  
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3744 2043 stillstands in the Central Andes under a wetter climate during the early Holocene,  
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3746 2044 although these glaciers apparently rapidly retreated a short time thereafter. In Patagonia  
3747  
3748 2045 and Tierra del Fuego, glaciers retreated during the early Holocene and most glaciers  
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3750 2046 approached their present size at that time.  
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3752 2047

## 3748 2048 **8. Discussion**

### 3750 2049 *8.1. The climatic meaning of the Last Glacial Termination*

3753 2050 Before comparing glacier behavior in the different regions of the Americas, we first  
3754  
3755 2051 summarize the state of knowledge of global climate evolution during the Last Glacial  
3756  
3757 2052 Termination and the mechanisms that caused it. An immediate problem in attempting  
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3759 2053 such a summary is that it is difficult to even define the start and end of this period. It  
3760  
3761 2054 encompasses a set of events that do not begin or end at the same time around the world.  
3762  
3763 2055 In addition, deglaciation may be caused, not only by changes in orbital forcing that  
3764  
3765 2056 regulate the amount of insolation that Earth receives (Broecker and van Donk, 1970), but  
3766  
3767 2057 also by internal forcing mechanisms and feedbacks, including changes in atmospheric  
3768  
3769 2058 circulation and composition, especially in CO<sub>2</sub> and CH<sub>4</sub> (Sigman and Boyle, 2000;  
3770  
3771 2059 Monnin et al., 2001; Sigman et al., 2010; Shakun et al., 2012; Deaney et al., 2017),  
3772  
3773 2060 changes in ocean circulation, the composition of the oceans and sea ice extent (Bereiter  
3774  
3775 2061 et al., 2018), and the interplay between the atmosphere and oceans (Schmittner and  
3776  
3777 2062 Galbraith, 2008; Fogwill et al., 2017). Finally, the Last Glacial Termination is difficult to

3777  
3778  
3779 2063 define because the two hemispheres experience opposing external forcing and potentially  
3780  
3781 2064 opposing internal forcing mechanisms that might induce a climate compensation effect  
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3783 2065 between the hemispheres termed the “bipolar seesaw” (Broecker and Denton, 1990).  
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3785 2066 Broadly speaking, glacial terminations initiate when the ice sheets of the Northern  
3786 2067 Hemisphere are at their maximum extent and with global sea level at its lowest (Birchfield  
3787 and Broecker, 1990; Imbrie et al., 1993; Raymo, 1997; Paillard, 1998). Additionally,  
3788 2068 global deglaciation during each termination operates over approximately the same length  
3789 2069 of time during each glacial cycle and is characterised by short-lived fluctuations of rapid  
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3791 2070 glacier retreat and occasional re-advances (Lea et al., 2003). Given these observations, it  
3792 2071 is important to determine the mechanisms responsible for the climatic and glacial changes  
3793 2072 that accompany deglaciations. To that end, several hypotheses have been proposed that  
3794 2073 are mainly based on the temperatures of the oceans (Voelker, 2002) and the composition  
3795 2074 of the atmosphere (Severinghaus and Brook, 1999; Stolper et al., 2016).  
3796 2075  
3799 2076 Any attempt to closely examine the Last Glacial Termination must account for changes  
3800  
3801 2077 in ocean temperature throughout this period. These temperature changes are simultaneous  
3802 2078 in the two hemispheres, but can shift in opposite directions, for example in the Atlantic  
3803 2079 Ocean. They are determined by the greater or lesser intensity of the AMOC (see syntheses  
3804 2080 in Barker et al., 2009, 2010). Even though these changes occur throughout deglaciation,  
3805 2081 the amount of CO<sub>2</sub> in the atmosphere tends to increase more or less continuously. A  
3806 2082 possible explanation for this apparent enigma is that the oceans in one hemisphere may  
3807 2083 cool while those in the other hemisphere warm and emit more CO<sub>2</sub>, redistributing heat  
3808 2084 across the planet (Barker et al., 2009).  
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3810 2085 Building on previous work (Cheng et al., 2009), Denton et al. (2010) propose that a  
3811 2086 concatenation of processes, with multiple positive feedbacks, drive deglaciation. They  
3812 2087 argue that deglaciation is initiated by coincident “excessive” growth of Northern  
3813 2088 Hemisphere ice sheets and increasing boreal summer insolation due to orbital forcing.  
3814 2089 The large volume of ice on northern continents results in maximum isostatic depression  
3815 2090 and an increase in the extent of the ice sheets that are marine-based. Even a small increase  
3816 2091 in insolation could, under these conditions, enlarge ablation zones and initiate the collapse  
3817 2092 of Northern Hemisphere ice sheets. Marine-based ice sheets can also be more vulnerable  
3818 2093 to collapse due to positive feedbacks associated with sea-level rise at the grounding line.  
3819 2094 Outbursts of meltwater and icebergs from these ice sheets cool the North Atlantic Ocean  
3820 2095 and weaken the AMOC, leading to an expansion of winter sea ice and very cold winters  
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2096 on the adjacent continents (Denton et al., 2010). Under these conditions, the northern  
2097 polar front expands, driving the ITCZ, the southern trade winds, and the southern  
2098 westerlies to the south (Denton et al., 2010). The Asian monsoon weakens, while the  
2099 cooling over the North Atlantic intensifies the South American monsoon (Novello et al.,  
2100 2017) and the southern westerlies. The result is an increase in upwelling in the southern  
2101 oceans, accompanied by enhanced ocean ventilation and a rise in atmospheric CO<sub>2</sub>.  
2102 During deglaciation, the Southern Hemisphere warms first, followed by warming over  
2103 the rest of the planet (Broecker, 1998). Southward migration of the southern westerlies  
2104 also contributes to a temperature rise in the southern oceans, which transfer heat to the  
2105 south (Denton et al., 2010). The intensive cooling in the Northern Hemisphere ends due  
2106 to a reduction in meltwater input, northward retreat of sea ice, and renewed warming of  
2107 the northern oceans, which reestablish the AMOC. Subsequently, the ITCZ returns  
2108 northward and the Asian monsoon intensifies. The northward migration of the southern  
2109 westerlies and the intensification of the AMOC cool the southern oceans, completing a  
2110 cycle in the recurrent bipolar seesaw that ultimately tends toward equilibrium (Denton et  
2111 al., 2010).

2112 The hypothesis for climate evolution during Last Glacial Termination summarized above  
2113 can be tested with our dataset on the behavior of glaciers in the Americas. It is clear that  
2114 sea level depends on how water is distributed between the ocean and Northern  
2115 Hemisphere ice sheets during glacials and also on the effects of land-based glacier ice  
2116 cover on the isostatic balance of the northern continents (Lambeck et al., 2014). The  
2117 hypothesis that deglaciation begins when northern ice sheets are extremely large is still  
2118 supported (Abe-Ouchi et al., 2013; Deaney et al., 2017). According to Cheng et al. (2016),  
2119 for example, ice sheets reach their maximum size after five precession cycles, which may  
2120 explain why glacial cycles finished after similar durations of about 115 ka (Paillard,  
2121 1998). In addition, it seems that the time needed for ice sheets to reach this extreme size  
2122 increased throughout the Pleistocene (Clark et al., 2006), and successively more  
2123 insolation energy was required to start deglaciation. This might explain why each glacial  
2124 cycle is longer than its predecessor (Tzedakis et al., 2017). Deglaciation begins when the  
2125 excessive size of the northern ice sheets coincides with: (i) increasing insolation in boreal  
2126 summer in the Northern Hemisphere, mainly at 65° N, the average latitude of large  
2127 northern ice sheets (Kawamura et al., 2007; Brook and Buizert, 2018); (ii) minimum CO<sub>2</sub>  
2128 in the atmosphere (Shakun et al., 2012); and (iii) maximum sea ice extent (Gildor et al.,

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3896  
3897 2129 [2014](#)). These conditions induce aridity and reduce vegetation cover, which in turn  
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3899 2130 increases atmospheric dust, reducing albedo on northern ice sheets ([Ellis and Palmer,](#)  
3900 2131 [2016](#)). Better knowledge of the activity of glaciers throughout the Americas may confirm  
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3902 2132 the hypothesis that is central to these models, namely that deglaciation begins in the North  
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3904 2133 and is transmitted to the South.  
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3906 2134 New information from ocean and polar ice cores reinforces the idea of climate  
3907 2135 compensation between the two hemispheres (the bipolar seesaw) during the Last Glacial  
3908  
3909 2136 Termination. Intensive sea-level rise occurs within the first 2 kyr of deglaciation,  
3910 2137 inducing retreat of marine-based ice sheets, and acts as a positive feedback for  
3911  
3912 2138 deglaciation ([Grant et al., 2014](#)). [Fogwill et al. \(2017\)](#) argue that, once deglaciation starts,  
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3914 2139 it is driven by global oceanic and atmospheric teleconnections. New data support the idea  
3915 2140 that meltwater cooling of the Northern Hemisphere reduced the AMOC strength ([Deaney](#)  
3916 2141 [et al., 2017](#); [Muschitiello et al., 2019](#)) and pushed the northern westerlies southward in  
3917 2141  
3918 2142 Asia ([Chen et al., 2019](#)), Europe ([Naughton et al., 2019](#)), and North America ([Hudson et](#)  
3919 2142  
3920 2143 [al., 2019](#)). The Asian summer monsoon weakened during these cold periods in the  
3921 2144 Northern Hemisphere ([Cheng et al., 2016](#); [Chen et al., 2019](#)), and the Indian summer  
3922 2144  
3923 2145 monsoon transferred Southern Hemisphere heat northward, promoting subsequent  
3924 2145  
3925 2146 Northern Hemisphere deglaciation ([Nilsson-Kerr et al., 2019](#)).  
3926  
3927 2147 New data have also highlighted the importance of CO<sub>2</sub> storage in the dense deep waters  
3928 2148 of the Southern Hemisphere during glacials ([Fogwill et al., 2017](#); [Clementi and Sikes,](#)  
3929 2148  
3930 2149 [2019](#)). Ventilation of these waters during deglaciation emits a large amount of CO<sub>2</sub> into  
3931 2149  
3932 2150 the atmosphere and significantly warms the planet ([Stephens and Keeling, 2000](#);  
3933 2151 [Anderson et al., 2009](#); [Skinner et al., 2010](#); [Brook and Buizert, 2018](#); [Clementi and Sikes,](#)  
3934 2151  
3935 2152 [2019](#)), favoring deglaciation ([Lee et al., 2011](#); [Shakun et al., 2012](#)). This increase in CO<sub>2</sub>  
3936 2152  
3937 2153 overrides the cooling effect from orbital variations in the Southern Hemisphere ([He et al.,](#)  
3938 2153  
3939 2154 [2013](#)).  
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3941 2155 Recent high-resolution data from ice cores in Antarctica and Greenland have helped  
3942 2156 verify the opposite temperature trends in the two polar areas during deglaciation, at least  
3943 2157 on a large scale. Moreover, these data confirm that the rise in CO<sub>2</sub> was synchronous with  
3944 2157  
3945 2158 the increase in Antarctic temperatures ([Ahn et al., 2012](#); [Beeman et al., 2019](#)). Antarctic  
3946 2158  
3947 2159 temperature seems to be more closely linked to changes in tropical ocean currents,  
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3949 2160 whereas Greenland is less affected by this phenomenon ([Wolff et al., 2009](#); [Landais et](#)  
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3951 2161 [al., 2015](#)). However, the intimate relationship between AMOC intensity and atmospheric

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2162 CO<sub>2</sub> concentrations has been clearly demonstrated (Deaney et al., 2017), and deglaciation  
2163 largely represents a period of imbalance between these two parameters. Therefore, when  
2164 the AMOC stabilizes, atmospheric CO<sub>2</sub> concentrations stabilize and the interglacial  
2165 period begins (Deaney et al., 2017). That said, some exceptions have been detected on a  
2166 centennial scale (Böhm et al., 2015). A sudden surge in the AMOC may cause a large  
2167 release of CO<sub>2</sub> into the atmosphere, although only for a few centuries (Chen et al., 2015).  
2168 New studies propose that mean ocean temperature and the temperature of Antarctica are  
2169 closely related, underlining the importance of the Southern Hemisphere ocean in  
2170 orchestrating deglaciation (Bereiter et al., 2018), as it is the main contributor of CO<sub>2</sub> to  
2171 the atmosphere (Beeman et al., 2019). Inverse temperature evolution and the latitudinal  
2172 migration of atmospheric circulation systems (fronts, ITCZ, trade winds and westerlies)  
2173 may have the greatest impact on the planet's glaciers and ice sheets, albeit in opposite  
2174 directions.

2175 Improved knowledge of the changing extent of glaciers throughout the Americas is  
2176 necessary to understand how glaciers are affected by the above-described evolution of  
2177 the climate system during the Last Glacial Termination. However, little attention has been  
2178 focussed on the differing behavior of glaciers between the hemispheres and how this  
2179 might reflect global ocean and atmospheric teleconnections during the last deglaciation.  
2180 Is there a glacial bipolar seesaw reflected in the behavior of mountain glaciers? In that  
2181 sense, it is necessary to consider that mountain glaciers today contribute about one-third  
2182 of the ice melt to the oceans (Gardner et al., 2013; Bamber et al., 2018). One of the few  
2183 studies that compares the behavior of mountain glaciers in both hemispheres is that of  
2184 Shakun et al. (2015a). These authors analyzed 1116 cosmogenic nuclide exposure ages  
2185 (mostly <sup>10</sup>Be ages) from glacial landforms located between 50°N and 55°S on different  
2186 continents, but mostly from the Americas. Inferred glacier behavior was evaluated using  
2187 a variety of climate forcings. Their results demonstrate that glaciers responded  
2188 synchronously throughout deglaciation, mainly due to the global increase of CO<sub>2</sub> in the  
2189 atmosphere and the subsequent increase in temperature. They note important regional  
2190 differences related to other factors, such as insolation in the Northern Hemisphere, a  
2191 seesaw response to changes in the AMOC in the Southern Hemisphere, and changes in  
2192 precipitation distribution and in tropical ocean currents.

2193 *8.2 Glaciers in the Americas during GLGM in a global context*

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2194 We note the similarity in the times of glacier advances in North and Central America  
2195 during the GLGM. Most mountain glaciers reached their maximum extent before or  
2196 during the GLGM, although in some areas (e.g. Teton Range and portions of the  
2197 Yellowstone Ice Cap), the maximum may also have encompassed HS-1. However, there  
2198 were many local differences within each region. Similarly, the LIS did not exhibit  
2199 uniform evolution along all parts of its margin, although it did reach its maximum extent  
2200 during the GLGM. Based on our synthesis, the LIS began to retreat about 21 ka ago at  
2201 the same time as the majority of the North and Central American glaciers, as well as  
2202 European glaciers. The Scandinavian Ice Sheet reached its maximum extent in the GLGM  
2203 and also started its retreat about 21 ka (Toucanne et al., 2015; Cuzzone et al., 2016;  
2204 Hughes A.L.C. et al., 2016; Hughes, P. et al., 2016; Stroeven et al., 2016; Patton et al.,  
2205 2017). As in the case of the LIS, the margins of the Scandinavian Ice Sheet retreated at  
2206 different times; retreat in some areas was delayed until HS-1. The same timing and  
2207 behavior has been reported for the Barents, British-Irish, and Icelandic ice sheets (Hormes  
2208 et al., 2013; Pétursson et al., 2015; Hughes A.L.C. et al., 2016). In all three cases, the  
2209 maximum was reached during the GLGM and deglaciation began about 21-20 ka,  
2210 although retreat did not begin in some areas until HS-1. Some sectors of the British-Irish  
2211 Ice Sheet began their retreat very early in the GLGM, although for reasons related to the  
2212 dynamics of the ice sheet rather than climate (Ó Cofaigh et al., 2019). In the case of the  
2213 Icelandic Ice Sheet, sea-level rise caused it to collapse after 19 ka (Pétursson et al., 2015).  
2214 In summary, the Last Glacial Termination started almost simultaneously in areas covered  
2215 by the LIS and the Eurasian ice sheets.

2216 Similarities are also evident between North and Central American and many European  
2217 mountain glaciers. The glacial maximum in Europe extended from 30 ka until the  
2218 beginning of deglaciation 21-19 ka, for example in the Alps (Ivy-Ochs, 2015), Apennines  
2219 (Giraudi et al., 2015), Trata Mountains (Makos, 2015; Makos et al., 2018) and the  
2220 Anatolia peninsula mountains (Akçar et al., 2017). However, glaciers in some mid-  
2221 latitude mountains in Europe achieved their maximum size much earlier, between MIS 5  
2222 to MIS 3, for example, glaciers in the Cantabrian Mountains and central Pyrenees on the  
2223 Iberian Peninsula (Oliva et al., 2019) and the High Atlas in North Africa (Hughes et al.,  
2224 2018). In contrast, glaciers in the eastern Pyrenees, the Central Range and the Sierra  
2225 Nevada on the Iberian Peninsula, which are also located in mid-latitudes, clearly attained  
2226 their maximum size during the GLGM (Oliva et al., 2019). In summary, mountain

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4074 2227 glaciers in Europe and North America evolved in a similar way, in spite of the local  
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4076 2228 differences within each region.  
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4078 2229 Comparing glacial behavior in South America to glaciers in other continents at similar  
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4080 2230 latitudes is more difficult. The extra-American glaciers of the Southern Hemisphere are  
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4082 2231 located in isolated mountains of Africa and Oceania and, with one exception, have not  
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4084 2232 been well studied. The exception is the Southern Alps of New Zealand, which are located  
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4086 2233 at the about the same latitudes as northern Patagonia.  
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4088 2234 Unlike most of North and Central America, the maximum advance of the last glacial cycle  
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4090 2235 throughout South America, except perhaps in Tierra del Fuego and in some mountains of  
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4092 2236 Patagonia, was reached long before the GLGM, which is between approximately 60 ka  
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4094 2237 and 40 ka. A similar pattern is also evident in mountains of East Africa ([Shanahan and](#)  
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4096 2238 [Zreda, 2000; Mahaney, 2011](#)), New Zealand ([Schaefer et al., 2015; Darvill et al., 2016](#))  
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4098 2239 and Kerguelen ([Jomelli et al., 2018](#)). Outside the Southern Hemisphere, an early  
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4100 2240 maximum advance during the last glacial cycle has also been proposed in some mountains  
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4102 2241 in Mexico ([Heine, 1988](#)), in the central Pyrenees, the Cantabrian Mountains ([Oliva et al.,](#)  
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4104 2242 [2019](#)), and in the High Atlas ([Hughes et al., 2018](#)). However, these cases are exceptional  
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4106 2243 and are not located in any latitudinal zone; rather they are purely regional. Some authors  
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4108 2244 have suggested the idea of an aborted termination around 65-45 ka in the Southern  
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4110 2245 Hemisphere, after glaciers had achieved their maximum extents ([Schaefer et al., 2015](#)).  
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4112 2246 Although the GLGM was not the last time that glaciers advanced in the Southern  
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4114 2247 Hemisphere, it was a period of widespread glacier expansion under a mainly cold and wet  
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4116 2248 climate. As in the Andes, many glaciers in the mountains of East Africa ([Shanahan and](#)  
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4118 2249 [Zreda, 2000; Mahaney, 2011](#)) and New Zealand ([Schaefer et al., 2015; Darvill et al.,](#)  
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4120 2250 [2016; Shulmeister et al., 2019](#)) advanced during the GLGM and left outer moraine  
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4122 2251 systems. This advance has been attributed to a southward migration of the ITCZ and  
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4124 2252 westerlies in response to strong cooling in the Northern Hemisphere ([Kanner et al., 2012;](#)  
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4126 2253 [Schaefer et al., 2015; Darvill et al., 2016](#)). Paleoclimate records from central Chile,  
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4128 2254 northwestern Patagonia, and the southeast Pacific, however, imply a northward shift in  
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4130 2255 southwesterly winds during the GLGM ([Heusser, 1990; Villagrán, 1988a, 1988b; Heusser](#)  
4130 2256 [et al., 1999; Lamy et al., 1999; Moreno et al., 1999, 2018](#)).  
4130 2257 Deglaciation in the Patagonian Andes began at 17.8 ka, consistent with Antarctic ice core  
4130 2258 records ([Erb et al., 2018](#)) and New Zealand glacial chronologies ([Schaefer et al., 2015;](#)  
4130 2259 [Darvill et al., 2016; Barrell et al., 2019; Shulmeister et al., 2019](#)). It occurred two or three

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4133 2260 millennia after the inception of deglaciation in the Northern Hemisphere, although  
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4135 2261 researchers have noted that ice recession and moderate warming took place during the  
4136 2262 Varas Interstade, between ~24 and ~19 ka (Mercer, 1972, 1976; Lowell et al., 1995;  
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4138 2263 Denton et al., 1999; Hein et al., 2010; Mendelova et al., 2017).

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4140 2264 The GLGM was more than a climatic period; it was the time when the world's glaciers  
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4142 2265 achieved their maximum extent after the previous interglacial. However, ice masses did  
4143 2266 not all behave in the same way because their activity was affected by topography, regional  
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4145 2267 changes in ocean and atmospheric circulation, local climatic conditions and climate  
4146 2268 feedbacks (Liakka et al., 2016; Patton et al., 2017; Liakka and Lofverstrom, 2018;  
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4148 2269 Licciardi and Pierce, 2018). Although many glaciers reached their maximum extent well  
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4150 2270 before the GLGM, especially in the Southern Hemisphere, the northern ice sheets grew  
4151 2271 more-or-less continuously towards the GLGM. Many glaciers, also in the Northern  
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4153 2272 Hemisphere, achieved their largest size just before the GLGM. Within each region,  
4154 2273 glaciers advanced many times, conditioned by the geographical constraints arising from  
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4156 2274 their own expansion. When the next warm orbital cycle began to affect the Northern  
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4158 2275 Hemisphere, around 21 ka, deglaciation started in all areas, although it was somewhat  
4159 2276 delayed in the Southern Hemisphere. Again, the duration and intensity of deglaciation  
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4161 2277 after the GLGM differed greatly and was regional rather than latitudinal.

4162  
4163 2278 *8.3 Glaciers in the Americas during HS-1 in a global context*

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4165 2279 Records of glacier behavior in the Americas during HS-1 are consistent with records from  
4166 2280 other continents, albeit with considerable local variability in each region. HS-1 did not  
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4168 2281 have a strong impact on the LIS and European ice sheets (Patton et al., 2017). Around  
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4170 2282 17.8 ka, however, some of the margins of these ice sheets stabilized or advanced and  
4171 2283 moraines were built at their margins. Retreat began again shortly thereafter, around 17.5  
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4173 2284 ka, with some local oscillations superimposed on overall retreat through the rest of HS-1  
4174 2285 (Cuzzone et al., 2016; Hughes A.L.C. et al., 2016; Peters et al., 2016; Stroeven et al.,  
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4176 2286 2016; Gump et al., 2017; Patton et al., 2017). Although there may have been an internal  
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4178 2287 reorganization of flow patterns and ice sheet geometry at this time, we note that there is  
4179  
4180 2288 little evidence for any major readvance of the LIS during during HS-1, and retreat likely  
4181 2289 continued in most regions including the southern margin (Heath et al., 2018). The  
4182  
4183 2290 European ice sheets evolved in a similar manner (Toucanne et al., 2015), with  
4184 2291 deglaciation beginning between 21 ka and 19 ka (Patton et al., 2017) and rapidly leading  
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4186 2292 to huge ice losses. The meltwater contribution to the North Atlantic, mainly from the LIS,



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2293 was enough to drastically reduce the AMOC (Toucanne et al., 2015; Stroeven et al.,  
2294 2016).

2295 Knowledge of the impacts of HS-1 on mountain glaciers in Europe is stronger than in  
2296 North and Central America. Glaciers advanced in the Alps during HS-1 (Gschnitz  
2297 stadial), occupying valley bottoms that had been deglaciated earlier. The main advance  
2298 was at the beginning of HS-1, around 17-16 ka, and its moraines are recognized in many  
2299 valleys (Ivy-Ochs, 2015). An advance of the same age has been documented in the Tatra  
2300 Mountains, (Makos, 2015; Makos et al., 2018). Up to three readvances have been  
2301 recognized in the Appenines during HS-1 (Giraudi et al., 2015), and glaciers advanced  
2302 on the Anatolian Peninsula near the end of HS-1 (Sarikaya et al., 2014, 2017). Glacial  
2303 advances around 17-16 ka are recognized in almost all mountain ranges on the Iberian  
2304 Peninsula (Oliva et al., 2019). As in the Alps, glaciers on the Iberian Peninsula reoccupied  
2305 the lower reaches of valleys, approaching the moraines of the GLGM (Palacios et al.,  
2306 2017a). This was the case in the central and eastern Pyrenees, the Central Range and the  
2307 Sierra Nevada (Oliva et al., 2019). All of these European advances happened at about the  
2308 same time as the HS-1 glacier advances in North and Central America. In the Rocky  
2309 Mountains, Mexico and Central America, maximum GLGM advances appear to have  
2310 extended into HS-1, although it is possible that glaciers readvanced during HS-1,  
2311 surpassing and erasing GLGM glacial landforms. The most recent summaries of the  
2312 glacial chronology of the California Sierra Nevada (Phillips, 2017), Wyoming's Wind  
2313 River Range (Dahms et al., 2018, 2019; Marcott et al., 2019), the European Alps (Ivy-  
2314 Ochs, 2015), and Iberian Sierra Nevada (Palacios et al., 2016) indicate that the glaciers  
2315 in these mountain systems behaved in similar ways during HS-1.

2316 The marked glacier advances in the tropical Andes during HS-1 have been attributed to  
2317 an intensification of the South American monsoon under a colder climate (Kanner et al.,  
2318 2012). The monsoon produced a wet period on the Altiplano and caused glaciers to  
2319 advance in the surrounding mountains, in many cases beyond the limits of the GLGM  
2320 moraines. Again, it is difficult to compare South American tropical glaciers to other  
2321 glaciers at similar latitudes. There is little information on the glaciers of East Africa; ages  
2322 bearing on deglaciation have a large margin of error that precludes assigning events to  
2323 HS-1 with confidence, although many of the glaciers show evidence of large late-glacial  
2324 oscillations (Shanahan and Zreda, 2000; Mahaney, 2011).

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4251 2325 It is much easier to compare glacier behavior between temperate southern latitudes, such  
4252 as Patagonia and the Southern Alps of New Zealand and Kerguelen Archipelago. In these  
4253 2326 mountain ranges, deglaciation accelerated during HS-1 (Darvill et al., 2016). The glaciers  
4254 2327 retreated throughout the entire HS-1 period in the Southern Alps (Putnam et al., 2013;  
4255 2328 Koffman et al., 2017), Kerguelen (Jomelli et al., 2018) and the same happened in  
4256 2329 Patagonia (Mendelova et al., 2017) and Tierra del Fuego (Hall et al., 2013). Moraines in  
4259 2330 some valleys of the Southern Alps dated to 17 ka were built at the end of a prolonged  
4261 2331 GLGM and mark the beginning of large-scale deglaciation (Barrell et al., 2019;  
4262 2332 Shulmeister et al., 2019).  
4263 2333  
4264 2334 On both continents, deglaciation began about 21 ka and became much more widespread  
4265 2335 after 19-18 ka, resulting in a steady rise in sea level, cooling of the North Atlantic, and  
4266 2336 reduction of the AMOC. HS-1 was a short period of stabilization and reduction in ice  
4267 2337 loss. Strong cooling occurred in temperate northern latitudes, and mountain glaciers  
4268 2338 advanced close to the limits reached during the GLGM, in some cases even surpassing  
4269 2339 them. This happened in spite of increased aridity, which was a consequence of the  
4270 2340 southward migration of the polar front. The ITCZ also migrated southward, particularly  
4271 2341 over the tropical Atlantic, thereby intensifying the South American monsoon and  
4272 2342 increasing precipitation in tropical latitudes, where glaciers advanced considerably. In  
4273 2343 contrast, in the temperate latitudes of the Southern Hemisphere, HS-1 was a warm period  
4274 2344 and glaciers began or continued their rapid retreat under a climate that was opposite that  
4275 2345 in the temperate latitudes of the Northern Hemisphere.  
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4277 2346 *8.4 Glaciers in the Americas during the B-A and ACR in a global context*  
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4279 2347 We have seen above that the southern and western margins of the LIS retreated during  
4280 2348 the B-A, whereas there was minimal retreat along its northern margin. In the rest of North  
4281 2349 and Central America, many glaciers retreated significantly or disappeared altogether by  
4282 2350 the end of HS-1 and during the B-A. However, some studies have suggested that there  
4283 2351 were short periods of glacier advance in Europe, as in North America, during this period  
4284 2352 of general deglaciation. The European ice sheet retreated from the sea and through central  
4285 2353 Europe during the B-A (Cuzzone et al., 2016) and separated into smaller ice sheets  
4286 2354 centered on the Scandinavian Peninsula, Svalbard and Novaya Zemlya (Hughes A.L.C.  
4287 2355 et al., 2016; Patton et al., 2017). Along the southern and northwestern margins of the  
4288 2356 Scandinavian Peninsula, there are moraine systems that mark the end of GS-2.1, and other  
4289 2357 moraines inboard of them relate to the cold stage of GI-1d, called the Older Dryas

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2358 (Mangerud et al., 2016, 2017; Stroeven et al., 2016; Romundset et al., 2017). Moraines  
2359 were built at the margin of the British-Irish ice sheet about 14 ka in the Older Dryas, but  
2360 that ice sheet had nearly disappeared along the B-A (Ballantyne et al., 2009; Hughes  
2361 A.L.C. et al., 2016; Wilson et al., 2019). The ice sheet covering Iceland retreated and left  
2362 important parts of the interior of the island free of ice during the B-A, with a climate  
2363 similar to that of today (Pétursson et al., 2015). In summary, we conclude that European  
2364 ice-sheets behaved in a similar manner to American glaciers during the B-A.

2365 Glaciers in the European Alps experienced the same rapid deglaciation during the B-A as  
2366 in North and Central America. After advances during HS-1, alpine glaciers retreated  
2367 considerably during the B-A and had practically disappeared by the end of this period  
2368 (Ivy-Ochs, 2015). Older Dryas (Daun stadial) moraines have been identified in many  
2369 alpine valleys, indicating stagnation or an advance of glaciers between the two  
2370 interstadials (Ivy-Ochs, 2015). After an advance in the Tatra Mountains at about 15 ka,  
2371 glaciers retreated rapidly (Makos, 2015; Makos et al., 2018). This retreat was interrupted  
2372 by readvances of glaciers during the Older Dryas (Marks et al., 2019). Glaciers also  
2373 rapidly retreated in the eastern Mediterranean during the B-A (Dede et al., 2017; Sarikaya  
2374 and Çiner, 2017; Sarikaya et al., 2017). The same pattern is evident in the Balkans (Styllas  
2375 et al., 2018), the Apennines (Giraudi et al., 2015), and the Iberian Peninsula where  
2376 glaciers disappeared from some mountain systems or retreated into the interior of cirques  
2377 (Oliva et al., 2019). Some moraines in these mountains may have been built in the Older  
2378 Dryas cold period, but uncertainties in the cosmogenic ages are sufficiently large that this  
2379 possibility cannot be confirmed. Examples are found in the central Pyrenees (Palacios et  
2380 al., 2017b).

2381 Conditions were warmer in Venezuela, and glaciers retreated, during B-A. The B-A warm  
2382 interstadial is not reflected in Central and Southern South American glacier behavior.  
2383 Rather, there is clear evidence from the northern and tropical Andes of advances during  
2384 the ACR. These advances occurred under cold and arid conditions caused mainly by  
2385 temperature changes related to the strengthening of the AMOC (Jomelli et al., 2014).  
2386 Although there is currently no evidence of these advances in northern Chile and the  
2387 central Andes of Argentina, they are clear in Patagonia, where after a retreat of glaciers  
2388 during HS-1, there was an advance during the ACR (Strelin et al., 2011; García et al.,  
2389 2012). Evidence has been found also of a glacier advance during the ACR in the Fuegian

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4369 2390 Andes on Tierra del Fuego (Menounos et al., 2013), although there is no conclusive  
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4371 2391 evidence of an ACR event in the adjacent Cordillera Darwin (Hall et al., 2019).  
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4373 2392 Again, a comparison of the behavior of South American glaciers to glaciers in other areas  
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4375 2393 of the Southern Hemisphere is almost impossible. In East Africa, as is the case on other  
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4377 2394 continents, it is very difficult to place the ages of some moraines within the ACR or YD  
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4379 2395 (Mahaney, 2011). However, in the Southern Alps of New Zealand, at Kerguelen (Jomelli  
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4381 2396 et al., 2018) as in Patagonia, there is evidence of glacier advances during the ACR, but  
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4383 2397 with large regional variations, possibly related to the westerlies (Darvill et al., 2016). In  
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4385 2398 any case, most of the possible ACR moraines in New Zealand have been dated to 13 ka,  
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4387 2399 at the boundary between the ACR and the YD (Shulmeister et al., 2019). There are  
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4389 2400 indications of a decrease in temperature of 2-3°C at that time, at least in some areas of the  
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4391 2401 Southern Alps (Doughty et al., 2013). In spite of the limited knowledge of glacier  
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4393 2402 evolution over much of this region, we conclude that South American glaciers evolved in  
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4395 2403 a similar way to glaciers at similar latitudes on other continents and opposite to that of  
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4397 2404 glaciers in the Northern Hemisphere.

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4399 2405 Deglaciation of the Northern Hemisphere accelerated under a warm climate in the lead-  
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4401 2406 up to the interglacial period. Again, we observe differences in glacial behavior within  
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4403 2407 each region, in both North America and Europe, but these differences are local and relate  
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4405 2408 to the geography of ice sheets and mountain glaciers and not to latitudinal trends. In the  
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4407 2409 best-studied regions where glacial landforms are well preserved, the impacts of short cold  
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4409 2410 intervals on glaciers have been detected, especially in the Older Dryas. However, in most  
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4411 2411 cases, uncertainties in dating preclude correctly assigning landforms to brief climatic  
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4413 2412 periods. Thus, it is not yet possible to determine whether any moraines in the Northern  
4414  
4415 2413 Hemisphere belong to the ACR or the Older Dryas. Glacier behavior in the Southern  
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4417 2414 Hemisphere is different from that in North America and on other northern continents. The  
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4419 2415 impact of cooling during the ACR is clear in some of its regions. In the Southern  
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4421 2416 Hemisphere, there was no massive, continuous glacier melt, but rather a tendency towards  
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4423 2417 stagnation or glacier advance. As was the case for HS-1, the north and the south  
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4425 2418 responded in opposite ways to climate change during the B-A/ACR period.

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4427 2419 *8.5 Glaciers in the Americas during the YD in a Global Context*

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4429 2420 In many cases, the glaciers in North and Central America responded to the YD by  
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4431 2421 advancing. Retreat of the LIS slowed, and some sectors advanced. At the end of the YD  
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4433 2422 period, the LIS renewed its retreat. Similarly, several fronts of the Fennoscandian Ice  
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2423 Sheet advanced during the YD, although there was great variability in its different  
2424 margins (Cuzzone et al., 2016; Hughes A.L.C. et al., 2016; Mangerud et al., 2016;  
2425 Stroeven et al., 2016; Patton et al., 2017; Romundset et al., 2017). It appears that the  
2426 maximum advance occurred at the end of the YD period, at least at some margins  
2427 (Mangerud et al., 2016; Romundset et al., 2017). At the beginning of the Holocene, retreat  
2428 of the Fennoscandian Ice Sheet began anew, although this was interrupted during the  
2429 short-lived Preboreal oscillation, at 11.4 ka. Afterwards, retreat continued until the ice  
2430 sheet disappearance at 10-9 ka (Cuzzone et al., 2016; Hughes A.L.C. et al., 2016;  
2431 Stroeven et al., 2016). During the YD, ice caps in some sectors of Britain, Franz Josef  
2432 Land and Novaya Zemlya also expanded (Hughes A.L.C. et al., 2016; Patton et al., 2017;  
2433 Bickerdike et al., 2018). The glaciers in Iceland recovered during the YD and again  
2434 brought their fronts close to the present shoreline and, in some cases, beyond it (Pétursson  
2435 et al., 2015). With the Holocene came rapid retreat, interrupted by the Preboreal  
2436 oscillation at 11.4 ka (Andrés et al., 2019). In summary, the remnant European ice sheets  
2437 grew during the YD, but similar growth is less evident for the LIS.

2438 The impact on glaciers of the YD is currently being studied in the mountains of North  
2439 and Central America. In Alaska, the only clear evidence reported to date is in the south.  
2440 However, it is evident that there were small advances during the YD in many valleys of  
2441 British Columbia and the North Cascades. Recent dating has provided much more  
2442 evidence of small advances in the Wyoming and the Colorado Rocky Mountains  
2443 (Leonard et al., 2017a; Dahms et al., 2018, 2019) and possibly the California Sierra  
2444 Nevada (Phillips, 2017). YD advances are clear in central Mexico and increasingly  
2445 certain in Central America. Glaciers advanced throughout the European Alps during the  
2446 YD (Egesen stadial) and built moraines intermediate in position between those of the  
2447 Oldest Dryas and those of the Little Ice Age (Ivy-Ochs, 2015). In some valleys, there are  
2448 moraines dating to the Preboreal oscillation that lie between those of the YD and the Little  
2449 Ice Age (Ivy-Ochs, 2015). From glacier ELA depressions, it can be inferred that the  
2450 annual temperature was 3-5°C cooler in the Alps, the Tatra Mountains and elsewhere in  
2451 the Carpathians (Rinterknecht et al., 2012; Makos, 2015). New information shows that  
2452 glaciers advanced in cirques in the Mediterranean mountains, for example in the  
2453 Anatolian Peninsula (Sarkaya and Çiner, 2017), the Balkans (Styllas et al., 2018), the  
2454 Apennines (Giraudi et al., 2015), the Iberian mountains (García-Ruiz et al., 2016; Oliva  
2455 et al., 2019), the French Pyrenées (Jomelli et al., under review) and the High Atlas

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2456 (Hughes et al., 2018). At this time, we can conclude that the activity of glaciers in North  
2457 America and Europe during the YD is more similar than it appeared a few years ago.

2458 Glaciers also apparently advanced in the northern Andes during the YD. However, in the  
2459 central Andes, the YD was a period of glacier retreat, with the exception of the Altiplano  
2460 where the Coipasa wet phase coincided with advances in the surrounding mountains.  
2461 Glaciers may also have advanced in these mountains at the beginning of the Holocene,  
2462 around 11 ka, but they all retreated after 10 ka. Glaciers in Patagonia and on Tierra del  
2463 Fuego retreated after the ACR, in the latter area probably beyond the limits of the Little  
2464 Ice Age. Glaciers in East Africa retreated immediately after constructing ACR or YD  
2465 moraines (Mahaney, 2011). In the Southern Alps, as in Patagonia, the YD was a period  
2466 of glacier retreat (Shulmeister et al., 2019) with temperatures about 1°C warmer than  
2467 today (Koffman et al., 2017).

2468 Glaciers in the Northern Hemisphere responded synchronously to YD cooling by either  
2469 stabilizing or advancing, but the timing of the maximum extent differs spatially, as does  
2470 the magnitude of advance; in many areas, there is no evidence for YD glacier activity. In  
2471 the Southern Hemisphere, the South American monsoon intensified, thereby increasing  
2472 humidity, which caused tropical glaciers to advance. However, in the temperate latitudes  
2473 of this hemisphere, glaciers retreated, once again showing their antiphase behavior  
2474 compared to those in the north.

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## 2476 **9. Conclusions**

2477 The decrease in temperature in the Americas during the GLGM was 4-8 °C, but changes  
2478 in precipitation differed considerably throughout this large region. Consequently, many  
2479 glaciers of North and Central America reached their maximum extent during the GLGM,  
2480 whereas others reached it later, during the HS-1 period. In the Andes, for example,  
2481 glaciers advanced during the GLGM, but this advance was not the largest of the Last  
2482 Glacial, except possibly on Tierra del Fuego. HS-1 was a time of glacier growth  
2483 throughout most of North and Central America; some glaciers built new moraines beyond  
2484 those of the GLGM. Glaciers in the tropical Andes stabilized or advanced during HS-1  
2485 and, in many cases, overrode GLGM moraines. However, glaciers in the temperate and  
2486 subpolar Andes retreated during this period. Glaciers retreated throughout North and  
2487 Central America during the B-A interstadial and, in some cases, disappeared. Glaciers

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2488 advanced during the ACR in some parts of the tropical Andes and in the south of South  
2489 America. This advance was strong in Patagonia. Limited advances have been documented  
2490 in high mountain valleys in North and Central America during the YD. In contrast,  
2491 glaciers retreated during this interval in South America, except in some sectors of the  
2492 northern Andes and on the Altiplano where glacier advances coincided with the highest  
2493 level of the Coipasa paleo-lake cycle.

2494 In summary, the GLGM was the culmination of glacier growth during the last glacial  
2495 cycle. Glaciers achieved their maximum extent in many sectors before the GLGM, and  
2496 even in individual sectors at different times, but the main northern ice sheets were largest  
2497 within the GLGM. The latter explains why orbital forcing triggered deglaciation  
2498 beginning about 21 ka across the Northern Hemisphere and somewhat later in the  
2499 Southern Hemisphere.

2500 Glaciers in North America and Europe exhibit common behavior at all latitudes through  
2501 the Last Glacial Termination. This synchronous behavior extended almost to the Equator.  
2502 This commonality was clearly influenced by pronounced shifts in ocean circulation (e.g.  
2503 the AMOC), but probably also reflected proximity to the great Northern Hemisphere ice  
2504 sheets that profoundly affected atmospheric circulation and temperature.

2505 Glaciers at temperate latitudes in the Southern Hemisphere fluctuated synchronously,  
2506 especially those in Patagonia and the Southern Alps of New Zealand. Their behavior is  
2507 generally opposite to that of Northern Hemisphere glaciers during HS-1 and the B-  
2508 A/ACR, but the two are similar at the beginning and end of Last Glacial Termination.

2509 Glaciers at tropical latitudes in the Southern Hemisphere show greater diversity in their  
2510 behavior, which is most likely related to shifts in the ITCZ. A striking feature of the  
2511 glacial history of Central America and the tropical Andes is the persistence of relatively  
2512 extensive mountain glaciers through the Younger Dryas, long after those in North  
2513 America and Europe had retreated close to Holocene limits. One significant difference  
2514 between much of the Andes and the Northern Hemisphere is that the combination of  
2515 extreme elevation and aridity produces a larger sensitivity to precipitation than for the  
2516 lower and wetter mountain ranges of North America and Europe.

2517 Once deglaciation began, there was a seesaw between the hemispheres, which affected  
2518 not only marine currents but also atmospheric circulation and glacier behavior. This  
2519 seesaw explains the opposing behavior of many glaciers in the Northern and Southern

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2520 Hemispheres during HS-1 and the B-A/ACR. At the end of the B-A, it appears that many  
2521 mountain glaciers and minor ice sheets had achieved sizes similar to those of the early  
2522 Holocene. Subsequently, the YD ended deglaciation in the south and led to the re-advance  
2523 of some glaciers in the north.

2524

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2532

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4550 **Figure captions**

4551 Figure 1. Locations of the main sites in North and Central America cited in the text.

4552 Figure 2. Locations of the main sites in South America cited in the text.

4553 Figure 3. Glacier extent during the Global Last Glacial Maximum in the Americas.

4554 Coloured areas represent regions containing glaciated mountain ranges and in many cases

4555 are, for purposes of visibility, much larger than actual glaciated areas. AK = Alaska, BC

4556 = British Columbia (Cordilleran Ice Sheet and northern Cascades), CS = central/southern

4557 Cascades, SN = Sierra Nevada, NRM = northern Rocky Mountains, SRM = southern

4558 Rocky Mountains, MX = Mexico, CA = Central America, NA = Northern Andes, PB =

4559 Peru/Bolivia, NCA = north-central Andes, ACA = arid central Andes, PA = Patagonia,

4560 TdF = Tierra del Fuego. The coastline corresponds to the GLGM sea-level low. Figure

4561 information comes from author interpretations and references cited in Table 1.

4562 Figure 4. Glacier extent during H-1 in the Americas. See Figure 3 for full caption. Figure

4563 information come from author interpretations and references cited in Table 2.

4564 Figure 5. Glacier extent during the B-A and ACR in the Americas. See Figure 3 for full

4565 caption. Figure information comes from author interpretations and references cited in

4566 Table 3.

4567 Figure 6. Glacier extent during the YD in the Americas. See Figure 3 for full caption.

4568 Figure information comes from author interpretations and references cited in Table 4.

4569 Figure 7. Glacier extent during the early Holocene in the Americas. See Figure 3 for full

4570 caption. Figure information comes from author interpretation and references cited in

4571 Table 4.

4572 Figure 8. Glacial landforms at Deming Glacier on Mount Baker in the North Cascade

4573 Range, Washington State. Late-glacial moraines below the present-day terminus of

4574 Deming Glacier are numbered 1 through 7. Moraines 1 and 7 and the dashed black line

4575 represent the maximum late glacial (Bølling?) limit of Deming Glacier. Moraine number

4576 4 marks the YD limit with its associated radiocarbon ages. Note Neoglacial - Little Ice

4577 Age terminus.

4578 Figure 9. GLGM moraines in in the Beartooth Range, Rocky Mountains, Montana. A)

4579 Location of the photos. B) Moraines; view west. C) Moraines; view east. The moraines

4580 have been dated to 19.8 ka (Licciardi and Pierce, 2018). Photos by Nuria Andrés.

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4581 Figure 10. A) Location of the Wind River Range in the Rocky Mountains of Wyoming,  
4582 Montana and Idaho, and the Middle and North forks of the Popo Agie River (red box) on  
4583 the southeast flank of the range. B) Overview of LLGM and Late-glacial (pre-Holocene)  
4584 moraines in the Middle and North fork catchments of the Popo Agie River. Only the  
4585 farthest extents of the dated moraines are indicated in the main valleys. Yellow –  
4586 Positions and  $^{10}\text{Be}$  ages of terminal LGM and post-LGM (Pinedale and post-Pinedale)  
4587 moraines. Green – Positions and  $^{10}\text{Be}$  ages of moraines associated with glacial activity  
4588 during the H-1/Oldest Dryas period. Red – Positions and  $^{10}\text{Be}$  ages of moraines associated  
4589 with the YD period. The glacial geology and chronology of this area were originally  
4590 described by [Dahms \(2002, 2004\)](#) and [Dahms et al. \(2010\)](#), and were subsequently  
4591 revised by [Dahms et al. \(2018, 2019\)](#). Source: Google Earth images.

4592 Figure 11. LLGM moraines on the east side of the Teton Range, Wyoming, near Taggart  
4593 and Jenny lakes. A) Locations of photos. B) Moraines east of Taggart Lake. C) Moraines  
4594 west of Jenny Lake. The moraines have been dated to between 14.4 and 15.2 ka ([Licciardi  
4595 and Pierce, 2018](#)). Photos by Nuria Andrés.

4596 Figure 12. Recessional moraines along the Yellowstone River in the Rocky Mountains of  
4597 Montana. A) Locations of photo. B) Moraines; view from the northeast. The moraines  
4598 have been dated to between 14.4 and 15.1 ka ([Licciardi and Pierce, 2018](#)). Photo by Nuria  
4599 Andrés.

4600 Figure 13. GLGM moraines in the Clear Creek watershed, Sawatch Range, central  
4601 Colorado Rocky Mountains). A) Location of photos. B) Moraines; view to the West. C)  
4602 Moraines; view to the east. The moraines have been dated to between 19.1 and 21.7 ka  
4603 ([Young et al., 2011](#)). Photos by Nuria Andrés.

4604 Figure 14. Overview of late Pleistocene moraines at Bishop Creek, eastern Sierra Nevada,  
4605 California. Extents of preserved moraines are indicated for the Tioga 1 (early GLGM),  
4606 Tioga 2/3 (late GLGM), and Tioga 4 (H-1) advances. Recess Peak moraines (B-A or YD)  
4607 are numerous throughout the headwaters area, but only one representative moraine is  
4608 shown. The obvious terminal moraines northeast of the Tioga 1 moraine date to MIS 6  
4609 age. The geology and chronology of this drainage are described in [Phillips et al. \(2009\)](#)  
4610 and updated in [Phillips \(2017\)](#).

4611 Figure 15. Glacial landforms between ~3100 and 4200 m a.s.l. in Gavidia Valley in the  
4612 Mérida Andes The valley has a U-shaped cross-profile, and has numerous outcrops of

8674  
8675  
8676 4613 striated and polished bedrock (roches moutonnées). Deglaciation happened in two stages:  
8677  
8678 4614 slow retreat between ~22 and 16.5 ka, followed by the complete deglaciation at ~16 ka  
8679 4615 ([Angel et al., 2016](#)). Photo by Eduardo Barreto.  
8680  
8681 4616 Figure 16. LLGM and H-1 glacial landforms at the top of Cerro Tunupa (19.8°S, 67.6°W;  
8682 4617 5110 m asl); view toward the southeast ([Blard et al, 2013a](#); [Martin et al., 2018](#)). In the  
8683 4618 background is the Salar de Uyuni. Photo by Pierre Henri Blard.  
8686  
8687 4619 Figure 17. Glacial landforms on Hualcahualca volcano, southern Peru. A) Locations of  
8688 4620 photos. B) Prominent well-preserved LLGM moraine on the east flank of the volcano. C)  
8689 4621 H-1 and YD moraines on the north flank of the volcano ([Alcalá-Reygosa et al., 2017](#)).  
8690 4622 Photos by Jesus Alcalá-Reygosa.  
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8693 4623 Figure 18. Glacial landforms in the arid Chilean Andes, from north to south: A) Glacial  
8694 4624 valley at El Tatio (22.3° S), view upvalley from the LGM right-lateral moraine that has  
8695 4625 yielded <sup>36</sup>Cl ages of 20-35 ka ([Ward et al., 2017](#)). Truck circled for scale. B) View  
8696 4626 upvalley from LGM right-lateral/frontal moraine near Co. La Torta (22.45° S) dated to  
8697 4627 25-30 ka ([Ward et al., 2015](#)). Dirt track visible for scale. C) Inner ridge of the western  
8700 4628 terminal complex of the former Chajnantor ice cap (23.0° S), last occupied at the LLGM  
8701 4629 ([Ward et al., 2015; 2017](#)). View to the southwest; backpack circled for scale. D) Eastern  
8702 4630 terminal moraine complex at Chajnantor, likely MIS 3 ([Ward et al., 2015, 2017](#)). View  
8703 4631 to the north; largest visible boulders are ~1.5 m in diameter. E) View upvalley from likely  
8704 4632 MIS 6 terminal moraine of the southern outlet glacier of the former ice field at Cordon  
8705 4633 de Puntas Negras (23.85° S). Sharper inner lateral/frontal moraines have yielded ages that  
8706 4634 support an LGM and/or MIS 3 age ([Thornton, 2019](#)). Locations of photos: 1 - A; 2 -B , 3  
8707 4635 - C and D, 4 - E. Photos by Dylan J. Ward.  
8708  
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8711 4636 Figure 19. Glacial landforms on the east side of Nevado de Chañi in the central Argentina  
8712 4637 Andes. A) Locations of the photos. B) View to the east (down-valley) of Refugio Valley  
8713 4638 showing the GLGM I (red) and H-1 (purple) moraines. The green circle marks a hut for  
8714 4639 scale. C) YD lateral/frontal moraines in the Chañi Chico valley, which is a tributary of  
8715 4640 Refugio Valley. These moraines are located inboard of those shown in panel A. Photos  
8716 4641 by Mateo Martini.  
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8721 4642 Figure 20. Glacial landforms on Tierra de Fuego. A) Locations of photos B-E. B) The  
8722 4643 inner LGM-age moraine at Bahía Inútil, southern Chile. View to the east, parallel to the  
8723 4644 moraine (marked by the white dashed line). The large boulders are granite derived from  
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4645 the Cordillera Darwin. C) Ice-scoured terrain in the Cordillera Darwin characteristic of  
4646 areas deglaciated during H-I. View to the north along the Beagle Channel. D) Aerial view  
4647 to the southwest of Marinelli fjord. The dashed line shows the location of a late Holocene  
4648 moraine, marking a historic glacial margin. The white dot shows the approximate location  
4649 of a core site that provides evidence that ice receded from the Strait of Magellan and back  
4650 into the fjord by ~17 ka. The black dot shows the location of photo E. E) Bog in Marinelli  
4651 fjord; view to the south. The late Holocene moraine is visible in the background. Ice had  
4652 cleared this site, within view of the historic position, by 16 ka. Photos by Brenda Hall.

4653 Figure 21. ACR moraine in Gueshgue valley in the Cordillera Blanca, Peru (dated by  
4654 [Stansell et al., 2017](#)). Photo by Joseph Licciardi.

4655 Figure 22. ACR moraine at Nevado Huaguruncho in the Eastern Cordillera of Peru (dated  
4656 by [Stansell et al., 2015](#)). Photo by Joseph Licciardi.

4657 Figure 23. ACR (continuous white line) and Holocene (dashed line) moraines in the  
4658 Tranquilo Valley, close to Mount San Lorenzo, Patagonia. Photo by Esteban Salgado.

4659 Figure 24. Glacial landforms in Alcalican Valley, southwest of Iztaccíhuatl in central  
4660 Mexico. Moraine from the Late Pleistocene-Holocene transition. Elevations are ~3865 m  
4661 at the bottom of the valley at the end of the moraine and ~5200 m a.s.l. on the mountain  
4662 summit. Moraines of this group have been <sup>36</sup>Cl-dated at 13-12 to ~10.5 ka and could be  
4663 YD in age. Note the three moraine ridges on the right side of the valley. Photo by Lorenzo  
4664 Vázquez.

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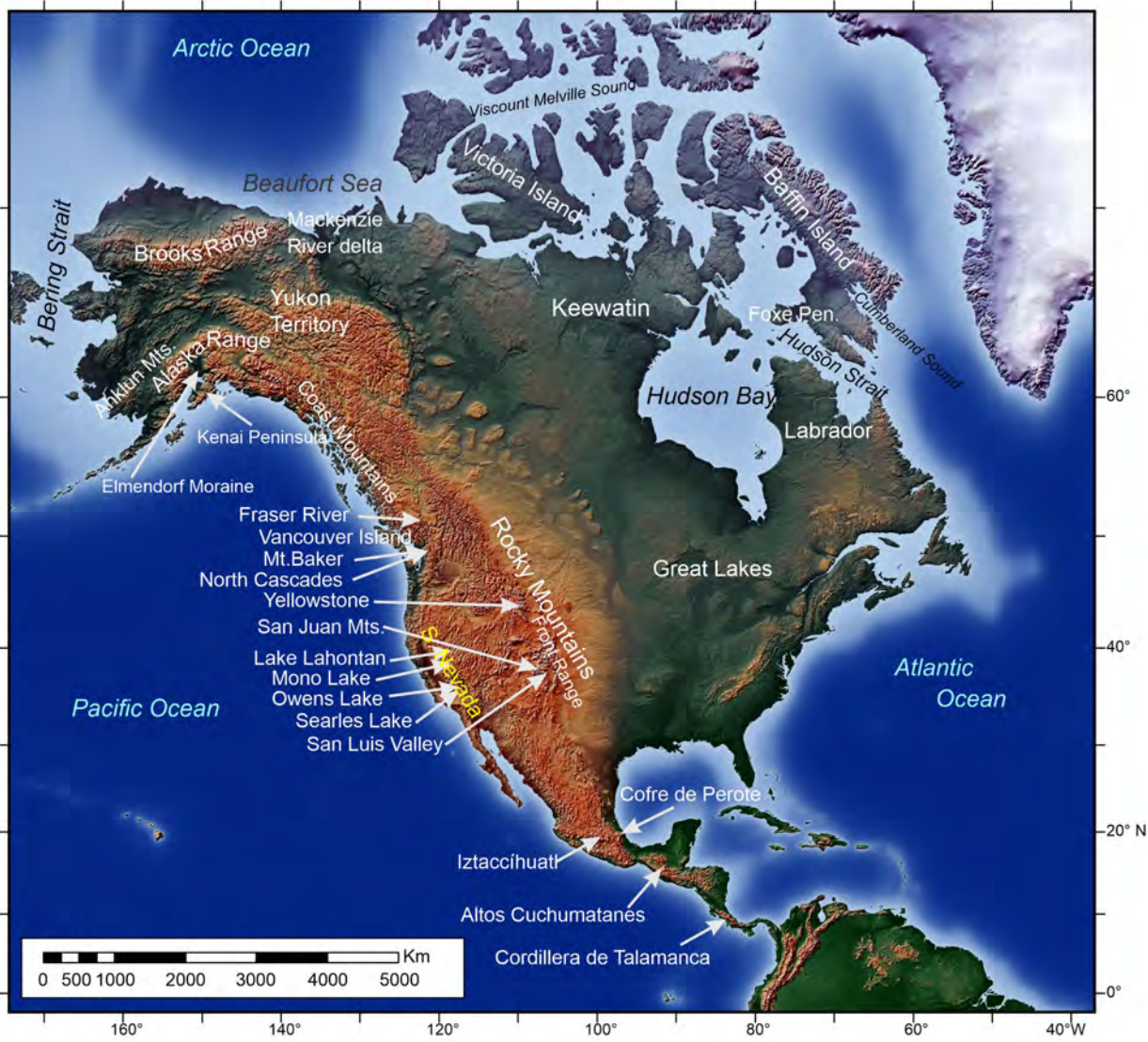
#### 4666 **Table captions**

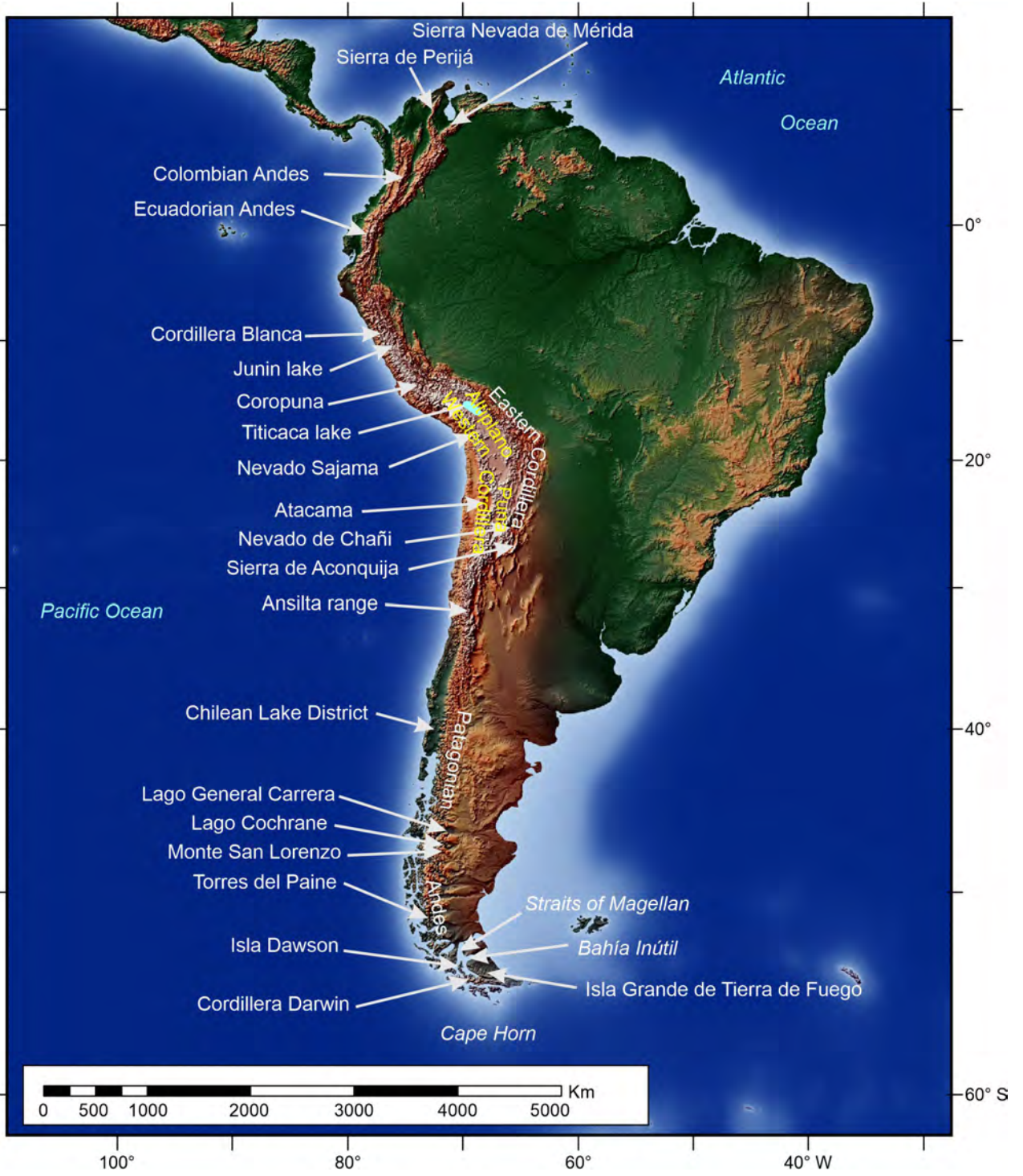
4667 Table 1. The main climate and glacial evolution features during the Global Last Glacial  
4668 Maximum (26.5-19 ka) in the Americas.

4669 Table 2. The main climate and glacial evolution features during the Heinrich 1 Stadial  
4670 (17.5-14.6 ka) in the Americas.

4671 Table 3. The main climate and glacial evolution features during the Bølling-Allerød  
4672 interstadial and Antarctic Cold Reversal (14.6-12.9 ka) in the Americas.

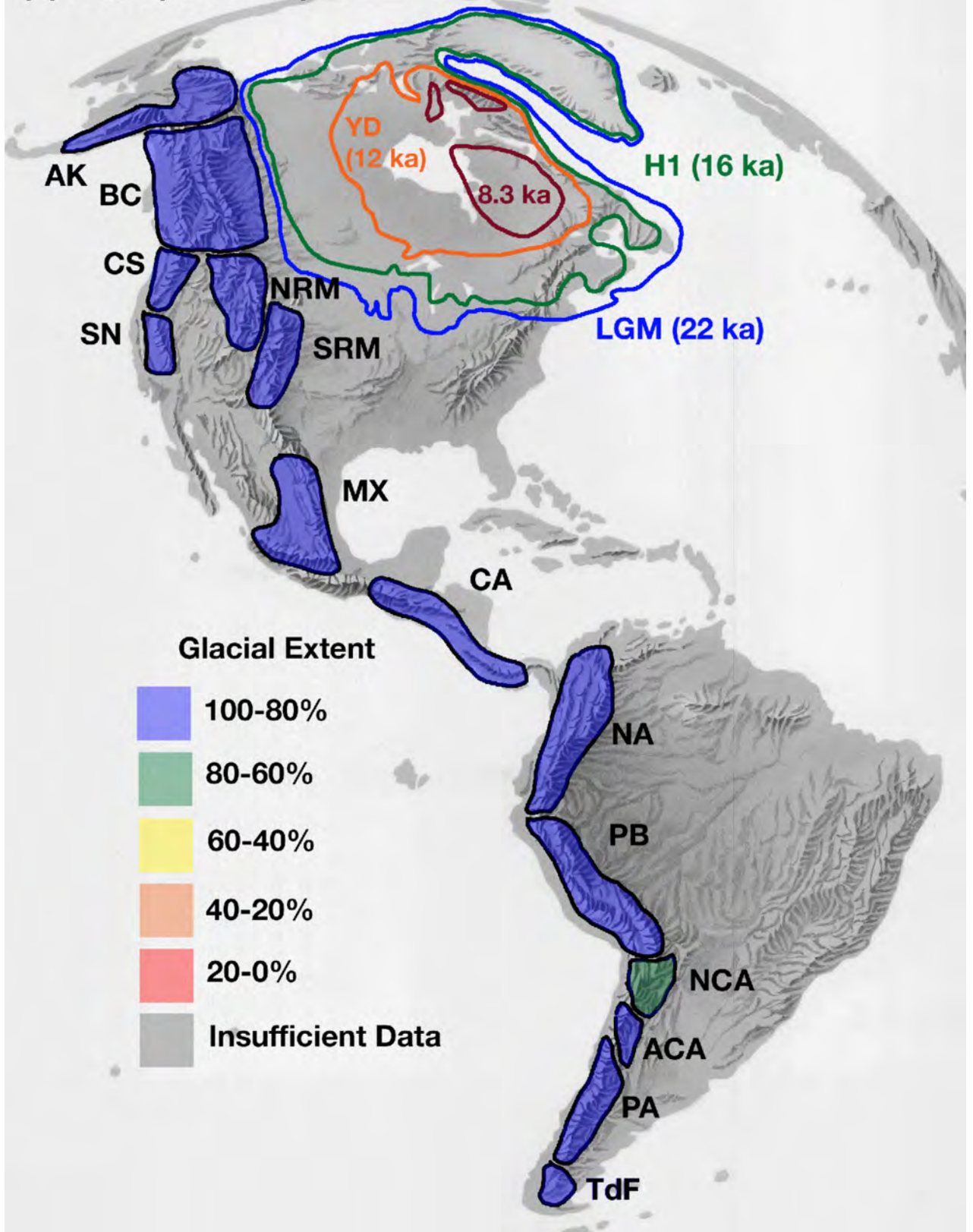
4673 Table 4. The main climate and glacial evolution features during the Younger Dryas (12.9-  
4674 11.7 ka) in the Americas.



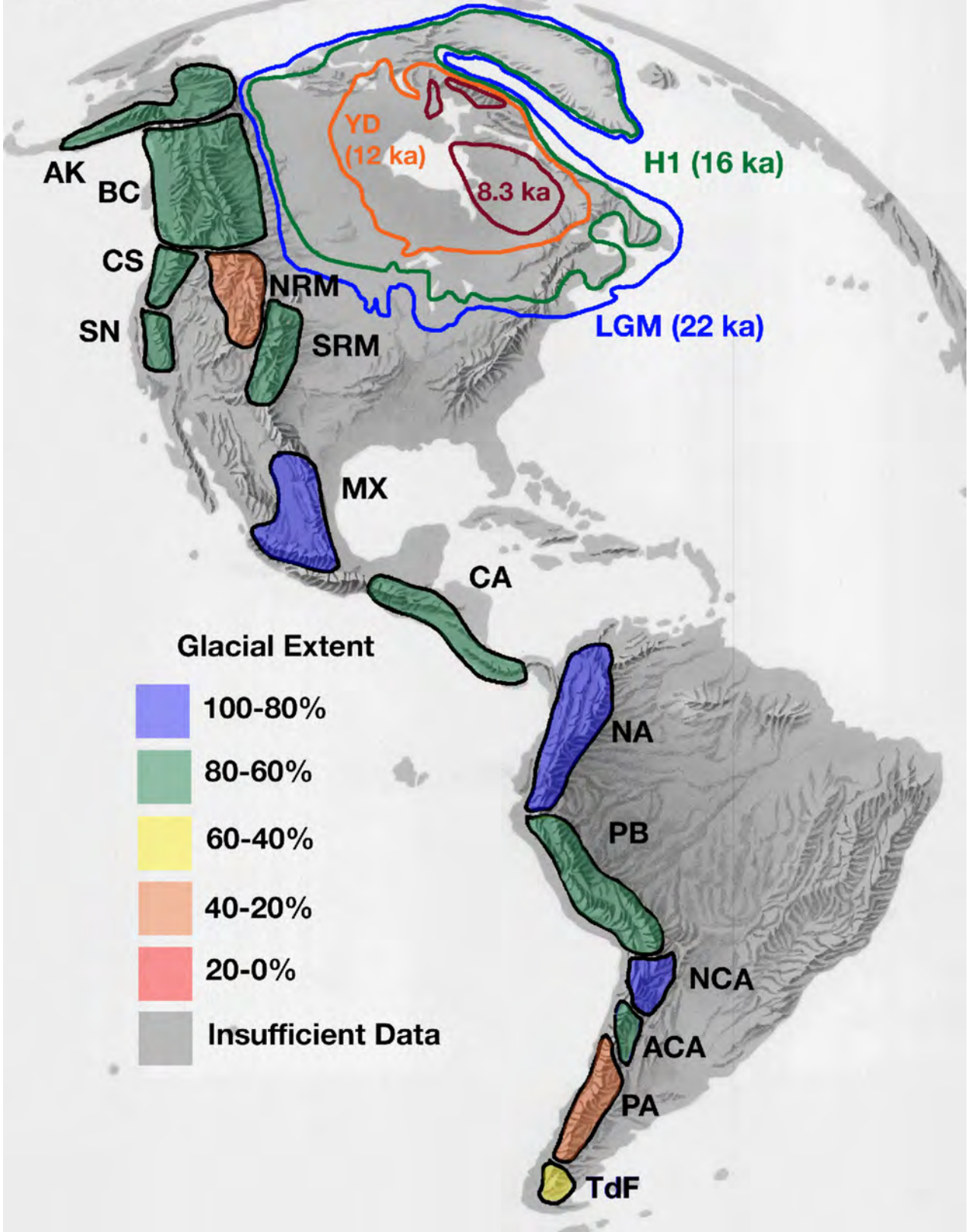




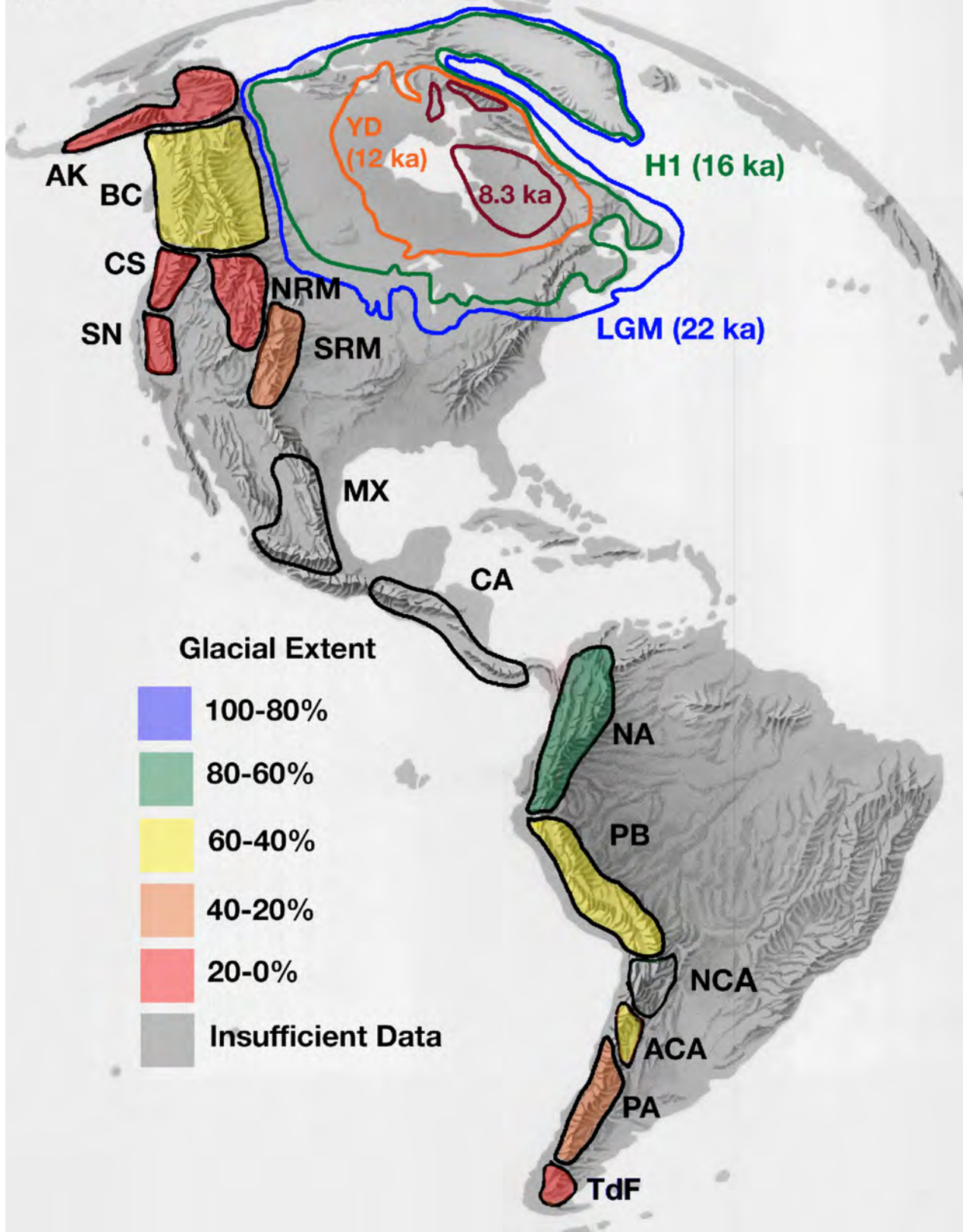
(a) LGM (26-21 ka)



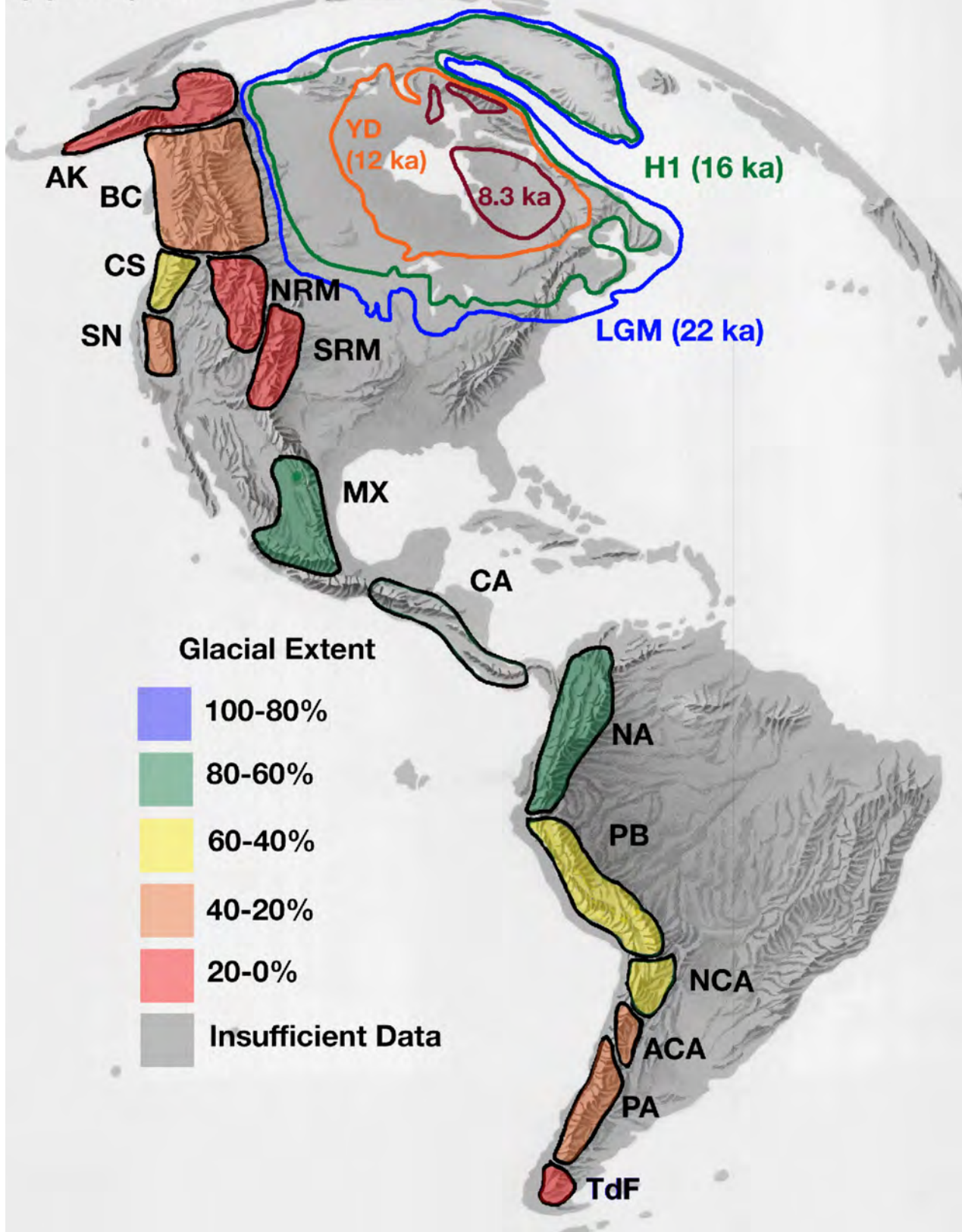
(b) H1 (16.8-14.7 ka)



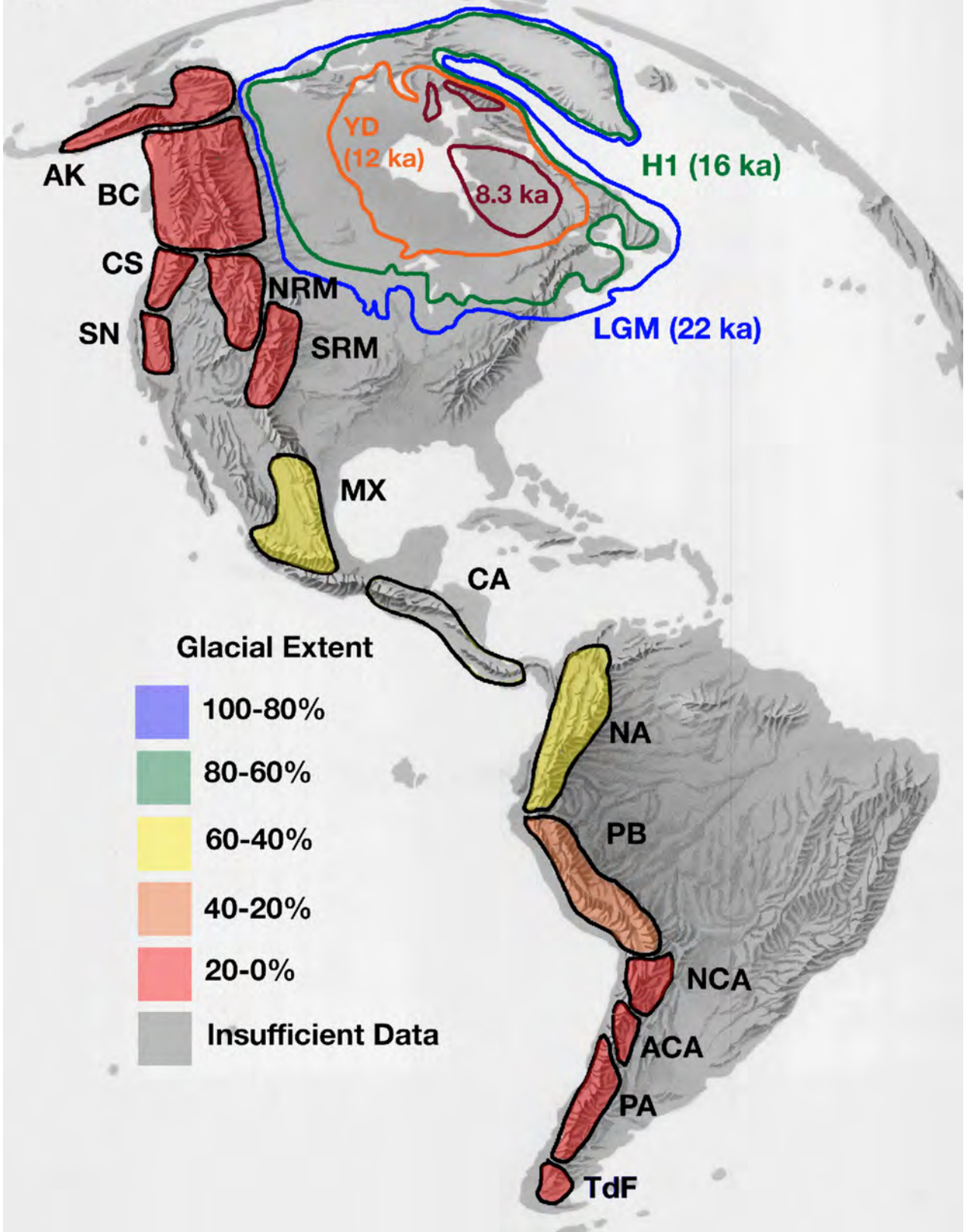
(c) Bølling/Allerød-ACR (14.7-12.9 ka)

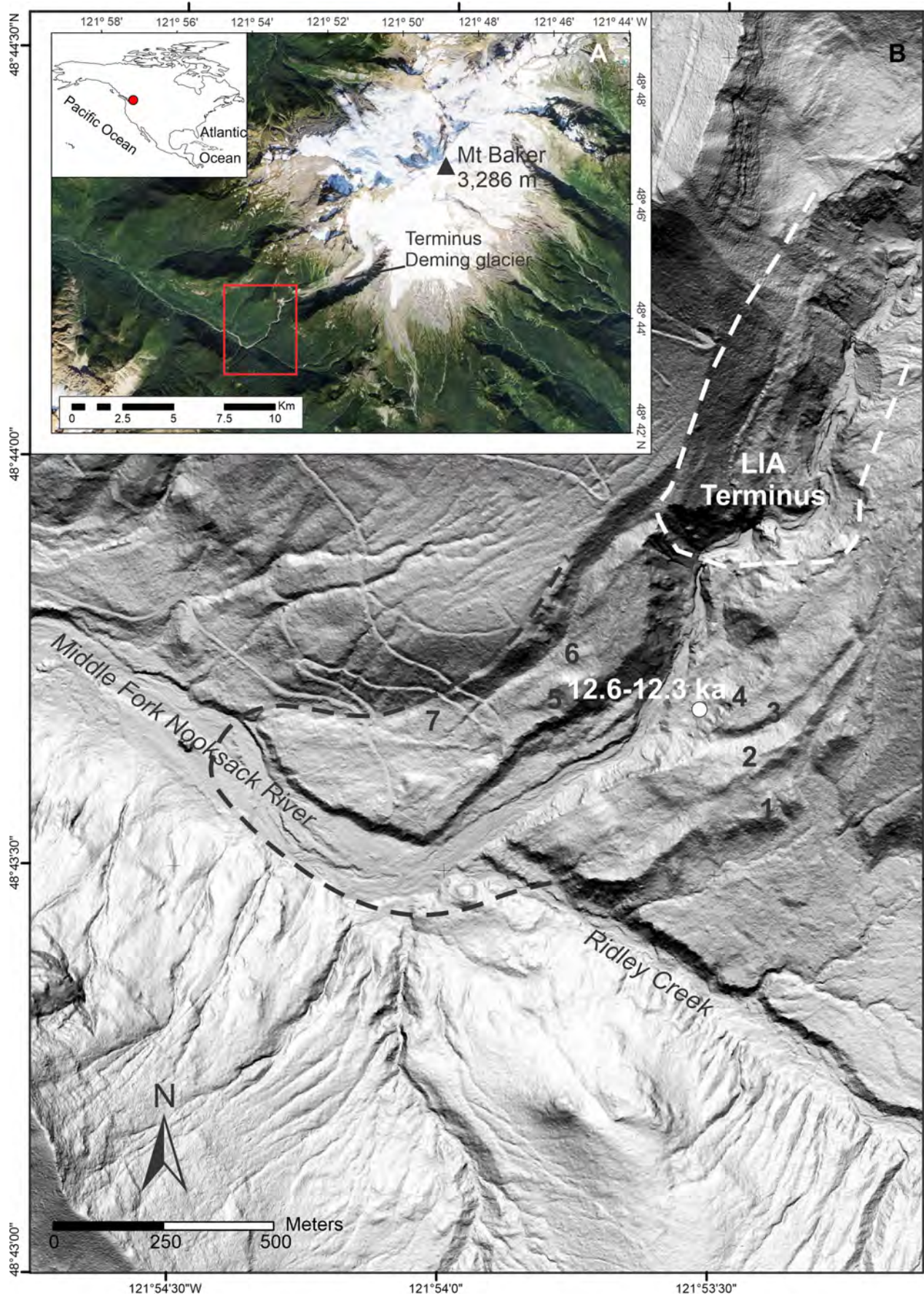


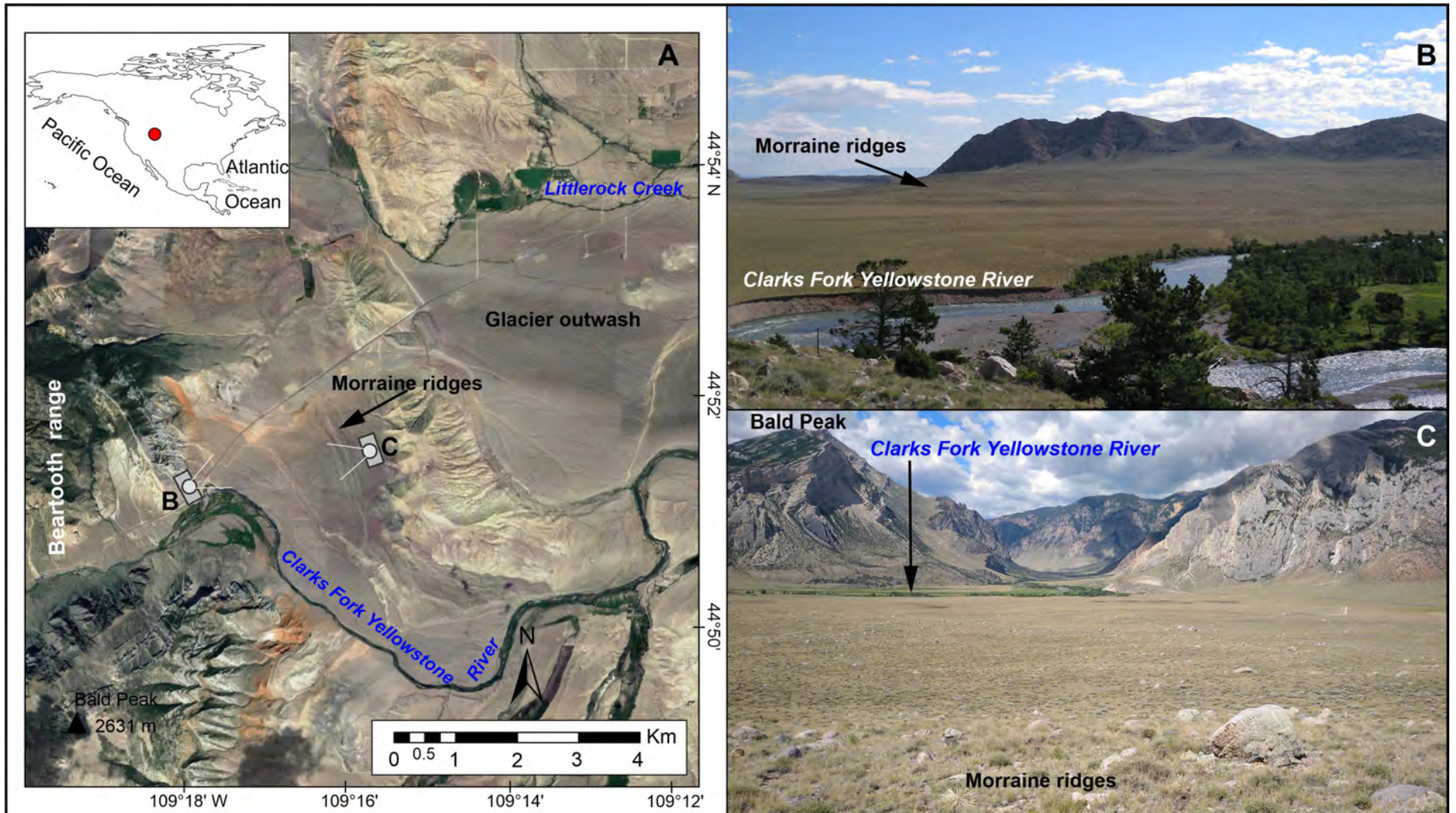
(d) YD (12.9-11.7 ka)

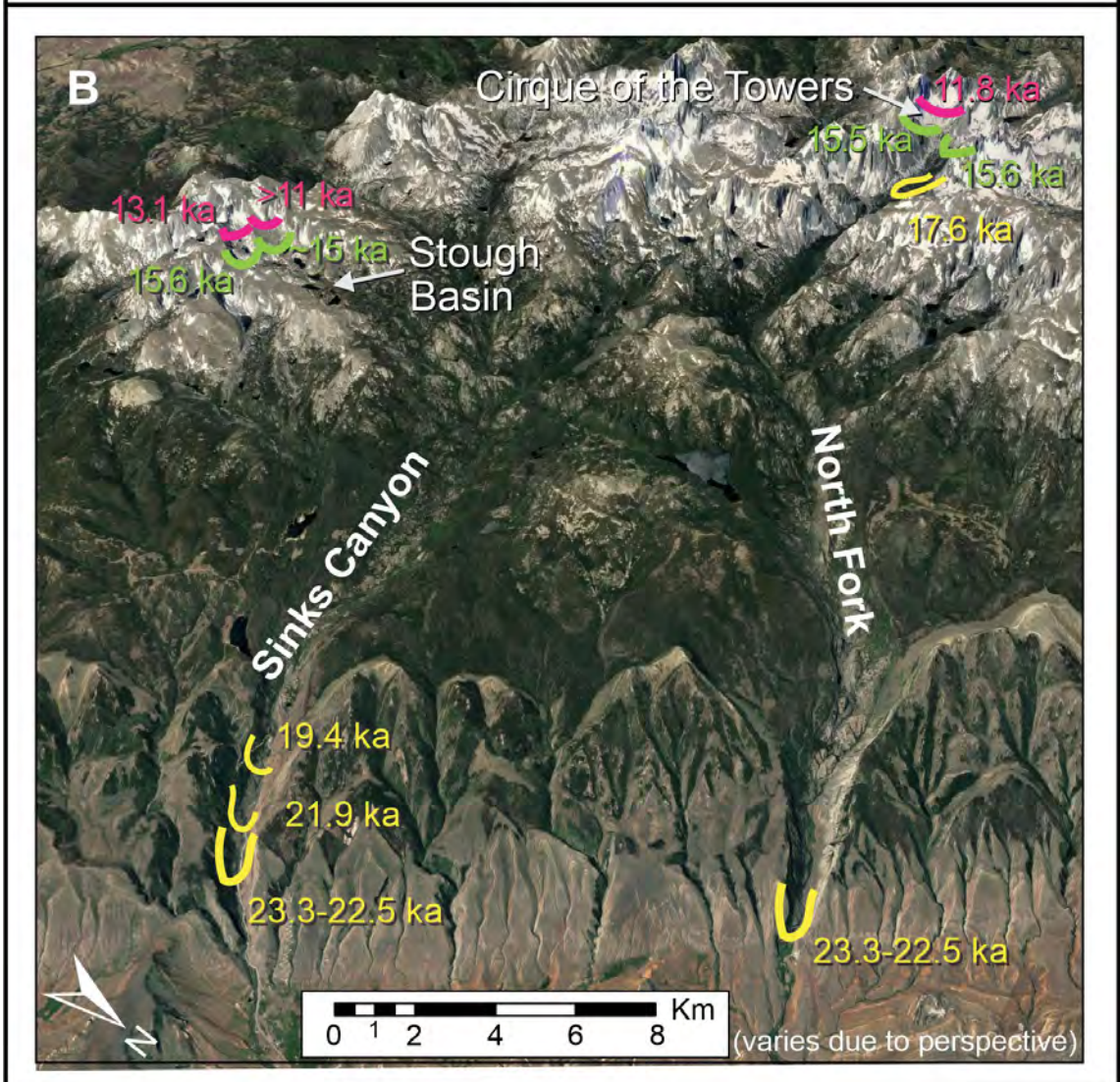
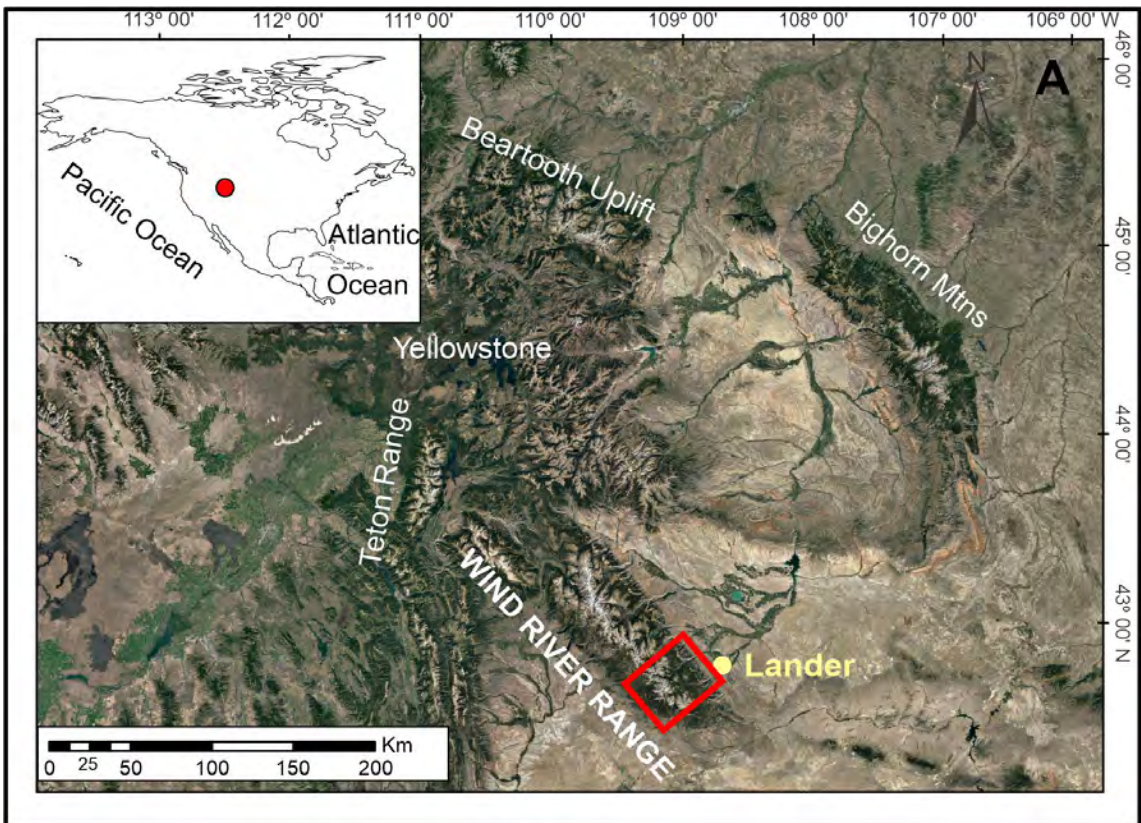


(e) Early Holocene (11.7-7.0 ka)

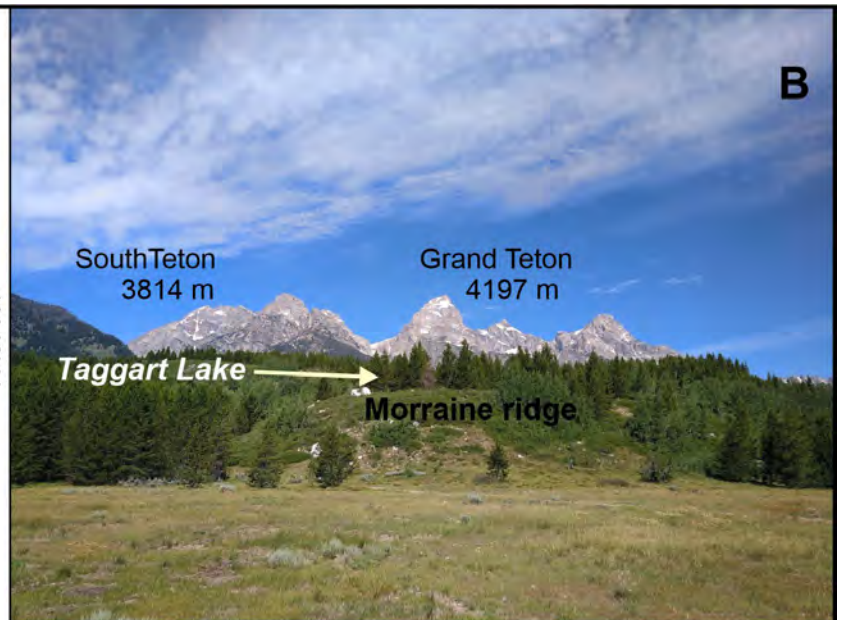
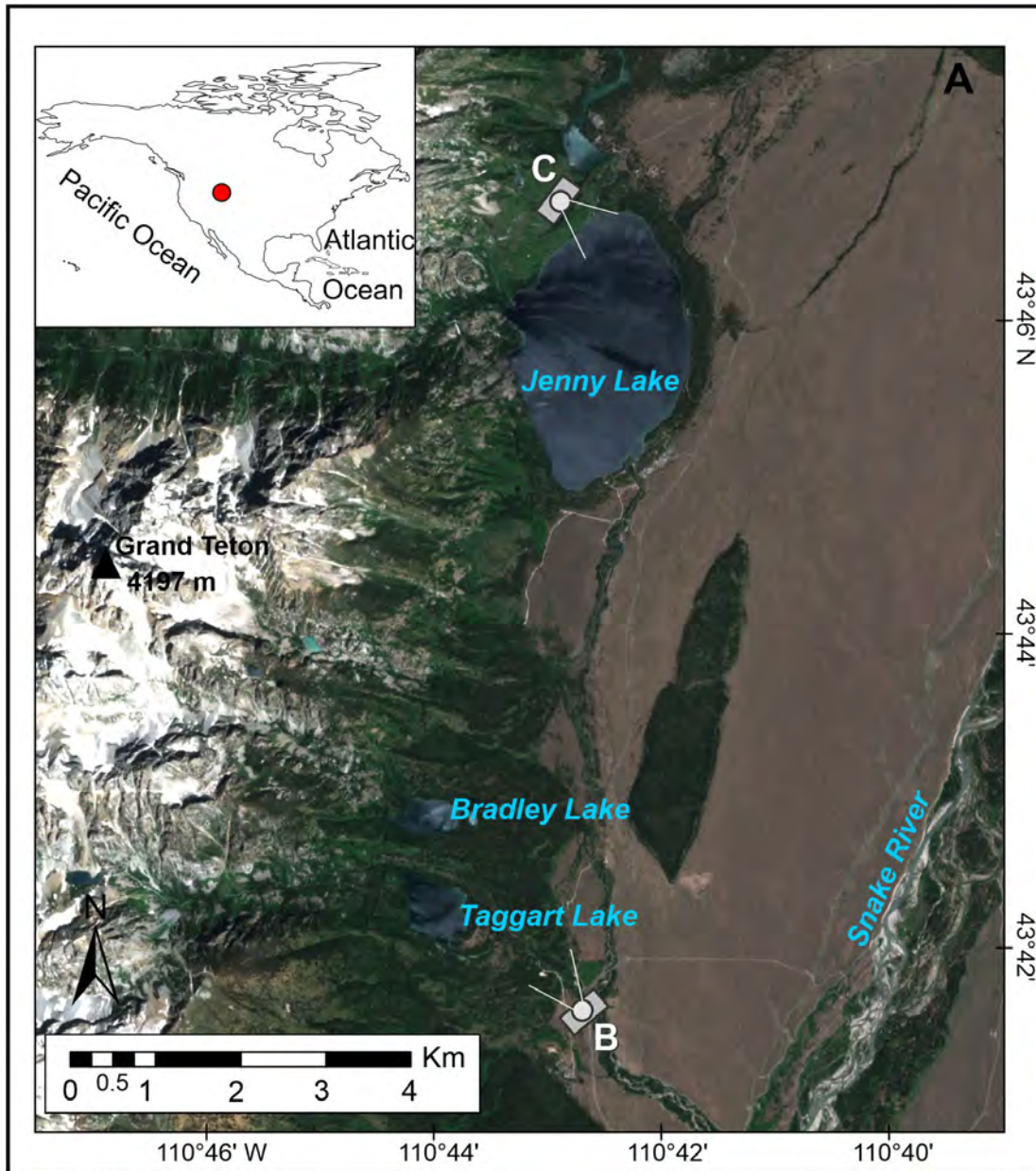


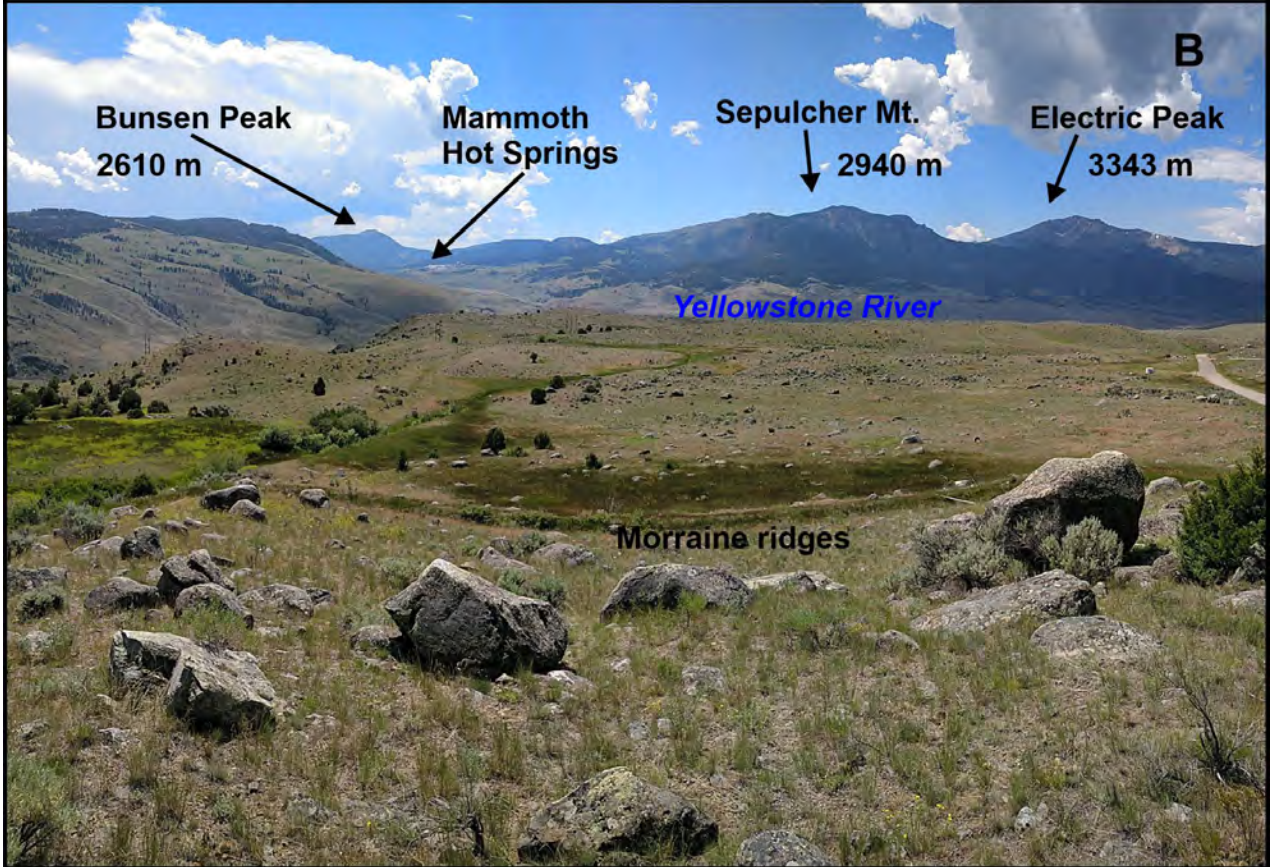
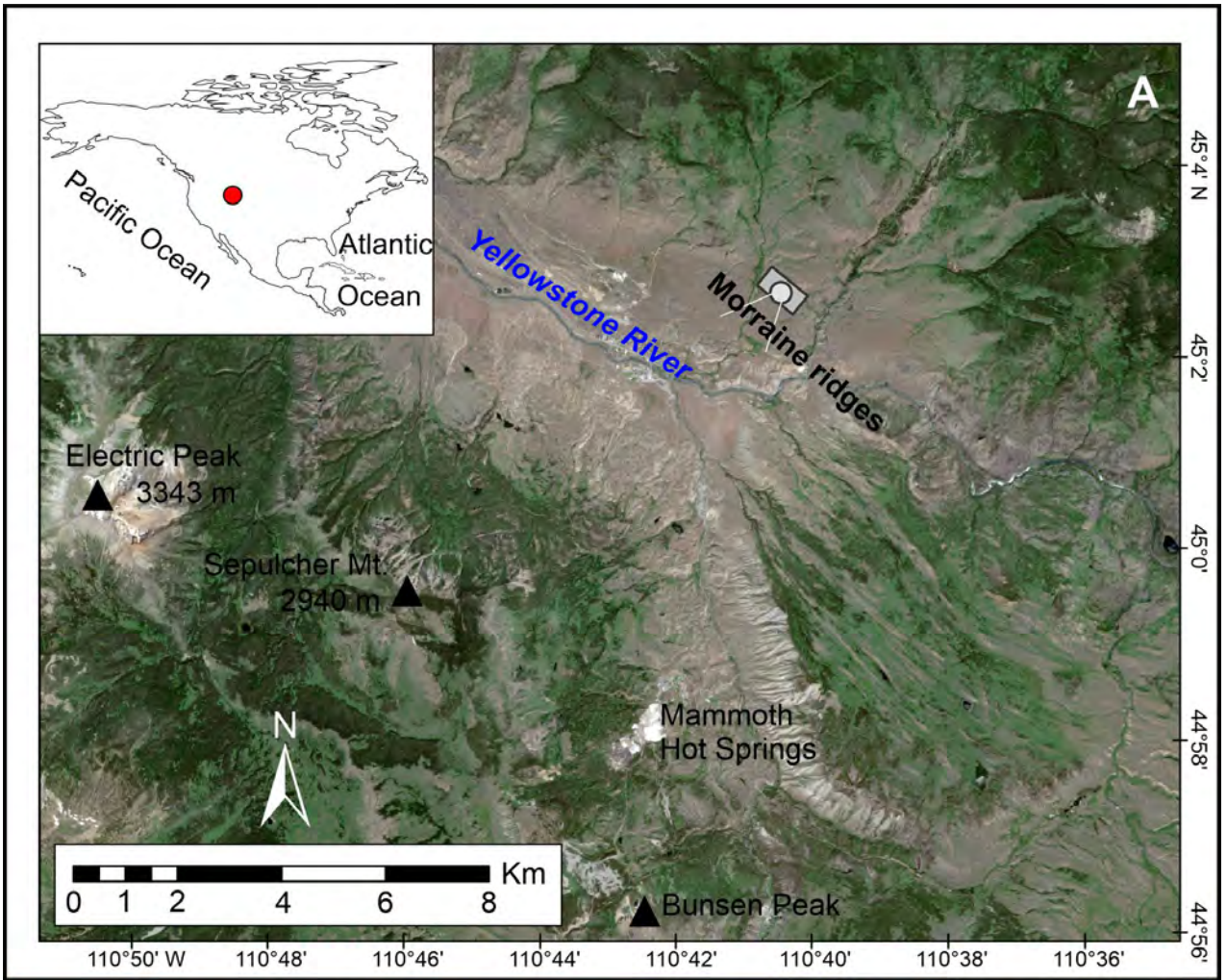


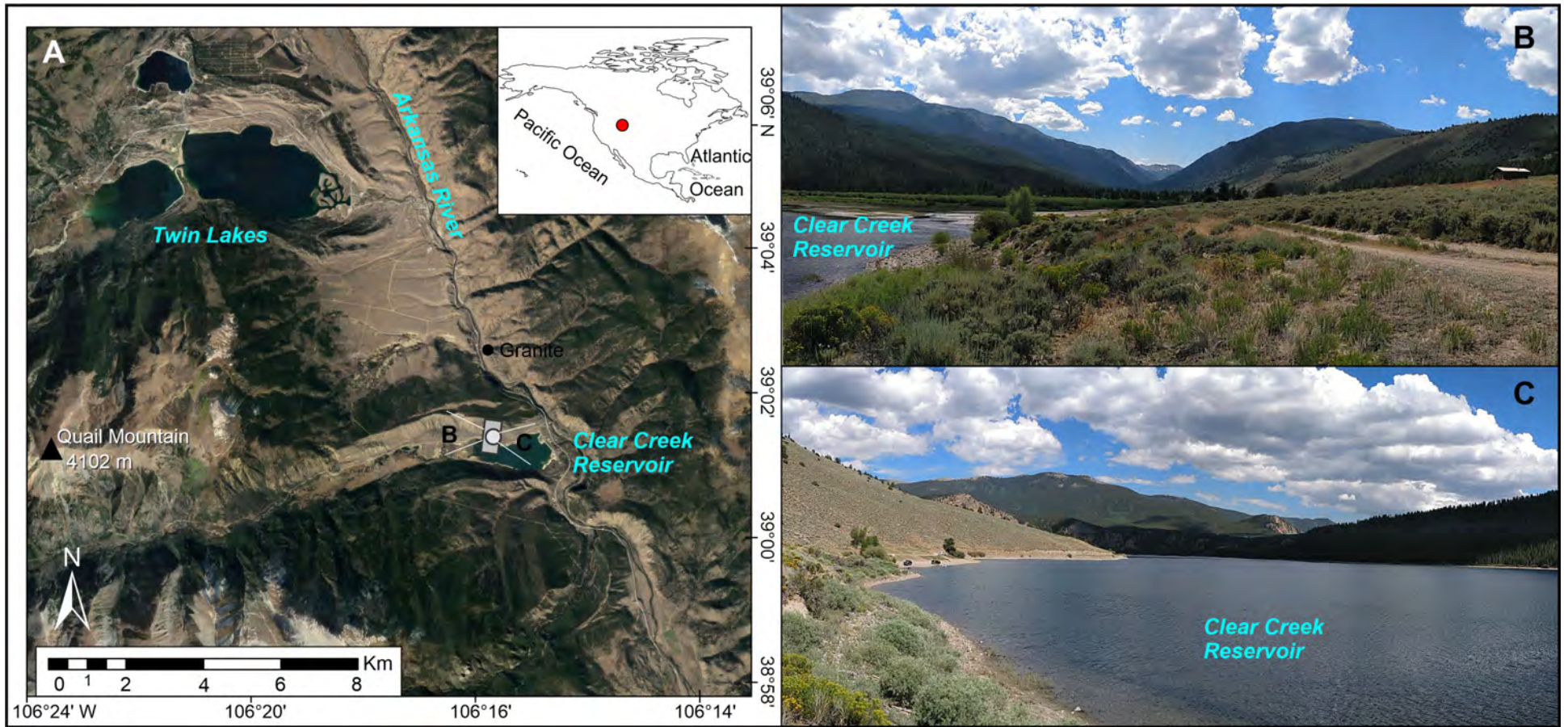


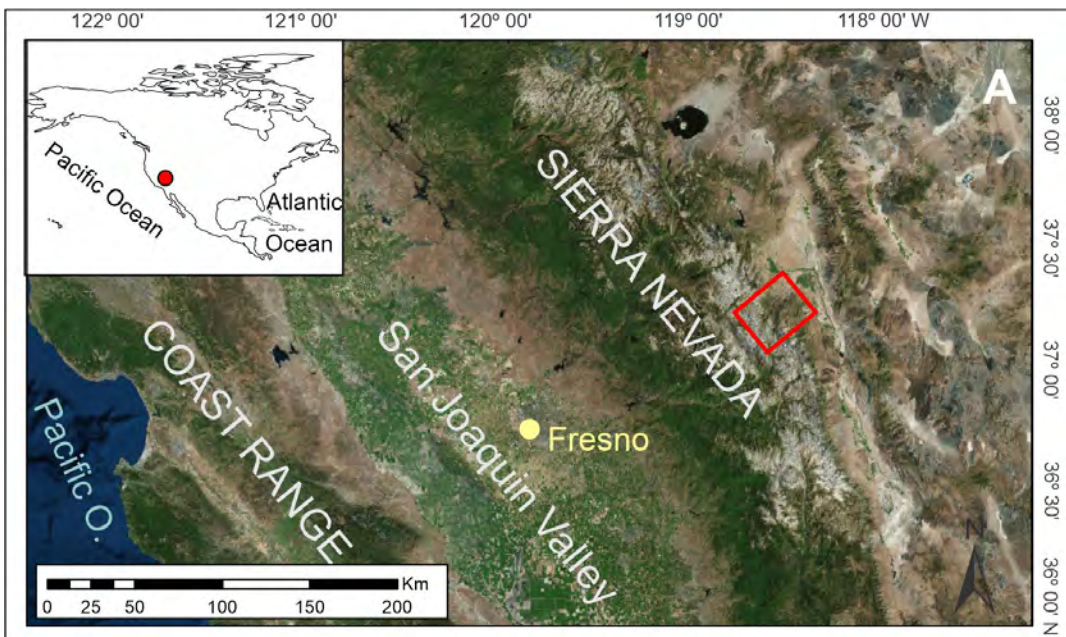


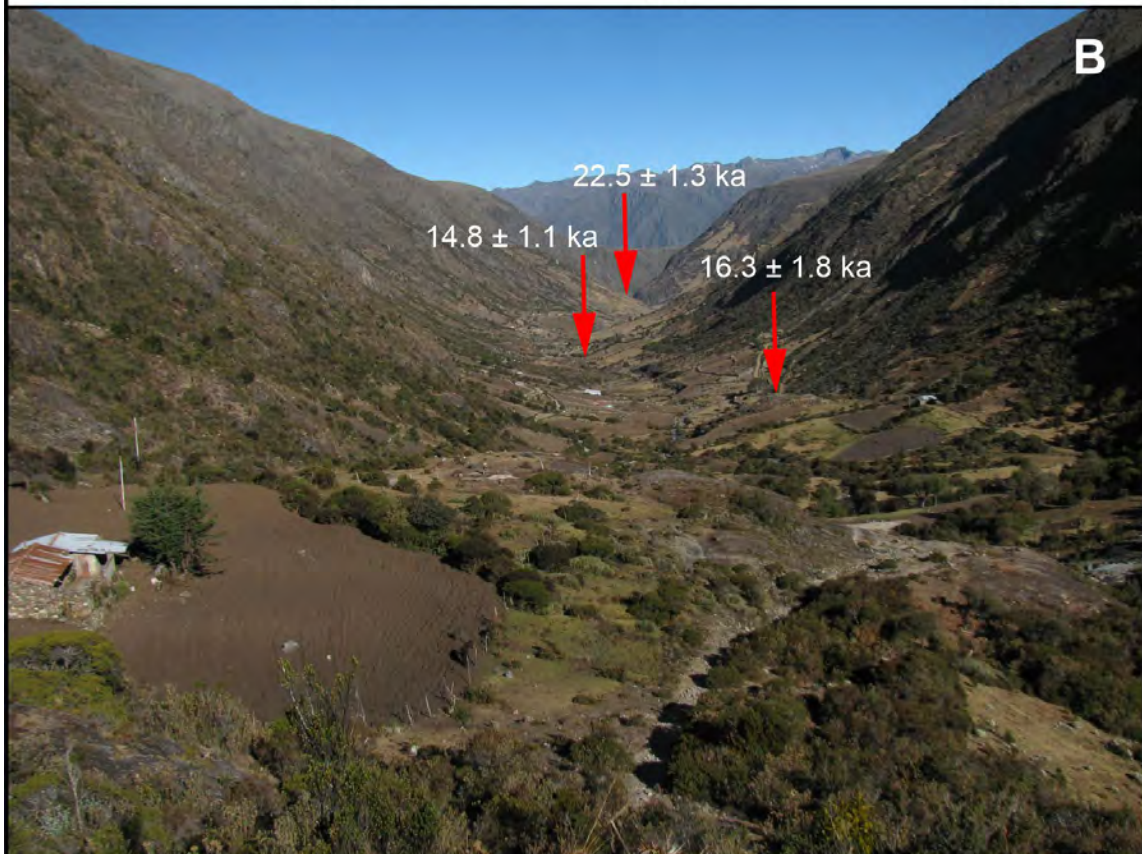
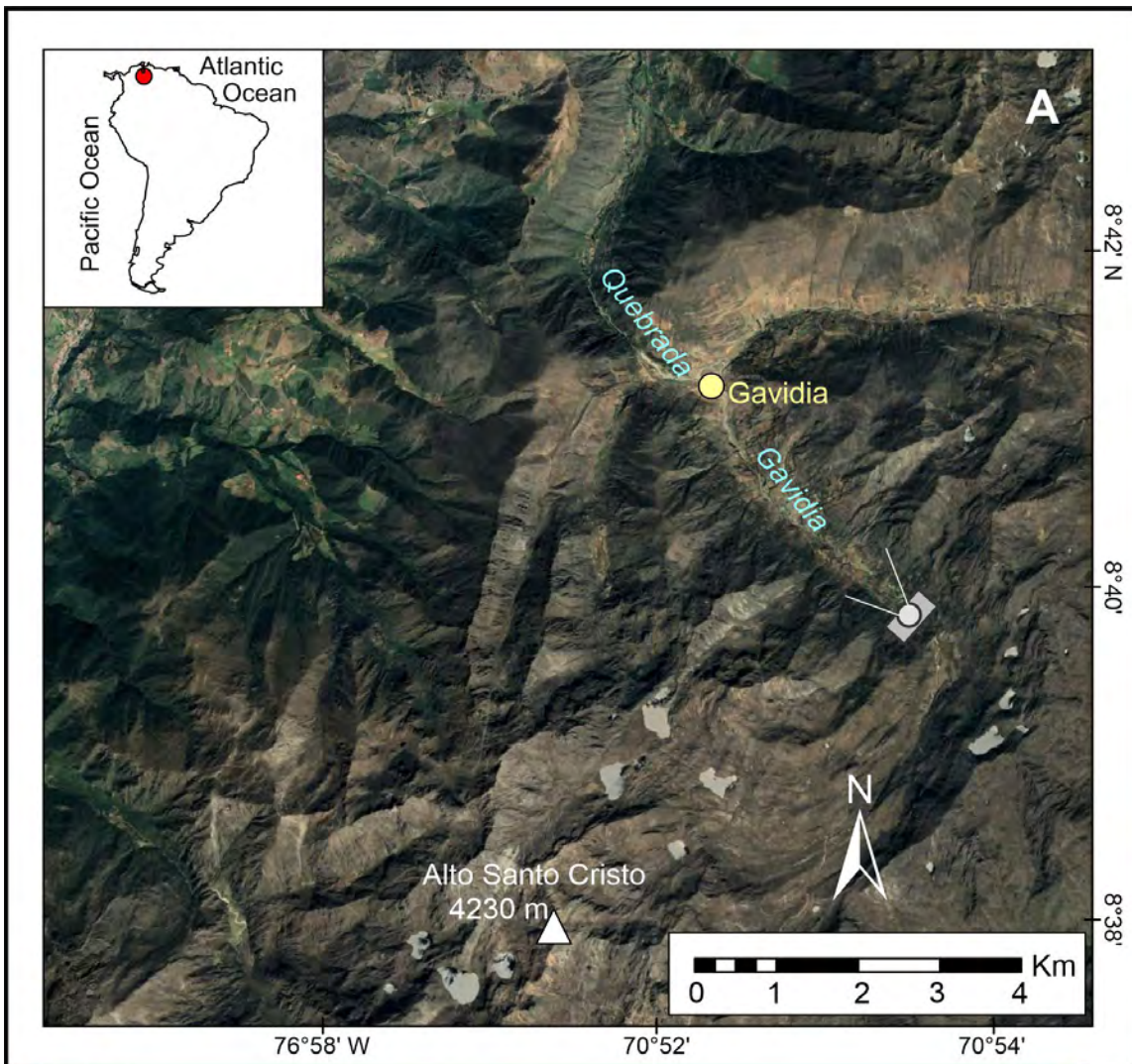


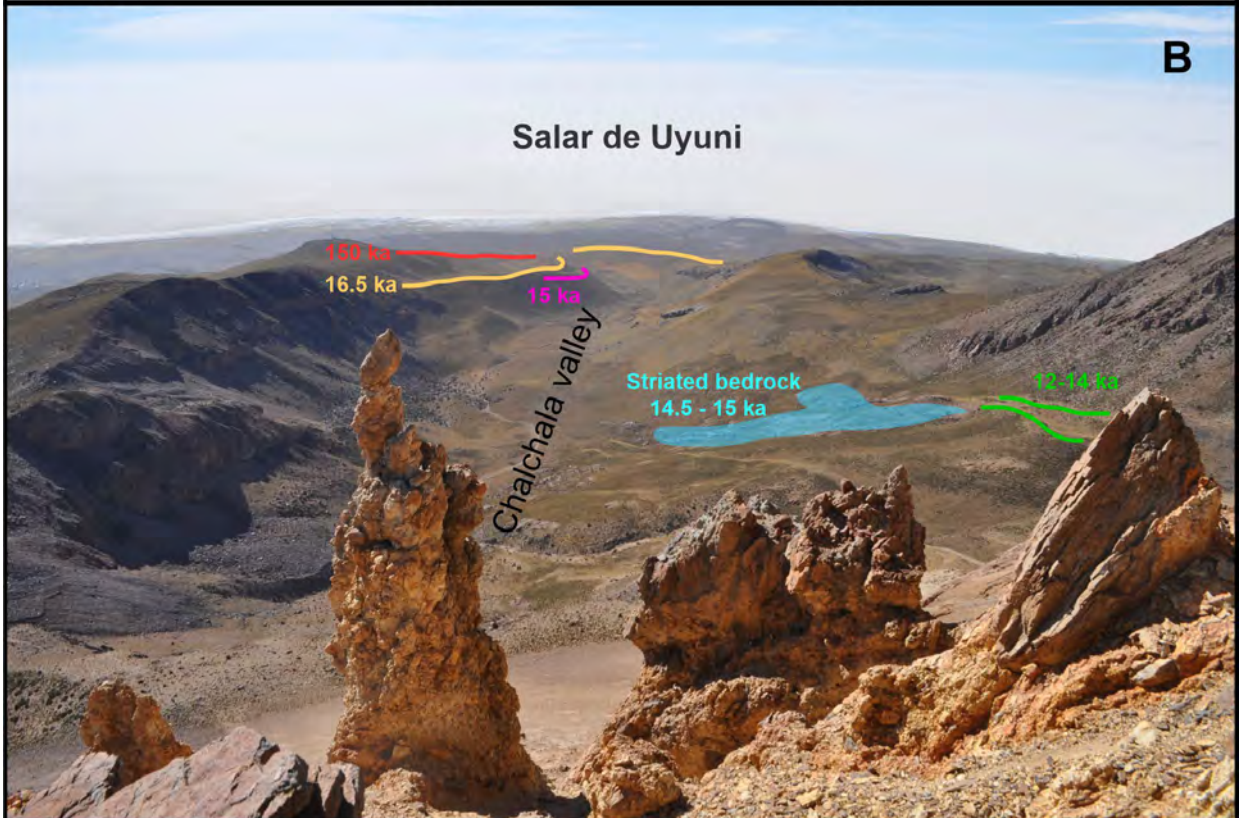
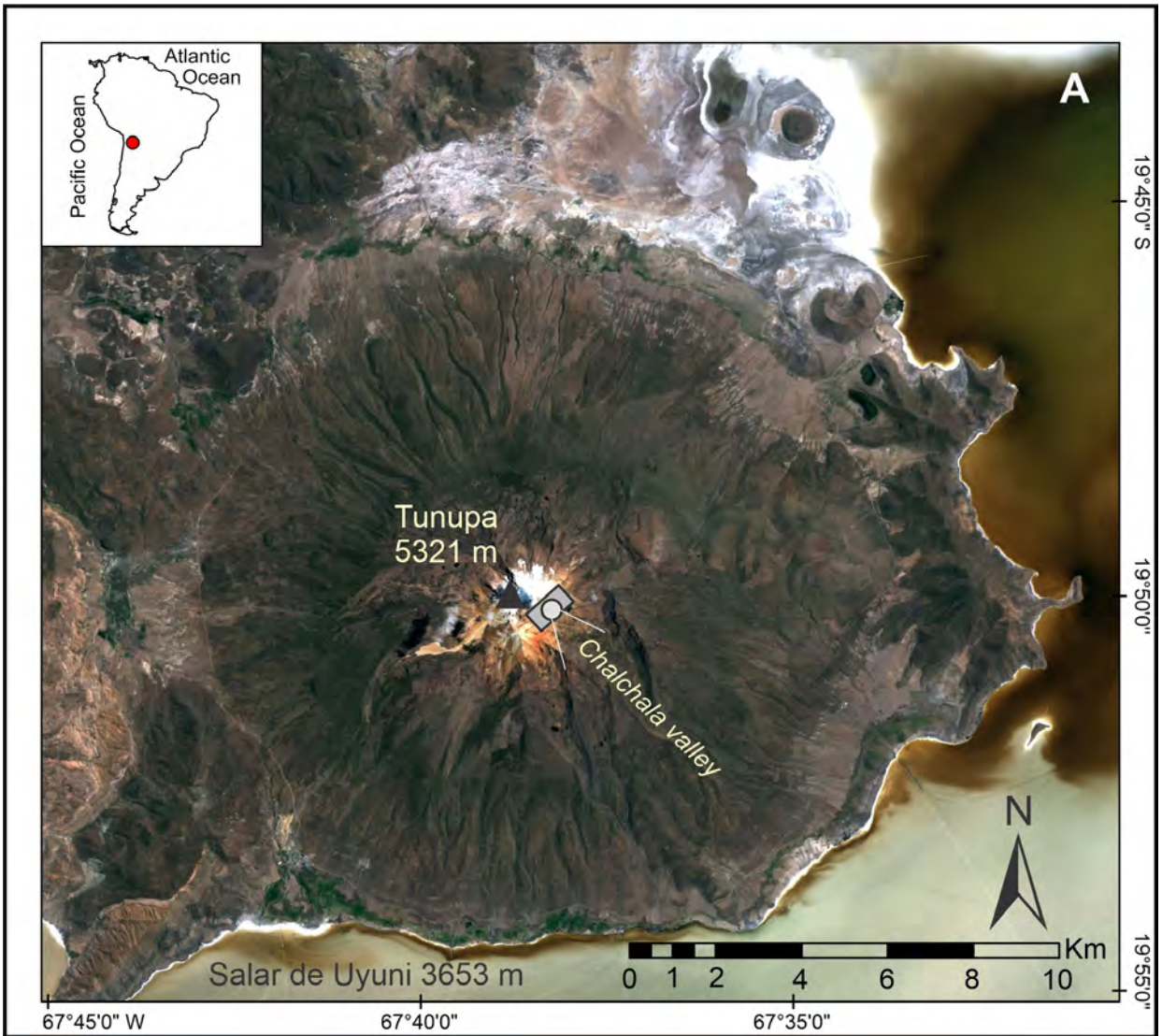


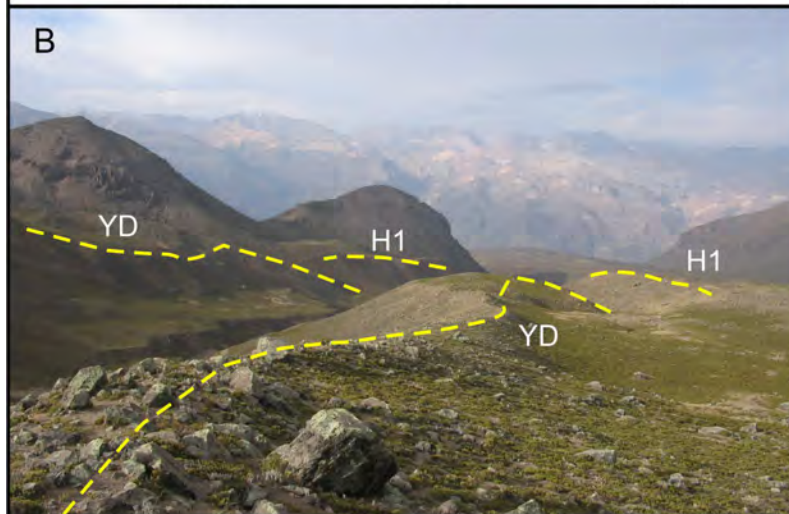
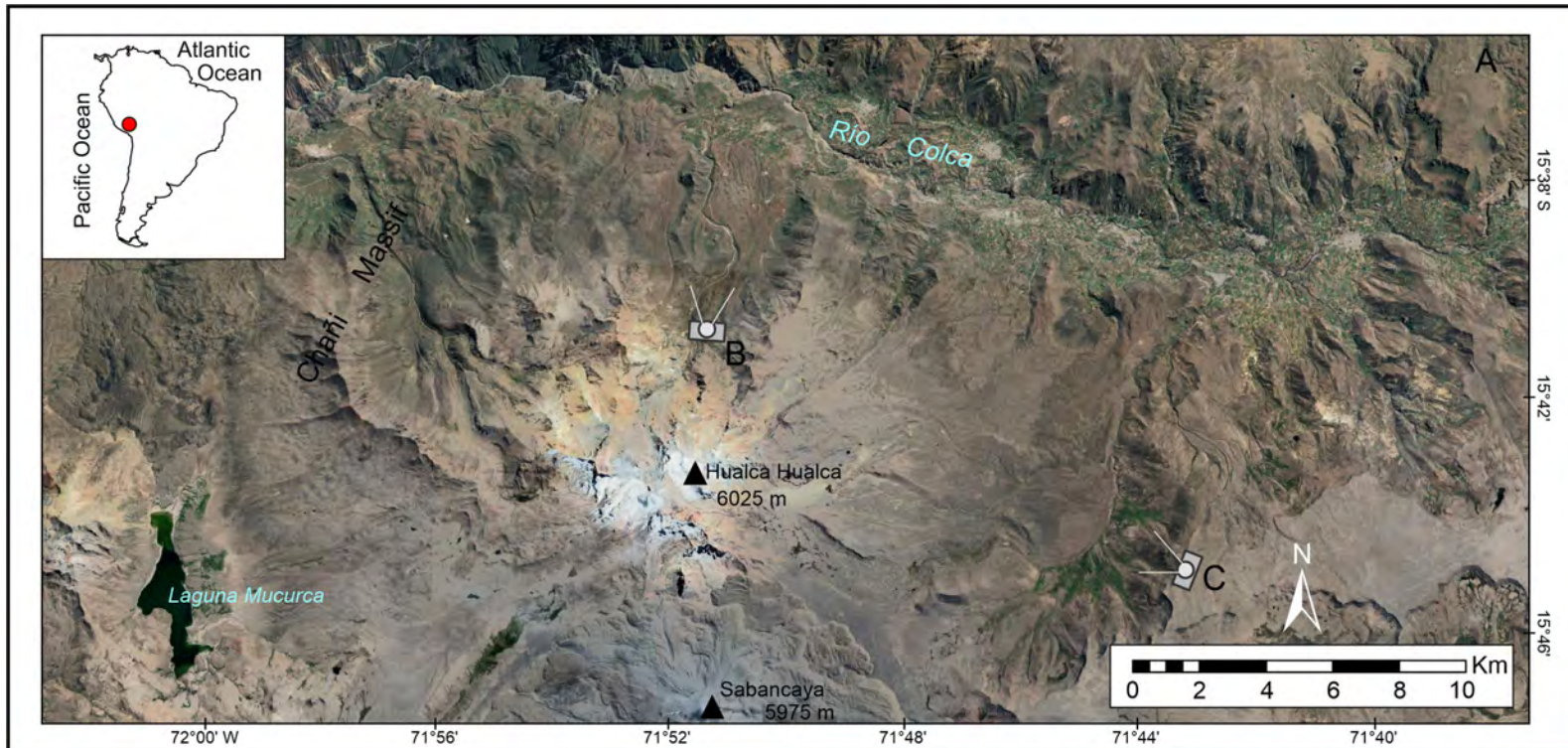


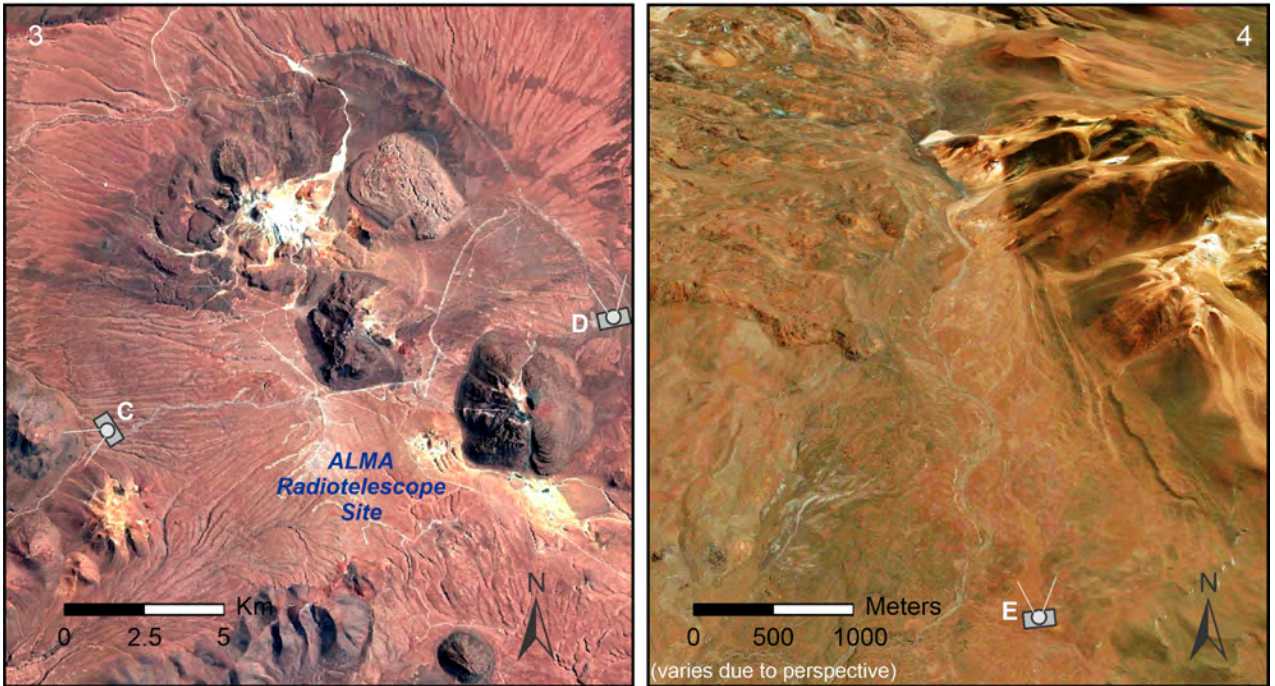
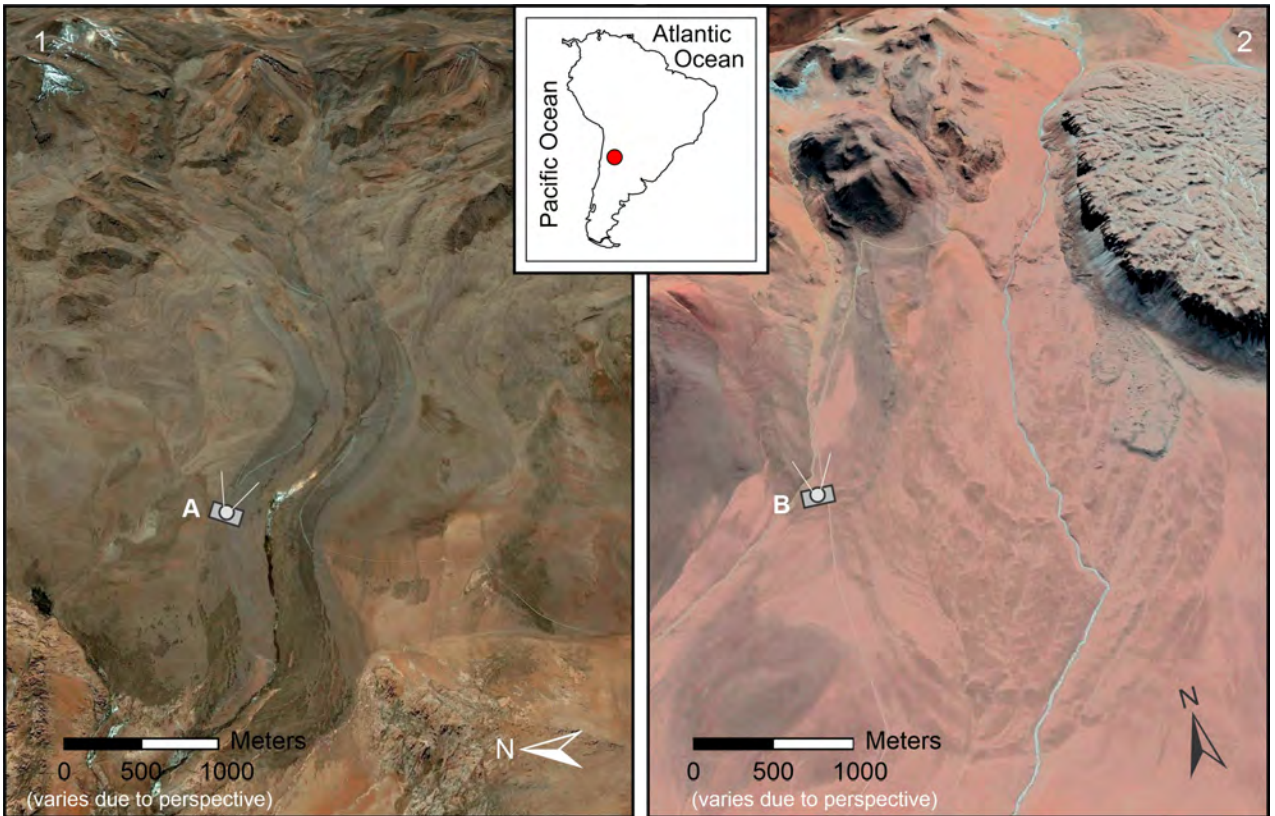




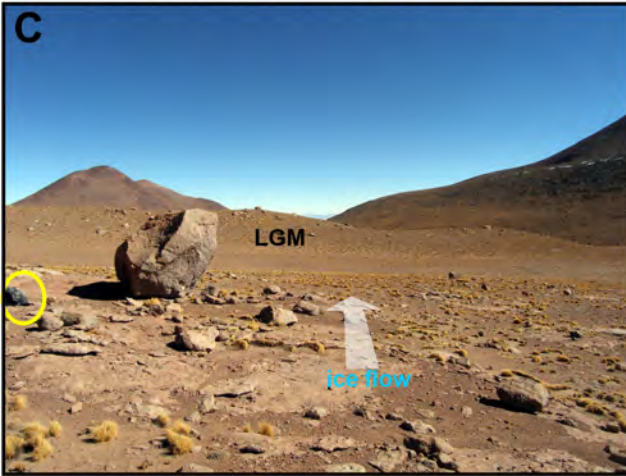
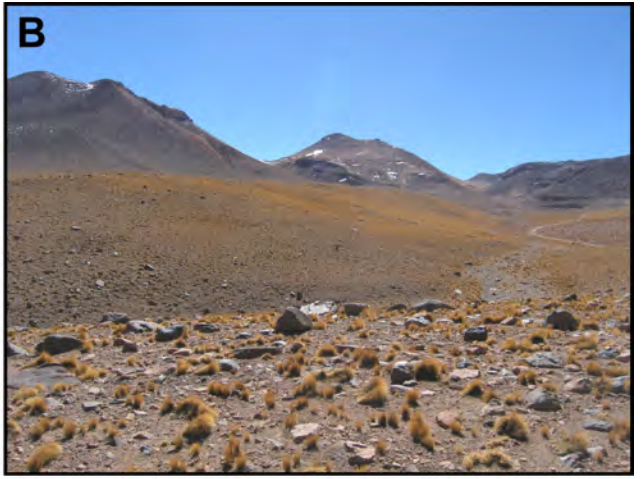


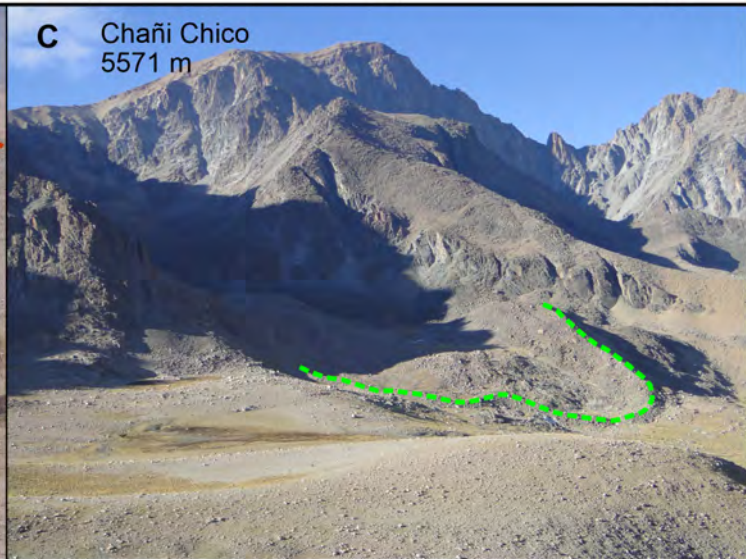
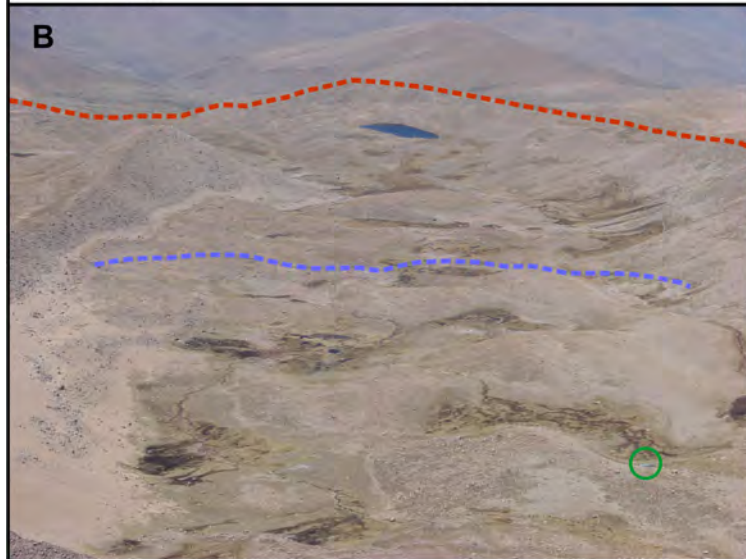
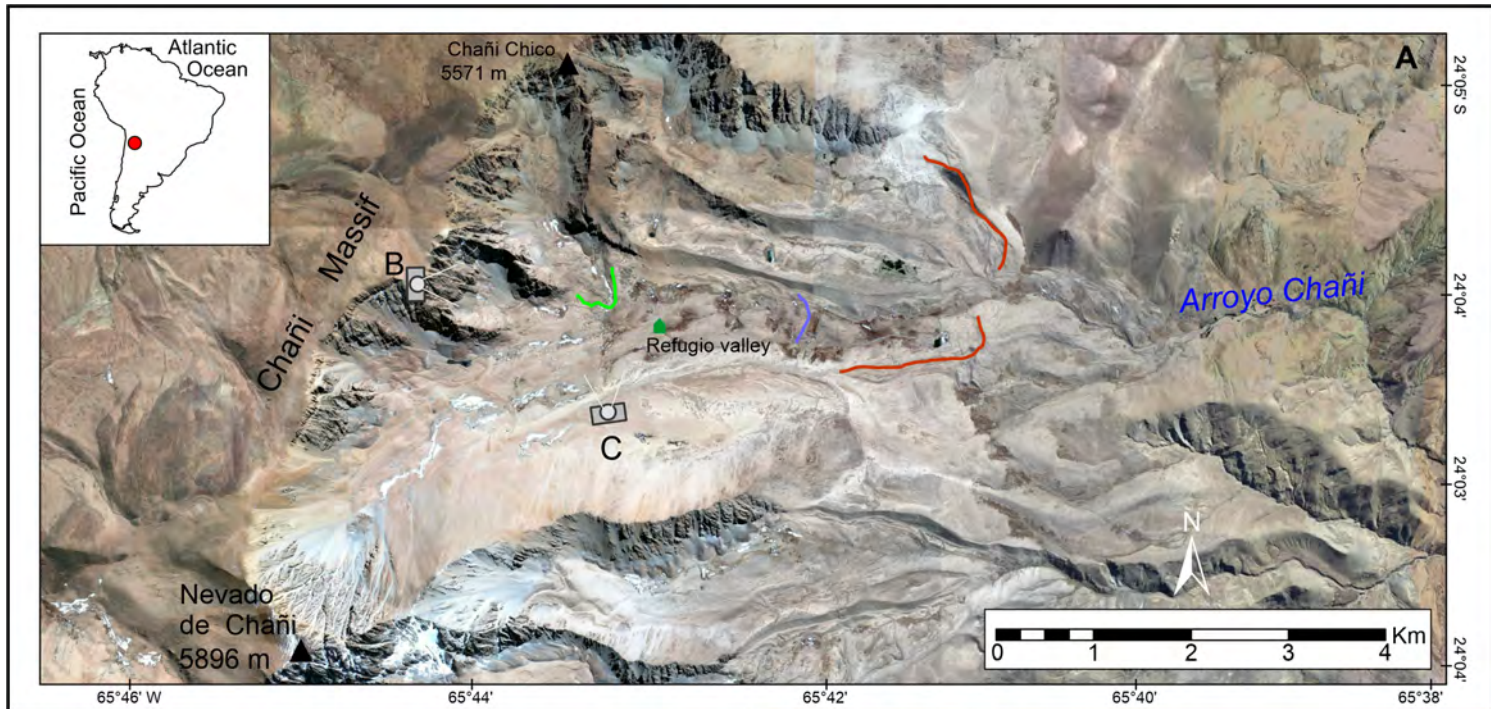


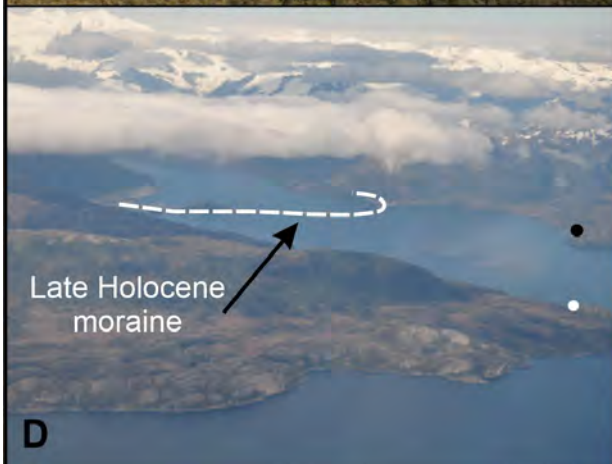
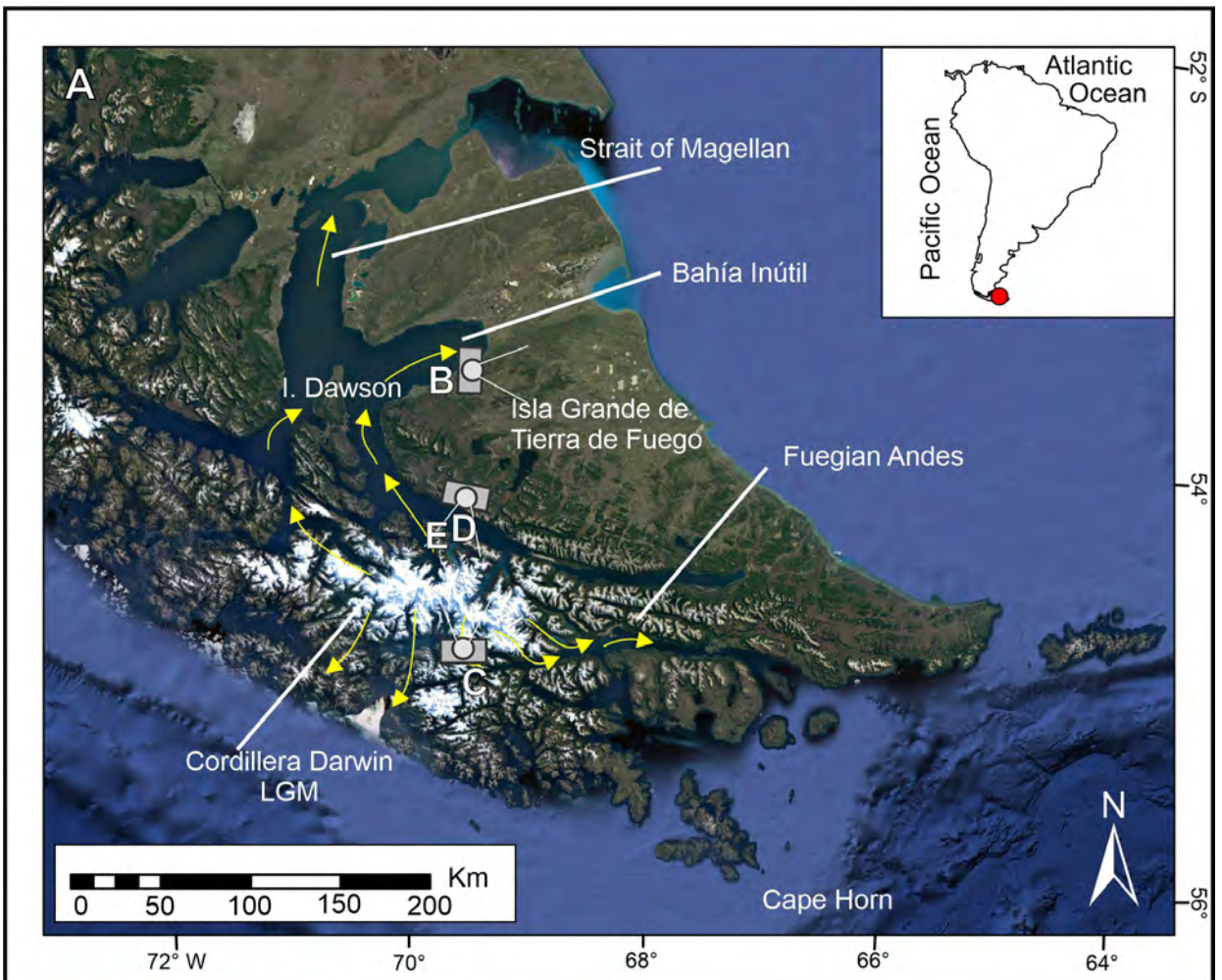


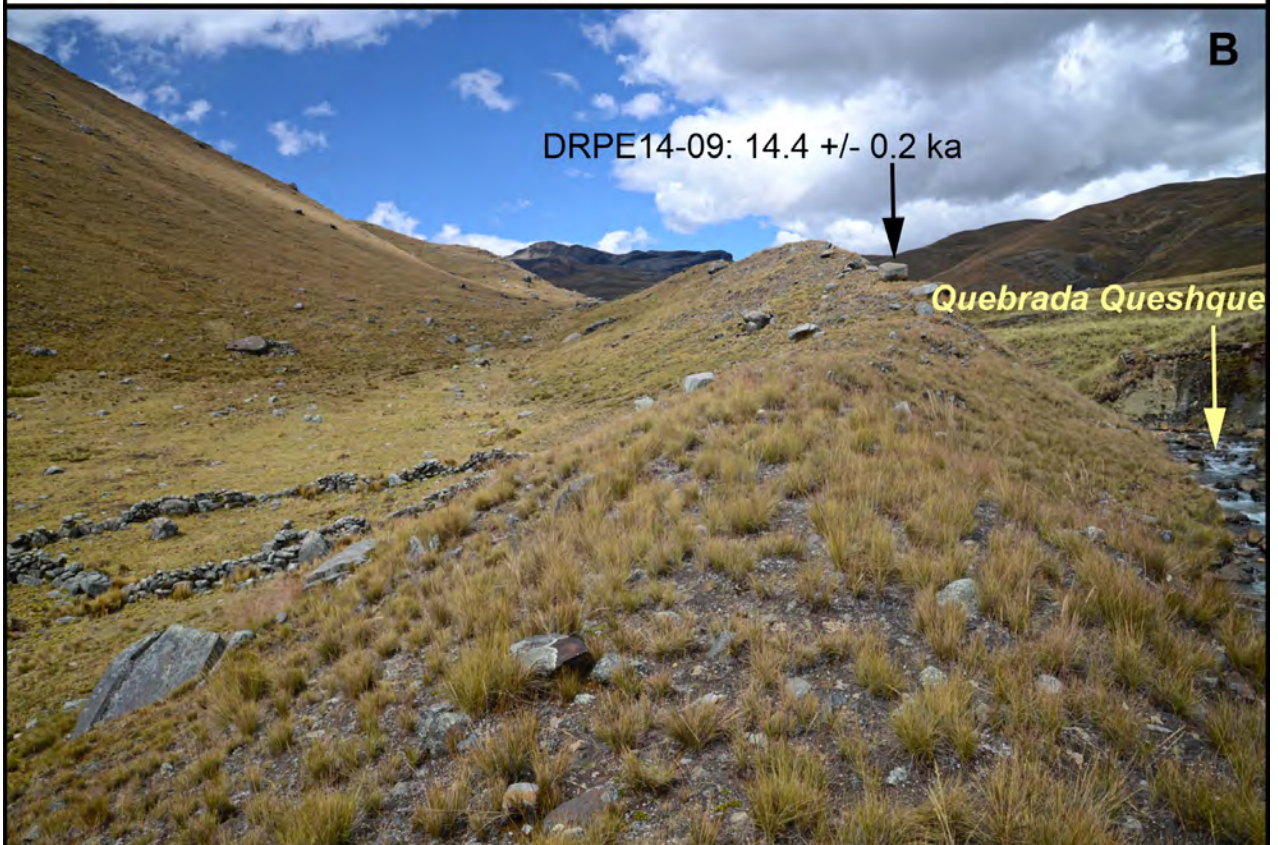
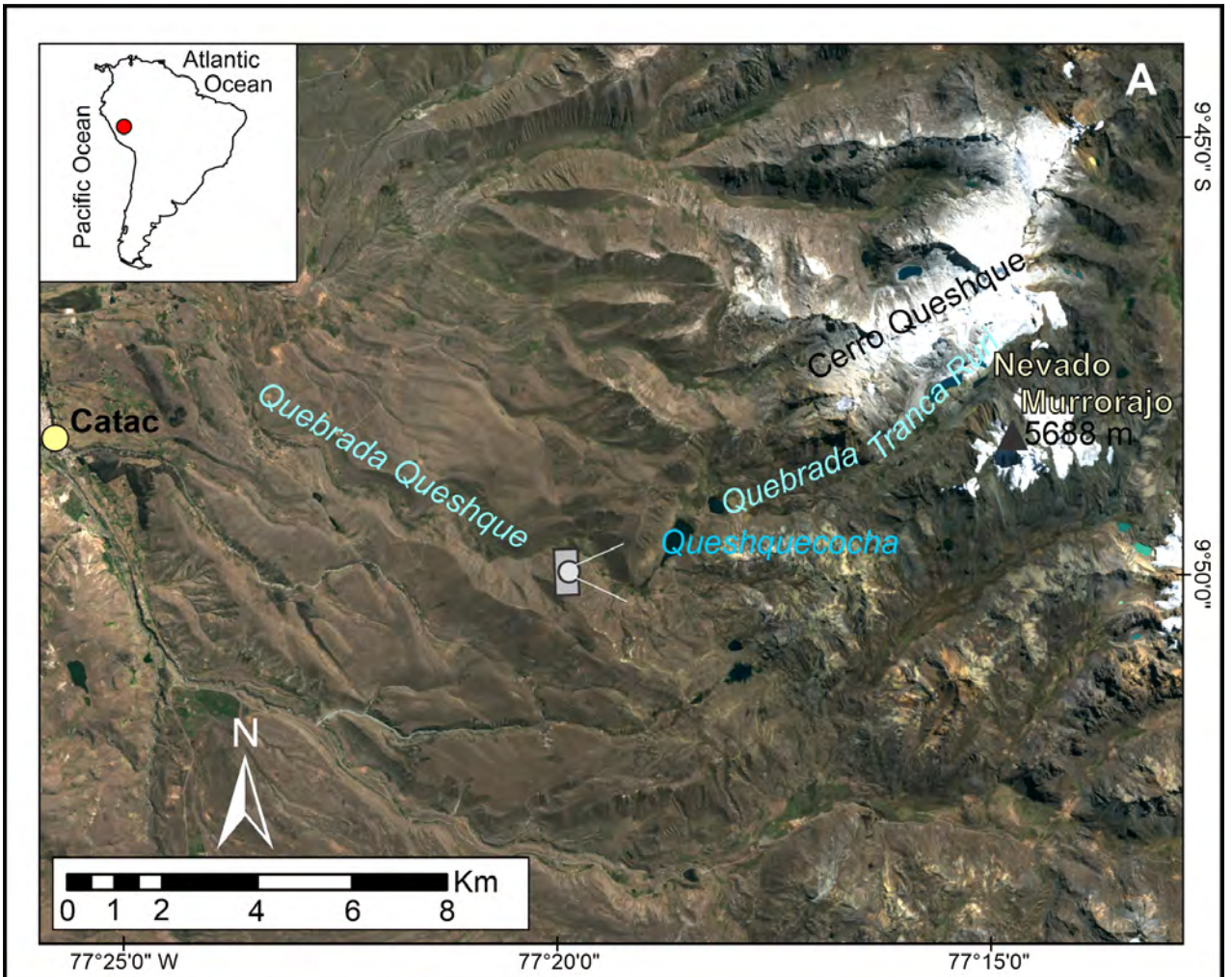


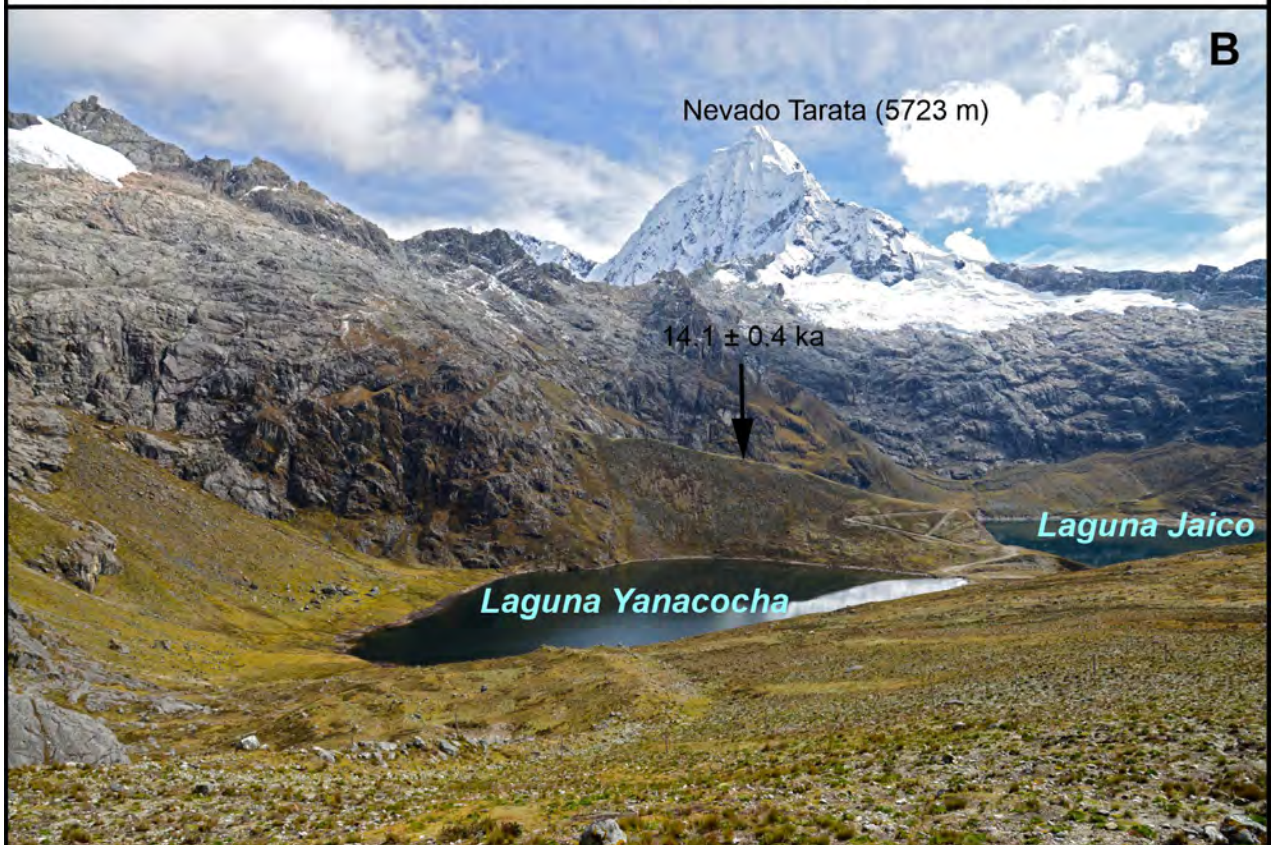
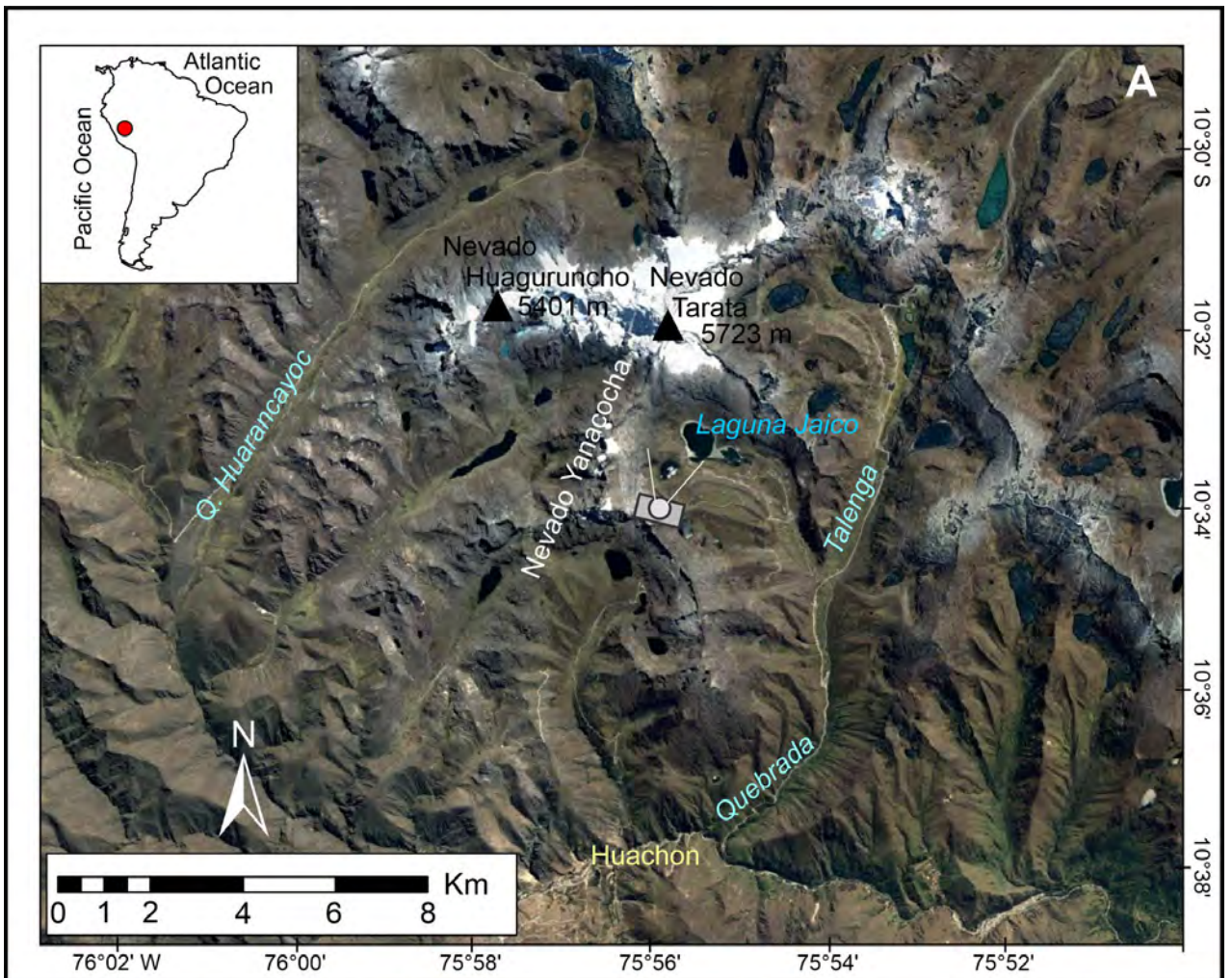


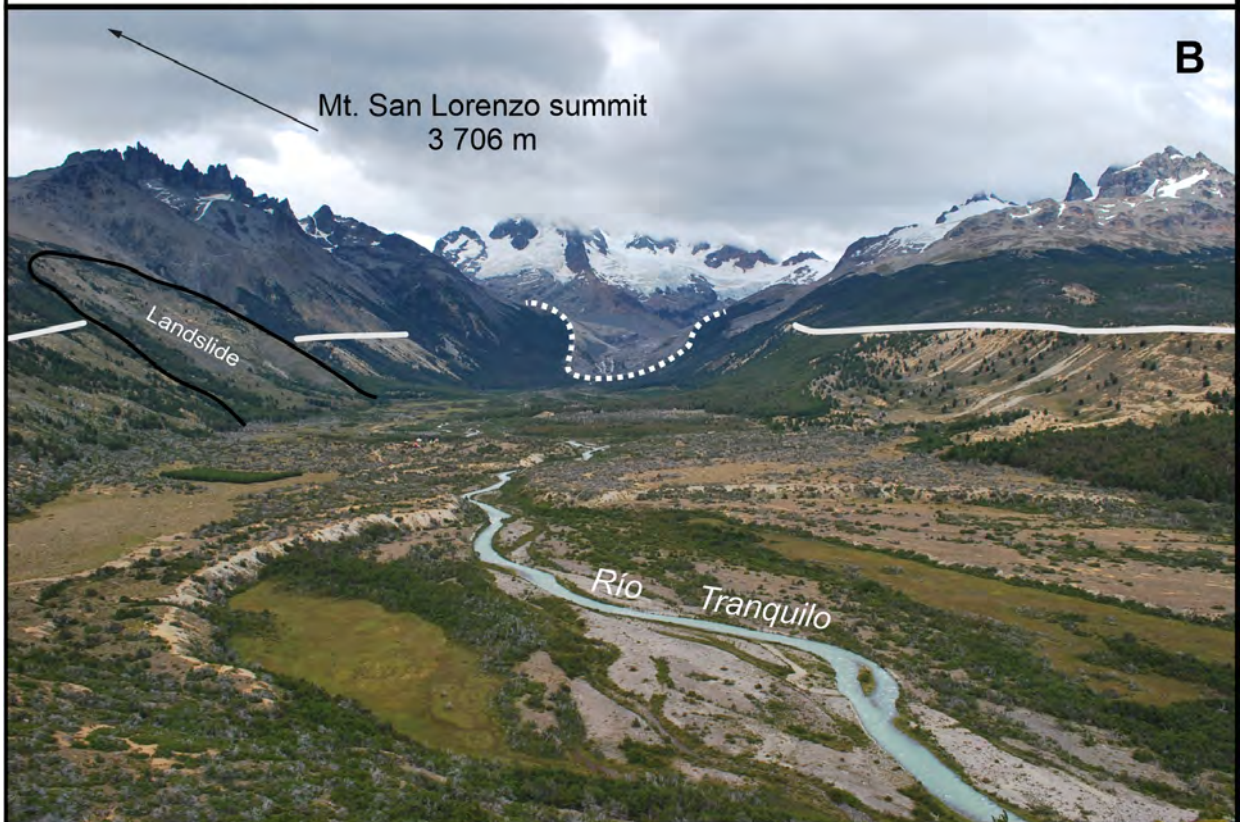
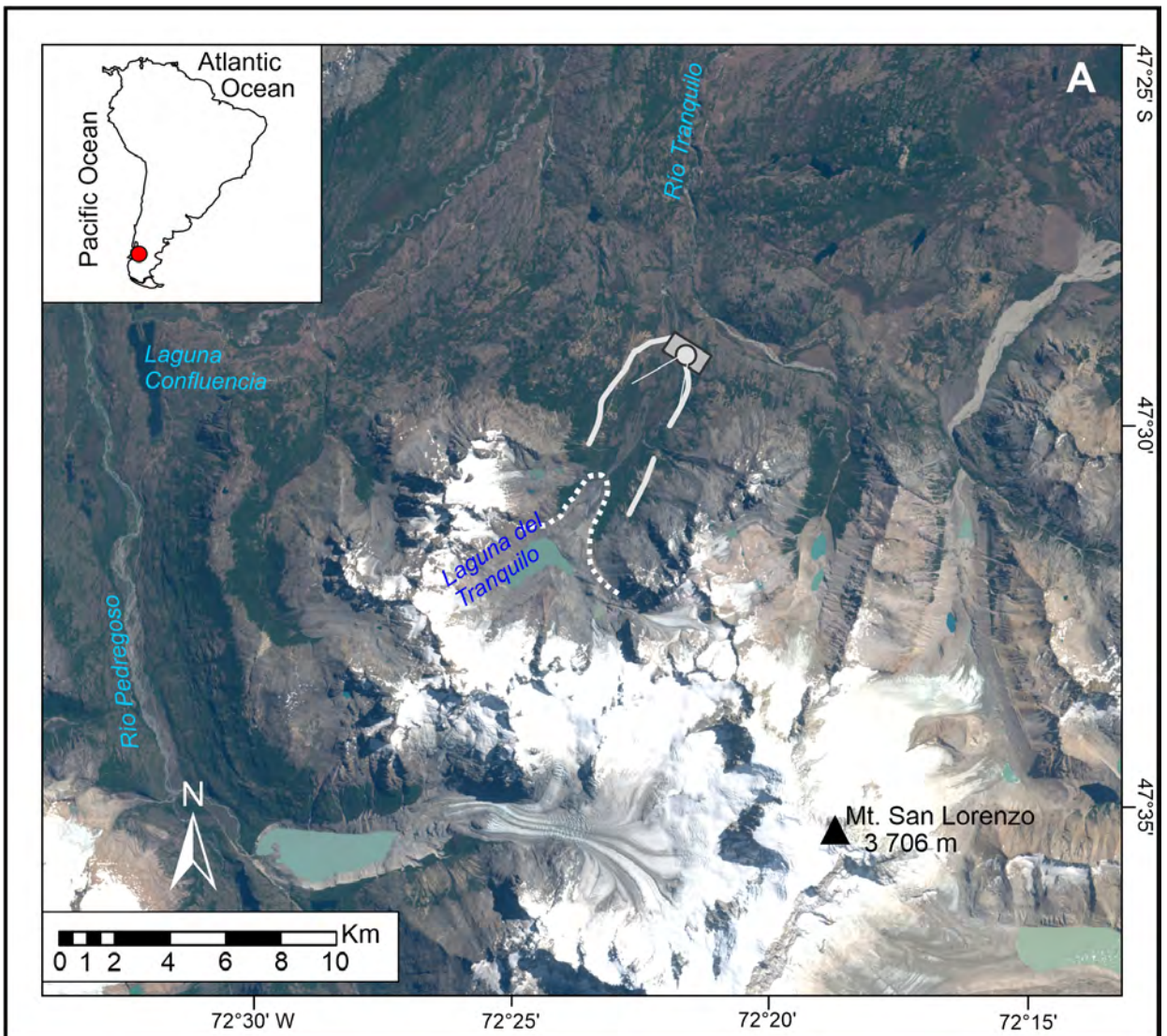












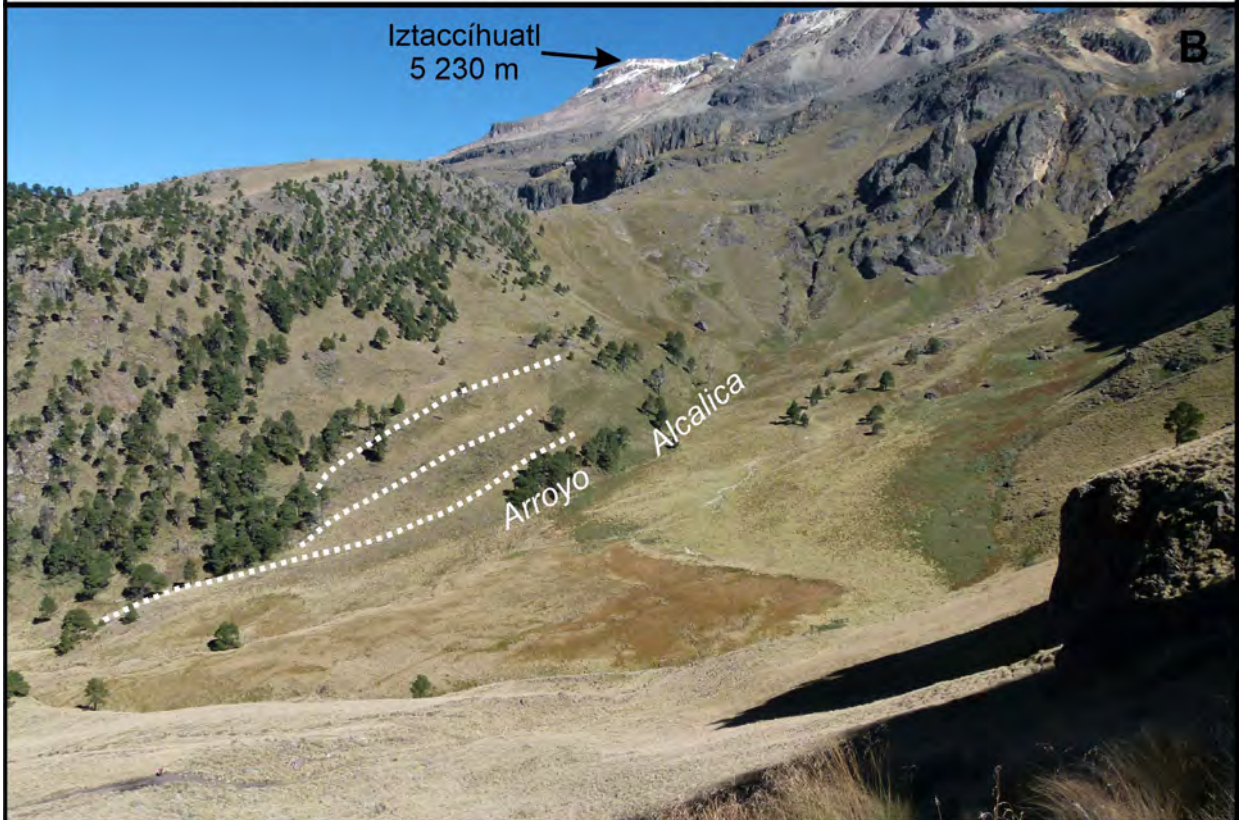
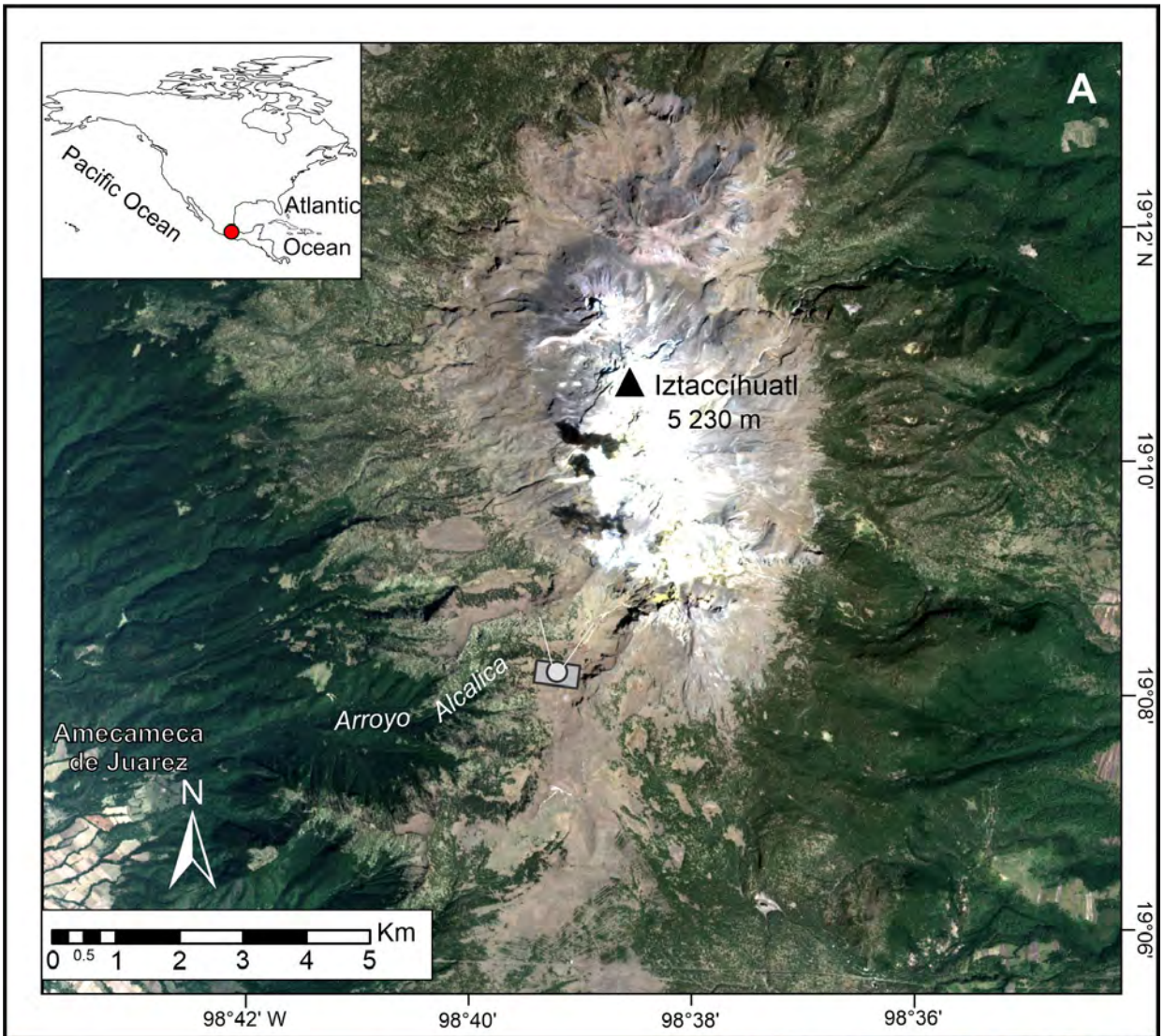


Table 1. The main climate and glacial evolution features during the Global Last Glacial Maximum (26.5 to 19 ka) in the Americas

REGION	Climate during GLGM in relation to present	ELA depression in LGM in relation to present	Local Last Maximum Ice Extension and its relation to GLGM	Initial deglaciation chronology	Key References
<b>Laurentia</b>	6-7 °C colder and great local variability in precipitation		During GLGM Probably from 28 to 20 ka (even 17 ka) with great local variability LIS near-maximum extent for several thousand years	23 to 20 ka, depending on area, but slow prior to 17 ka	Dyke et al., 2002; Ullman et al., 2015b; Robel and Tziperman, 2016; Stokes et al., 2016; Margold et al., 2015; 2018; Stokes, 2017
<b>Alaska</b>	More arid and relatively warm conditions	200-500 m	During GLGM From 24 to 21 ka in the whole region	~20 ka	Kaufman et al., 2003; Tulenko et al., 2018; Pendleton et al., 2015; 2012; Putnam et al., 2013; Briner et al., 2017
<b>Cordillera Ice Sheet and North Cascades</b>	6-7 °C colder and 40% lower mean annual precipitation	1000–750 m	CIS in the north during GLGM and in the west and south, later up to 16 ka During LGM in North Cascades, from 25 to 21 ka	From 21 to 16 ka	Clague, 2017; Riedel et al. 2010; Bartlein et al. 2011; Riedel, 2017
<b>Yellowstone-Tetons</b>	Around 5.4°C colder and similar precipitation as at present	900 m	During GLGM From ~22 ka, but in some valleys up to 17-16 ka	~20 to 17 ka, depending of the valleys	Licciardi and Pierce, 2018; Pierce, et al., 2018
<b>Wind River Range, WY</b>	(Est.) > 5.4°C colder and similar precipitation as at present	~900 m	During GLGM From ~24-22 ka	<22 ka	Dahms 2004; Dahms et al., 2018, 2019; Shakun et al., 2015
<b>Colorado</b>	Around 5.4-8°C colder and similar precipitation as at present	900 m	During GLGM From ~24 ka, but some valleys up to 17-16 ka	~20 to 17 ka, depending of the valleys	Young et al., 2011; Leonard et al., 2017a, b ; Brugger et al., 2019
<b>Sierra Nevada</b>	5-6°C lower, 140% higher precipitation	1200 m	During GLGM From 26 to 20 ka	Began at 19 ka and accelerated rapidly after 18 ka	Plummer, 2002; Phillips, 2016, 2017
<b>Mexico</b>	6 - 9°C colder, precipitation lower than modern	1000 m in the interior; 1200-1500 m	During GLGM. Interior part of central Mexico at 21-19 ka, extending to 20-14 ka near the coasts	15 to 14 ka	Vazquez-Selem and Heine, 2011; Vázquez-Selem and Lachniet, 2017



		near the coasts			
<b>Central America</b>	6 - 9°C colder	1100-1400 m	During GLGM and post-LGM Maximum at ~25-23 ka in Cordillera Talamanca; subsequent deglaciation and moraines formation ca. 18-16 ka	~17 ka	Roy and Lachniet, 2010; Cunningham, 2019; Potter et al., 2019
<b>Northern Andes</b>	8°C colder and wetter	1420-850 m	Pre-GLGM but during the GLGM, ~21 ka in Sierra Nevada, Venezuela	From ~21 to 16 ka	van der Hammen, T., 1981; Schubert and Rinaldi, 1987; Thouret et al., 1996; Stansell et al., 2007; Brunshön and Behling, 2009; Angel, 2016; Angel et al., 2016; Angel et al., 2017
<b>Peru and Bolivia</b>	~5-8° C colder precipitation was slightly higher	Variability, from 800 to 1200 m	Pre-GLGM; GLGM and post GLGM, with average of 25 ka, from 32 to 17 ka	From ~22 to ~17 ka	Seltzer et al., 2002; Mark et al., 2005; Rodbell et al., 2008; Mark, 2017
<b>Northern Chile</b>	Colder and strong spatial gradients in moisture availability from north to south	900 m	Pre-GLGM 45-35 ka 30-25 ka and GLGM ~22-19 ka, and in some sectors 17-15 ka,	North Arid Diagonal, up to < 15 ka. South 17 ka	Ward et al., 2015; Cesta and Ward, 2016; Ward et al., 2017; Martin et al., 2018
<b>Central Andes of Argentina</b>	Colder and slightly wetter conditions	800 m	Pre-GLGM (~40-50 ka), with LGM minor and variable expansion: between ~26 and ~19 ka	From ~21 ka	Martini et al., 2017; Moreiras et al., 2017; Zech J. et al., 2017; Terrizano et al., 2017; D'Arcy et al., 2019
<b>Patagonia</b>	6-7° C colder and wetter in northern Patagonia	ca. 900 m	Pre-GLGM (~48 and ~30 ka), with LGM minor expansion and variable: between ~26.9 and 18.8 ka	between 19.5-17.5-ka	Moreno et al. 2015; Darvill et al., 2016; Hein et al., 2017; Garcia et al., 2018; Hubbard et al., 2005; Denton et al., 1999
<b>Tierra del Fuego</b>	6-7°C colder	~1000 m?	Pre GLGM? GLGM expansion by ~25 ka, but pre-GLGM moraines may be more extensive	~18 ka	Hall et al., 2013, 2017; Menounos et al., 2013

Table 2. The main climate and glacial evolution features during the Heinrich 1 Stadial (17.5 to 14.6 ka) in the Americas.

REGION	Climate during H-1 in relation to present	ELA depression in H1 in relation to present	Glacial evolution during H-1	Key References
LAURENTIA			No clear evidence for major glacial readvances during or soon after H-1. Important changes in the location of ice dispersal centres, with subsequent effects on ice flow patterns and some lobe advances. In general, ice sheet thinning tendency	Stokes et al., 2005; Koester et al., 2017;
ALASKA	Cold, mild	200-500 m	Identified some standstills or re-advances in glacial fronts at ~17-16.5 ka, but there was significant recession of most of glaciers over all this period	Pendleton et al., 2015; Kopczynski et al., 2017; Briner et al., 2017
Cordillera Ice Sheet and North Cascades	6-7 °C colder and 40% lower mean annual precipitation influenced by CIS expansion	1000–750 m few hundred meters above the LGM	The ice sheet reached its maximum extent during H-1 Some glaciers left moraines closely nested inside the LGM moraines, that could belong to the H-1 expansion	Cosma et al., 2008; Troost, 2016; Clague, 2017; Kaufman et al., 2004; Porter and Swanson, 2008; Riedel et al., 2010
Yellowstone-Tetons	Colder and similar precipitation	900 m, with great local variability	Some valleys reach their local maximum ice advance at ~17 ka, while others experienced extensive recession.	Licciardi and Pierce, 2018; Pierce et al., 2018
Wind River Range, WY	Est. colder w/ similar precipitation	60-170m (S-to-N)	Stagnant ice remains w/in 13 km of LLGM max in some trunk valleys while riegels become ice-free, suggesting de-coupling of cirque from valley ice ~18-17 ka. ‘Temple Lake’ moraines form 15-14 ka	Shakun et al., 2015a; Dahms et al., 2018, 2019; Marcott et al., 2019.
Colorado	Colder and similar precipitation	900 m, with great local variability	Some valleys reach their local maximum ice advance at ~17-16 ka, while others experienced extensive recession. Rapid retreat afterwards.	Laabs et al., 2009; Young et al., 2011; Shakun et al., 2015a; Leonard et al., 2017a, b; Brugger et al., 2019
Sierra Nevada	3° colder and 160% of precipitation	900 m	Strong evidence of glacial advance during H1 at 16.2 ka. Extensive retreat afterwards	Plummer, 2002; Phillips, 1996, 2017
Mexico	Cold and dry	900 m in the interior; 1200-1500 m near the coasts	Minor recession in the interior but overall strong evidence of glacial advance or stillstand coeval to H-1 (17-15 ka)	Vázquez-Selem and Lachniet, 2017

<b>Central America</b>	Cold conditions	Unknown	Moraine formation dated at ca. 18-16 ka; glacier recession by 15 ka	<a href="#">Cunningham et al., 2019</a> ; <a href="#">Potter et al., 2019</a>
<b>Northern Andes</b>	Cold temperatures but higher than in the LGM	Unknown, but likely with great variability	Local glacial advances from 17.5-15 ka in the context of general deglaciation	<a href="#">Rull, 1998</a> ; <a href="#">Brunschön and Behling, 2009</a> ; <a href="#">Angel et al., 2016</a> ; <a href="#">2017</a>
<b>Peru and Bolivia</b>	The coldest period since LGM with cooling of ~3°C. Drier in central Peru. In the Altiplano the first part of H1 (~18 to 16.5 ka) was drier, followed by the Lake Tauca highstand from 16.5 to 14.5 ka with an increase of precipitation > 130 %	Large variability, from similar to LGM to few hundred meters	Most of the glaciers retreated just prior to H-1, but multiple glaciers re-advanced during H-1, in Peru and Bolivia, with average moraine age of 16.1 ka with a standard deviation of 1.1 ka	<a href="#">Alcalá-Reygosa, 2017</a> ; <a href="#">Mark, 2017</a> ; <a href="#">Martin et al., 2018</a>
<b>Northern Chile</b>	~3.5°C colder and sharp precipitation gradient with decreasing precipitation from Lake Tauca to the south.	900 m of regional difference in relation to precipitation distribution	Moraines dating to the LGM and earlier were overridden during the Tauca highstand wet phase (~17-15 ka) in mountains surrounding Lake Tauca, but to the south the Tauca-phase moraine is either absent or found high in the valleys.	<a href="#">Ward et al., 2017</a> ; <a href="#">Martin et al., 2018</a>
<b>Central Andes of Argentina</b>	Glacial advances occurred synchronous with the expansion of Altiplano lakes	620 m	New advances in some sectors between 17 and 15 ka and in other sectors, deglaciation from 17 ka. Moraines of H-1 located up-valley from those of the LGM.	<a href="#">Zech J. et al., 2009</a> , <a href="#">2017</a> ; <a href="#">Martini et al., 2017</a> ;
<b>Patagonia</b>	Warming tendency throughout the period. Average interglacial temperatures by 16.8 ka	Trending toward values similar to the present	Widespread deglaciation along the region, with exception of short stabilization at ~16.9-16.2 ka in a few glaciers. Glaciers in central Patagonia were near to modern ice limits before ~16 ka	<a href="#">Boex et al. 2013</a> ; <a href="#">Moreno et al. 2015</a> ; <a href="#">Henriquez et al., 2017</a> ; <a href="#">Mendelova et al., 2017</a> ; <a href="#">Bendle et al., 2017</a>
<b>Tierra de Fuego</b>	Sudden, large-scale warming.	Unknown, but likely within a few hundred meters of present by 16.8 ka	Rapid glacier recession, with no evidence of stillstands. Cordillera Darwin icefield had contracted to within the present-day fjords by 16.8 ka	<a href="#">McCulloch et al., 2005b</a> ; <a href="#">Kaplan et al., 2008</a> ; <a href="#">Menounos et al., 2013</a> ; <a href="#">Hall et al., 2013</a> , <a href="#">2017</a>

Table 3. The main climate and glacial evolution features during the Bølling-Allerød interstadial and Antarctic Cold Reversal (14.6–12.9 ka) in the Americas.

REGION	Climate during B-A/ACR in relation to present	ELA depression during BA/ACR in relation to present	Glacial evolution during B-A/ACR	Key References
LAURENTIA			General ablation in marginal areas in B-A and marked retreat acceleration along the southern and western margins, with development of proglacial lakes. Some readvances not related to climatic forcing. Northern margin stable.	<a href="#">Carlson et al., 2012</a> ; <a href="#">Ullman et al., 2015b</a> ; <a href="#">Margold et al., 2015</a> ; <a href="#">2018</a> ; <a href="#">Dyke, 2004</a> ; <a href="#">Stokes, 2017</a>
ALASKA	Unknown but some data suggest an important climate change		Widespread glacier retreat around the time of the Bølling onset. Glaciers were smaller than their eventual late Holocene extents by 15 ka	<a href="#">Badding et al., 2013</a> ; <a href="#">Pendleton et al., 2015</a>
Cordillera Ice Sheet and North Cascades	Positive temperature anomaly of 1–2 °C early in the BA, and increase in mean annual precipitation of 250 mm. Brief cold periods with -1.5 °C temperature drop caused the small glacier advances later in the BA	CIS lowered more than 500 m in the early BA. ELAS ~500-700 m	Rapid disintegration of the CIS in the North Cascades and western Canada from 14.5–14.0 ka Glacial tongues transformed into a labyrinth of dead ice in valleys and between CIS and LIS Minor glacial advances between 13.6 and 13.3 ka. Top-down deglaciation of the ice sheet led to exposure of valley heads and cirques before adjacent valley floors	<a href="#">Liu et al. 2009</a> ; <a href="#">Peltier et al. 2015</a> ; <a href="#">Menounos et al. 2017</a> ; <a href="#">Lambeck et al. 2017</a> ; <a href="#">Riedel 2017</a> ; <a href="#">Clague 2017</a>
Yellowstone-Tetons	warming climate	Increasing to present day	Some valleys continue advancing after 16 ka in the SW, due to their exposure to greater precipitation, the Yellowstone as a whole experienced an intensive deglaciation 15-14 ka	<a href="#">Licciardi and Pierce, 2018</a> ; <a href="#">Pierce et al., 2018</a>
Wind River Range, WY	Warming prior to cooling during IACP/ACR	Increasing to near present	Glaciers retreat behind their H-1 moraines by ~15 ka, possibly to cirque headwalls. <sup>10</sup> Be ages suggest some begin to re-advance ~13.6-3 ka correlative with the Inter-Ållerød Cold Period (IACP) and the ACR	<a href="#">Dahms et al., 2018, 2019</a> ; <a href="#">Shakun et al., 2015</a> .
Colorado	temperature and summer insolation increasing	Increasing to present day	Deglaciation was culminated by ~14 ka, and by 13 ka most of the glaciers had disappeared	<a href="#">Laabs et al., 2009</a> ; <a href="#">Young et al., 2011</a> ; <a href="#">Shakun et al., 2015a</a> ; <a href="#">Leonard et al, 2017a,b</a>
Sierra Nevada	summer-warm and high-insolation	Increasing to present day, with a short depression of ~150 m	Glaciers had retreated to cirque headwalls by about 15.5 ka, well before the start of the BA at 14.7 ka.	<a href="#">Bowerman, 2011</a> ; <a href="#">Phillips, 2016, 2017</a>

			Glacier advanced following the Bølling-Ållerød transition, for short interval, correlative with both the Inter-Ållerød Cold Period and the ACR, but also the YD	
<b>Mexico</b>	Warming tendency	rose at least 200 m	Initial deglaciation, allowing for the formation of small recessional moraines close to those of the maximum advance from 15 to 14 ka and accelerated from 14 ka.	<a href="#">Vázquez-Selem and Lachniet, 2017</a>
<b>Central America</b>	Warming trend	Unknown	Glacier retreat and standstills 15-10 ka	<a href="#">Potter et al., 2019</a>
<b>Northern Andes</b>	2.9±0.8°C colder and 10% increase in annual precipitation	Depression of 500 m	Glacier advances have been related to ACR in the Sierra Nevada del Cocuy, Colombia	<a href="#">Jomelli et al., 2014, 2016</a>
<b>Peru and Bolivia</b>	Colder in the beginning and end of the ACR and variability in the precipitation	Great variability, from similar to present to depression of 500 m	Glaciers advancing in multiple regions of Peru of Bolivia around 13.5 ka	<a href="#">Jomelli et al., 2014, 2016, 2017</a> ; <a href="#">Stansell et al., 2015, 2017</a>
<b>Northern Chile</b>	Drier than H1 or YD		No glacial advances have been dated with sufficient precision to distinguish between the YD and the ACR, but possible ACR moraines at recessed positions	<a href="#">Ward et al., 2015; 2017</a>
<b>Central Andes of Argentina</b>	Similar to present-day precipitation		No generalized glacial activity in the region, but possible ACR moraines	<a href="#">D'Arcy et al., 2019</a>
<b>Patagonia</b>	1.6-1.8°C colder	Depression of 260 m	Several advances during ACR south of 47°S but no evidence of a glacial advance north of 47°S	<a href="#">Moreno et al., 2009</a> ; <a href="#">Glasser et al., 2011</a> ; <a href="#">Sagredo et al., 2011</a> ; <a href="#">Strelin et al., 2011</a> ; <a href="#">García et al., 2012</a> ; <a href="#">Nimick et al., 2016</a> ; <a href="#">Sagredo et al., 2018</a>
<b>Tierra de Fuego</b>	Colder		Cirque moraines in the Fuegian Andes have been dated to the ACR. Overall, little work has been done on ACR ice extent in this region.	<a href="#">Menounos et al., 2013</a>

Table 4. The main climate and glacial evolution features during Younger Dryas (12.9–11.7 ka) in the Americas.

<b>REGION</b>	<b>Climate during YD in relation to present</b>	<b>ELA depression during YD in relation to present</b>	<b>Glacial evolution during YD</b>	<b>Final stages of deglaciation</b>	<b>Key References</b>
<b>LAURENTIA</b>			Glacial recession slowed and some moraine systems were built because of some marginal re-advances. Several ice streams switched on caused by climatically forced. However, there was rapid retreat along other margins.	Following the YD, deglaciation occurred rapidly. Retreat towards the positions of the major ice dispersal centres with some interruptions, such as the 8.2 ka event, to be completed by 7 ka	<a href="#">Dyke, 2004</a> ; <a href="#">Margold et al., 2018</a> ; <a href="#">Lakeman et al., 2018</a> ;
<b>ALASKA</b>	Cooling is notable only in the South and in the beginning of YD. The rest was warmer than the late Holocene. Increase in precipitation due to transgressive flooding of Bering Strait around the time of the YD.	<0-80 m	Limited evidence for glacier re-advances and most of the glaciers retreated up-valley of their late Holocene extent prior to the YD		<a href="#">Briner et al., 2002</a> ; <a href="#">2017</a> ; <a href="#">Kaufman et al., 2010</a>
<b>Cordillera Ice Sheet and North Cascades</b>	The climate was variable at the century time scale, with maximum cooling of ~2–3 °C	ELA 200-400 m below modern values in the North Cascades, but fluctuated 100-200 m	Many alpine glaciers and at least two remnant lobes of the CIS advanced. They were small advances, only several hundred meters beyond late Holocene maximum positions attained during the Little Ice Age. In North Cascades there are multiple closely nested YD moraines	Ice-free by 11.6 ka in many sites in western Canada. In British Columbia extent was only slightly larger at 11 ka than today.	<a href="#">Osborn et al., 2012</a> ; <a href="#">Menounos et al., 2017</a> ; <a href="#">Riedel, 2017</a> ;
<b>Yellowstone-Tetons</b>			The only glacial advance detected is in the Lake Solitude cirque in the east slope of		<a href="#">Licciardi and Pierce, 2008</a>

			Teton Range, with a moraine that closed a small cirque at $12.9 \pm 0.7$ ka		
<b>Wind River Range, WY</b>	Cooling; Precip ?	~40-80m (S-to-N)	'Alice Lake' moraines in fifteen cirque valleys 0.1-0.5 km behind H-1/Oldest Dryas moraines. 13.6-11.2 ka in Stough Basin and Cirque of the Towers; 13.3 ka in Titcomb Basin.	Ice-free until Late Holocene. Two re-advances (pre-LIA pro-talus/moraines; LIA moraines).	<a href="#">Dahms, 2002</a> ; <a href="#">Dahms et al., 2010, 2018, 2019</a> ; <a href="#">Shakun et al., 2015</a> .
<b>Colorado</b>	Cooling evidence		Most of the valleys were largely ice-free by ~15-13 ka with no clear evidence of re-advance during the YD.		<a href="#">Leonard et al., 2017a</a>
<b>Sierra Nevada</b>	Evidence of 1°C cooling and 140 % precipitation increase		Unclear whether Recess Peak advance is YD or pre-YD		<a href="#">Bowerman, 2011</a>
<b>Mexico</b>	~4-5°C colder and relatively dry conditions	650-800 m	In the mountains of central Mexico >4200 m glaciers formed a distinctive group of closely spaced moraines at 3800-3900 m from 13-12 to ~10.5 ka	Mountains <4000 m in central Mexico were ice-free by 11.5 ka; mountains <4200 m ice-free between 10.5 and 10 ka; highest peaks of central Mexico (>4400 m) have evidence of a short but distinctive advance 8.5 to 7.5 ka	<a href="#">Vázquez-Selem and Lachniet, 2017</a>
<b>Central America</b>			Conflicting evidence: full deglaciation prior to YD, ca. 15.2 ka ( <a href="#">Cunningham et al., 2018</a> ); glaciation coeval to YD and full deglaciation before 9.7 ka ( <a href="#">Orvis and Horn, 2000</a> ; <a href="#">Potter et al., 2019</a> )	Ice-free before 9.7 ka or probably as early as 15.2 ka, depending on the authors	<a href="#">Orvis and Horn, 2000</a> ; <a href="#">Cunningham et al., 2019</a> ; <a href="#">Potter et al., 2019-</a>
<b>Northern Andes</b>	2.2-3.8°C colder than today, and drier climate in the Venezuelan Andes		Evidence of glacier advance is limited and mainly located at elevations higher than 3800 m a.s.l.		<a href="#">Salgado-Labouriau and Schubert, 1977</a> ; <a href="#">Carrillo et al., 2008</a> ; <a href="#">Rull et al., 2010</a> ; <a href="#">Stansell et al., 2010</a> ; <a href="#">Angel et al. 2017</a>
<b>Peru and Bolivia</b>	More precipitation in the South indicated by high level of Coipasa paleolake	Scarce records and great variability	The ice was in a general retreating phase during the YD. Only local evidence of advances at the start of YD, mainly in southwestern Peru.	Multiple glaciers advanced or experienced stillstands in the Central Andes during the early Holocene. Glaciers then seem	<a href="#">Mark et al., 2017</a> ; <a href="#">Bromley et al., 2011</a> ; <a href="#">Alcalá-Reygosa et al., 2017</a>

				to have rapidly retreated through the remaining early Holocene	
<b>Northern Chile</b>	100-150% wetter than modern		There are no glacial landforms in the immediate vicinity of the Arid Diagonal dated with sufficient precision to distinguish between the YD and the ACR; possible YD moraines in recessed positions at a few locations.		<a href="#">Ward et al., 2015</a>
<b>Central Andes of Argentina</b>	Cooler and wetter conditions. Synchronous with the expansion of Altiplano lakes	480 m	Evidence of local advances during YD	After the YD or H-1 there is no widespread glacial activity	<a href="#">Martini et al., 2017</a> ; <a href="#">D'Arcy et al., 2019</a>
<b>Patagonia</b>	No clear cooling and great variability in precipitation		After reaching their maximum late-glacial extent during the ACR, Patagonian glaciers underwent net recession and thinning during the YD. This general trend was interrupted by stillstands or minor readvances that deposited small moraines south of 47°S	A widespread warm/dry interval is evident between 11-8 ka and the glaciers continued decreasing through the early Holocene, when most glaciers approached their present-day configuration	<a href="#">Moreno et al., 2010</a> ; <a href="#">Strelin et al., 2011</a> ; <a href="#">Kaplan et al., 2016</a> ; <a href="#">Glasser et al., 2012</a> ; <a href="#">Sagredo et al., 2018</a> ; <a href="#">Moreno et al., 2018a,b</a>
<b>Tierra de Fuego</b>	No clear cooling		A glacier in the Fuegian Andes had reached positions comparable to the Little Ice Age by 12.5-11.2 ka. Probably glaciers were in recession throughout Tierra del Fuego.	Recession may have been rapid in the early Holocene	<a href="#">Menounos et al., 2013</a> ; <a href="#">Hall et al., 2013</a> ; <a href="#">2017</a>



## Declaration of interests

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests: