

1           **WARM SUMMERS AND RICH BIOTIC COMMUNITIES DURING**  
2                                   **N-HEMISPHERE DEGLACIATION**

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33 **ABSTRACT**

34

35 Detailed studies on fossil remains of plants or animals in glacial lake sediments are rare. As a result,  
36 environmental conditions right at the moment of deglaciation of the large N-Hemisphere ice-sheets  
37 remain largely unknown. Here we study three deglacial phases of the Fennoscandian Ice Sheet as a  
38 unique, repeated element in a long sediment record preserved at Sokli in northern Finland. We  
39 summarize extensive multi-proxy data (diatoms, phytoliths, chironomids, pollen, spores, non-pollen  
40 palynomorphs, macrofossils, lithology, loss-on-ignition, C/N) obtained on glacial lake sediments dated  
41 to the early Holocene (ca. 10 kyr BP), early MIS 3 (ca. 50 kyr BP) and early MIS 5a (ca. 80 kyr BP). In  
42 contrast to the common view of an unproductive ice-marginal environment, our study reconstructs rich  
43 ecosystems both in the glacial lake and along the shores with forest on recently deglaciated land. Higher  
44 than present-day summer temperatures are reconstructed based on a large variety of aquatic taxa. Rich  
45 biota developed due to the insolation-induced postglacial warming and high nutrient levels, the latter  
46 resulting from erosion of fresh bedrock and sediment, leaching of surface soils, decay of plant material  
47 under shallow water conditions, and sudden decreases in lake volume. Aquatic communities responded  
48 quickly to deglaciation and warm summers and reflect boreal conditions, in contrast to the terrestrial  
49 ecosystem which responded with some delay probably due to time required for slow soil formation  
50 processes. Birch forest is reconstructed upon deglaciation of the large LGM ice-sheet and shrub tundra  
51 following the probably faster melting smaller MIS 4 and MIS 5b ice-sheets. Our study shows that glacial  
52 lake sediments can provide valuable palaeo-environmental data, that aquatic biota and terrestrial  
53 vegetation rapidly accommodated to new environmental conditions during deglaciation, and that glacial  
54 lake ecosystems, and the carbon stored in their sediments, should be included in earth system modeling.

55

56 **Keywords: glacial lake sediment, ice-marginal environment, climate, ecosystem response, carbon**  
57 **storage, Fennoscandian Ice Sheet**

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59

## 60 **1 Introduction**

61

62 Glacial lake sediments are a persistent element in the geological record (Carrivick and Tweed, 2013;  
63 Lunkka et al., 2015) and former extents of glacial lakes have been mapped in different regions globally  
64 (Lundqvist, 1972; Björck, 1995; Dyke, 2004; Glasser et al., 2016). Particularly the drainage of large  
65 glacial lakes has received significant attention as it has the potential of influencing ocean circulation and  
66 climate on a global scale (Broecker and Denton, 1990; Barber et al., 1999).

67 Extensive ice-marginal retreat in the Fennoscandian and Laurentide Ice Sheets occurred at the last  
68 peak of high-latitude summer insolation around ca. 10 kyr cal BP (Dyke, 2004; Stroeven et al., 2016).  
69 The postglacial warming allowed for the production of large amounts of meltwater and, with the land-  
70 surface inclined towards the retreating ice-fronts due to isostatic depression, glacial lakes were formed  
71 extensively (Fig. 1). Although the sediments deposited in these lakes and fossil remains stored in them  
72 form a potential natural archive for palaeo-environmental and -climatic studies, the biotic content of  
73 glacial lake sediments has been rarely analyzed in detail. The lack of fossil analysis on glacial lake  
74 sediments is probably due to the generally low fossil content and often broken/corroded nature of fossil  
75 remains (e.g. Lunkka et al., 2015), making the analysis particularly time-consuming. Also, fossil remains  
76 are often considered as redeposited from older deposits as the ice-marginal environment is classically  
77 interpreted as harsh and unproductive (e.g. Stroeven et al., 2016). The difficulty of recovering the  
78 minerogenic glacial lake sediment at the base of Holocene peat or gyttja sequences with the help of  
79 conventional hand-coring adds yet another reason for the lack of studies on these sediments.

80 The north-eastern margin of the Fennoscandian Ice Sheet receded from the Sokli basin in northern  
81 Finland (Fig. 1) several times during Marine Isotope Stages (MIS) 5 to 1. Glacial lake deposits dated to  
82 early MIS 5e, 5a, and 3 and the early Holocene are found here in stratigraphic sequence with glacial till  
83 beds and non-glacial lacustrine and fluvial deposits. During each deglaciation episode there was a  
84 progression of lake stages associated with deglaciation (stage 1), maximum extent of the glacial lake  
85 (stage 2), and glacial lake drainage (stage 3). The Sokli sequence uniquely preserves lake sediments  
86 from four deglaciation episodes due to a non-typical bedrock setting combined with frozen-bed  
87 conditions at the base of the Fennoscandian Ice Sheet (Helmens et al., 2007a, and references therein).

88           The Sokli sedimentary sequence provides an exceptional opportunity to study the ice-marginal  
89 environment during the retreat of a continental ice-sheet as a repeating element in the Late Quaternary  
90 geological record at a single site. We here integrate multi-proxy data (diatoms, phytoliths, chironomids,  
91 pollen, spores, NPP's (non-pollen palynomorphs), macrofossils, lithology, LOI (loss-on-ignition), C/N)  
92 that are available for the glacial lake sediments in the Sokli basin dated to the early Holocene (Shala et  
93 al., 2014a, 2014b, 2017) and early MIS 3 (Engels et al., 2008; Bos et al., 2009; Helmens et al., 2009),  
94 and further compare this extensive data-set with newly obtained data for multiple proxies from the  
95 glacial lake sediments of early MIS 5a age. The overall coarse-grained glacial lake sediment found at  
96 the base of the MIS 5e gyttja deposit contains very few fossil remains (Pliikk et al., 2016) and is not  
97 included in this study. The aim of our study is to reconstruct in detail biota in the glacial lake, along its  
98 shores and on the surrounding land, as well as successive developments in these ecosystems upon  
99 deglaciation and glacial lake evolution, for a series of Late Quaternary deglaciation episodes at the same  
100 site. Furthermore, the multi-proxy data are used to make climate inferences. Our study provides valuable  
101 paleo-environmental data and new understanding on the aquatic and terrestrial productivity of the ice-  
102 marginal environment during N-Hemisphere deglaciation.

103

## 104 **2 Environmental and stratigraphic setting**

105

### 106 **2.1 The study site**

107

108 The Sokli site is situated in the northern boreal forest of north-eastern Finland (lat. 67°48' N, long. 29°18'  
109 E, elevation ca. 220 m a.s.l.) and on the main water divide that separates drainage into the Barents and  
110 White Seas to the east and the Baltic Sea to the southwest (Fig. 1). Bedrock in the region is Precambrian  
111 Shield with the exception of the immediate surroundings of the study site which is underlain by a  
112 Paleozoic magma-intrusion known as Sokli Carbonatite Massif (Fig. 2B). The Sokli wetland (Sokliaapa)  
113 is drained by the Sokli rivulet (Soklioja), the latter flowing south-westwards into the Yli-Nuortti river.  
114 Within the Sokli basin, Lake Loitsana occupies a depression associated with a NW-SE trending esker  
115 chain.

116 Present climate at Sokli is cold boreal with mean July and February temperatures at 13°C and -  
117 14°C, respectively; mean annual precipitation amounts to 500-550 mm (Drebs et al., 2002). Lakes in  
118 the area are ice-covered between October and end of May. Birch (*Betula pubescens* and *B. pendula*),  
119 pine (*Pinus sylvestris*) and spruce (*Picea abies*) are the dominant tree species. Mires of the aapa-type  
120 (i.e. a patterned fen) with *Sphagnum* spp., *Rubus chamaemorus*, Ericales, *Betula nana* (dwarf birch),  
121 *Salix* spp. and *Carex* spp. are extensively present in the region. Spruce reaches its northern limit some  
122 100 km north of Sokli. Farther north, pine forest predominates, succeeded northwards and upwards by  
123 birch-pine forest and then sub-arctic birch forest. The forest limit, which is situated some 300 km north  
124 of Sokli, is formed by the polycormic mountain birch *B. pubescens* subsp. *czerepanovii*, syn. *tortuosa*.  
125 The vegetation of the tundra region beyond the forest limit is low-arctic dwarf-shrub tundra dominated  
126 by *B. nana* and Ericales.

127

## 128 **2.2 Late Quaternary stratigraphy in the Sokli basin**

129

130 An unconsolidated sedimentary sequence up to ca. 30 m in thickness, including several organic units  
131 and till beds, is present in the central part of the Sokli Carbonatite Massif where two main fault zones  
132 cross (Fig. 2A, B). The unusually long Sokli sediment record, which spans the last ca. 140 kyr, has been  
133 protected from glacial erosion due to its sheltered position in a steep depression formed in the deeply-  
134 weathered rocks of the Sokli Massif.

135 The composite lithological log in Figure 2A is based on a series of boreholes collected between  
136 1996-2010 (Fig. 2B) and is an updated version from lithological columns earlier presented in Helmens  
137 et al. (2000, 2007a). The Sokli sediments have been dated by AMS <sup>14</sup>C dating on macrofossils of  
138 terrestrial plants, TL and IRSL dating, and OSL dating on quartz using SAR dose protocol (Helmens et  
139 al., 2000, 2007a; Alexanderson et al., 2008). OSL dates on quartz yielded large standard errors mainly  
140 due to small sample sizes, relatively poor luminescence characteristics, and uncertainties in dose-rate  
141 determinations. The OSL ages are, however, in sequence and (together with <sup>14</sup>C ages) group according  
142 to stratigraphic units (Fig. 2A); furthermore, the absolute chronology is in agreement with earlier made

143 land-sea comparisons (Fig. 2C; Helmens *et al.*, 2000, 2007a; Alexanderson *et al.*, 2008). The  
144 stratigraphy in the Sokli basin (Fig. 2A, right side) is according to Helmens (2014).

145 A diatom gyttja deposit up to 9 m in thickness, and dated to the Eemian Interglacial (MIS 5e),  
146 stretches as a marker horizon near the base of the Sokli basin infill (Plikk *et al.*, 2016; Fig. 2). The diatom  
147 gyttja overlays coarse-grained glacio-lacustrine sediment resting on till (MIS 6; Ilvonen, 1973). A sand  
148 and gravel deposit of fluvial origin (MIS 5d) separates the MIS 5e lake sediment from sandy gyttja that  
149 was deposited in an oxbow lake during MIS 5c (Helmens *et al.*, 2012). The upper part of the Sokli  
150 sedimentary sequence consists of till interlayered with two additionally well-sorted, fine-grained  
151 lacustrine sequences (MIS 5a and MIS 3). The deposition of glacio-lacustrine silt and clay in early MIS  
152 5a was followed by the accumulation of gyttja (this study). The glacio-lacustrine sediment of early MIS  
153 3 age is capped by till (Helmens *et al.*, 2009). Lake Loitsana holds a 7m-thick gyttja deposit underlain  
154 by a glacio-lacustrine deposit of early Holocene age (Shala *et al.*, 2014a, 2014b). Glaciation of the Sokli  
155 basin is dated to MIS 5b, MIS 4 and late MIS 3-2 (Helmens *et al.*, 2007a).

156

### 157 **2.3 The Sokli Ice Lake**

158

159 The Late Quaternary glacial lake sediments preserved in the Sokli basin were deposited in the Sokli Ice  
160 Lake (Fig. 1). Figure 3 gives a schematic representation of the evolution of the later stages of the Sokli  
161 Ice Lake, i.e. from the moment of deglaciation of the Sokli site (Lake Stage 1) until its final drainage  
162 through the Nuortti canyon (Lake Stage 3). Glacier meltwater was trapped between the hills of the Värriö  
163 tunturi and the retreating ice-margin during the early stages of the glacial lake (Johansson, 1995).

164 The ice-marginal retreat pattern in Figure 3 is guided by NW-SE oriented eskers that are dated to  
165 the last deglaciation in the early Holocene (Johansson, 2007; Shala *et al.*, 2014a). Importantly, a similar  
166 evolution of these final stages of the Sokli Ice Lake has been reconstructed for early MIS 3 (Helmens *et al.*,  
167 2009) based on N-S trending, till-covered eskers in the region dated by OSL at  $65 \pm 13$  kyr BP  
168 (Johansson, 2007). The direction of ice-marginal retreat in early MIS 5a was to the northwest  
169 (Johansson, 1995) and, also during this deglaciation episode, the glacial lake evolution was probably  
170 similar as during the early Holocene. The reconstruction in Figure 3 is based on geomorphologic (e.g.

171 meltwater channels) and DEM (Digital Elevation Model) data and corresponds to the one originally  
172 presented in Johansson (1995). As the area invoked in the glacial lake reconstruction is relatively small  
173 (ca. 30 x 30 km), possible errors generated due to non-uniform isostasy are considered to be insignificant  
174 (cf. Jansson 2003).

175 During each deglaciation episode, the Sokli Ice Lake varied in surface area and water depth  
176 depending on the position of the ice-front and the location of spillways/overflow sites (Fig. 3). During  
177 Lake Stages 1 and 2, the level of the glacial lake was controlled by the col at the head of the Törmäoja  
178 canyon which has a present elevation of 240 m. The glacial lake reached its maximum spatial extension  
179 during Lake Stage 2. Opening of the Nuortti canyon resulted in partial (Lake Stage 3; col at 220 m  
180 elevation) and final drainage of the Sokli Ice Lake.

181

### 182 **3 Methods**

183

184 A total of 2 m of glacial lake silts and clays, below 7 m of gyttja, was retrieved with the Russian peat  
185 corer from ice-covered Lake Loitsana. The 1-2 m thick early MIS 3 and MIS 5a glacial lake deposits,  
186 which underlie the LGM (Last Glacial Maximum) till, were recovered using heavy-equipment  
187 percussion drilling from the frozen surface of the Sokli wetland. Methods used for multi-proxy analysis  
188 on the sediments of early MIS 5a age follow those applied to the glacial lake sediments of early MIS 3  
189 and early Holocene age. The latter are presented in Helmens et al. (2009) and Shala et al. (2014a, 2014b)  
190 (siliceous microfossils and LOI); Engels et al. (2008) and Shala et al. (2014b) (chironomids); Bos et al.  
191 (2009) and Shala et al. (2014b, 2017) (pollen/spores/NPP's and macrofossils); and Shala et al. (2014b)  
192 (C/N). For all three time-slices (early Holocene, MIS 3 and 5a), diatom and chironomid diversities are  
193 calculated using Hill's  $N_2$  (effective number of occurrences; Hill, 1973).

194 A minimum of 500 diatom valves, 50 chironomid head capsules and ca. 400 terrestrial pollen  
195 grains were identified in each sample from the early MIS 5a glacial lake deposit. The sample size for  
196 the macrofossil analysis was mainly 10-15 cm<sup>3</sup>. Furthermore, we submitted two samples from gyttja  
197 found overlying the early MIS 5a deglacial sediment for AMS <sup>14</sup>C dating.

## 198 **4 Results and Discussion**

199

### 200 **4.1 Timing and duration of the Sokli Ice Lake**

201

202 <sup>14</sup>C dating on seeds of tree birch encountered in the youngest glacial lake deposit has provided an age  
203 of ca. 10.5 kyr cal BP for the last deglaciation of the Sokli basin (Shala et al., 2014a, 2014b). This date  
204 is in accordance with the early Holocene deglaciation chronology for north-eastern Fennoscandia as  
205 presented in Johansson (2007) and Stroeven et al. (2016). It also shows that the *Betula* seeds occur in-  
206 situ, i.e. they are not reworked from older deposits. The in-situ preservation of fossil remains in the  
207 glacial lake sediment is further demonstrated by <sup>14</sup>C ages on wood from the early MIS 3 sediment  
208 sequence (Helmens et al., 2000, 2007b) which are in line with OSL dating results (Alexanderson et al.,  
209 2008; Figs. 2 and 4B). The latter deglaciation episode probably occurred at ca. 53 kyr BP during  
210 prominent Greenland Interstadial (GI) 14 (Helmens et al., 2007b; Helmens and Engels, 2010). New <sup>14</sup>C  
211 datings of macrofossils from gyttja that overlies the oldest glacial lake deposit gave, as expected, infinite  
212 ages (Table 1). This episode of deglaciation has been earlier dated to early MIS 5a based on OSL dating  
213 of glacio-fluvial sand and gravel, that occur beneath the glacial lake sediment, at ca. 80 kyr BP and of  
214 minerogenic sediment at the top of the overlying gyttja at ca. 74 kyr BP (Alexanderson et al., 2008; Figs.  
215 2 and 4C).

216 The duration of the Sokli Ice Lake following the last deglaciation of the Sokli site is estimated at  
217 less than 100 yr (Johansson, 2007; Stroeven et al., 2016; Fig. 3). A duration of some 400 kyr is inferred  
218 based on <sup>14</sup>C dating evidence from the early Holocene glacial lake deposit in the Sokli basin (Shala et  
219 al., 2014a, 2014b) and counting of the distinct, possibly annual laminae in the glacial lake sediment of  
220 early MIS 3 age (Helmens et al., 2009). However, the 50 to 100-year error margins on the <sup>14</sup>C dates (Fig.  
221 4A), and the large spread in calendar ages for the earliest Holocene, hamper a detailed age determination  
222 for glacial lake duration.

223

224



## 225 **4.2 Aquatic and telmatic ecosystem changes in the ice-marginal environment during deglaciation**

226

227 The evolution of the Sokli Ice Lake as depicted in Figure 3 can be clearly traced in the sediment sequence  
228 of each deglaciation episode (early Holocene, MIS 3 and 5a; Fig. 4). The ice-marginal retreat during  
229 Lake Stages 1 and 2 is recorded by an upward-fining sequence of sandy and silty sediment grading into  
230 rhythmically laminated silts and clays at the base of each glacial lake sequence. Following the decrease  
231 in level and size of the glacial lake associated with the opening of the Nuortti canyon (Lake Stage 3),  
232 silts (early Holocene) and silts grading into sandy sediment (early MIS 3 and 5a) were deposited at the  
233 coring-sites. Lacustrine sedimentation is recorded to continue with the accumulation of organic silts  
234 followed by gyttja in the Sokli basin during early MIS 5a, and gyttja in Lake Loitsana in the early  
235 Holocene. The early MIS 3 glacial lake sediments were found to be covered by till. Furthermore, the  
236 Sokli Ice Lake evolution has been traced in XRF-based geochemical data available for the early  
237 Holocene glacial lake sediment (Shala et al., 2014a). The latter shows a decrease in sediment input  
238 enriched in elements (Ca, S and Nb) typical for the local bedrock (Sokli Carbonatite Massif) during  
239 Lake Stage 2, i.e. when the Sokli Massif became entirely submerged in the expanding glacial lake (Fig.  
240 3).

241 The biotic proxy diagrams obtained from the early Holocene (Fig. 4A), early MIS 3 (Fig. 4B) and  
242 early MIS 5a glacial lake sediments (Fig. 4C) reveal marked compositional shifts in biotic assemblages  
243 in concordance with the changes in size and depth of the Sokli Ice Lake. This shows that biota responded  
244 quickly to environmental changes during deglaciation. It also provides further support for the in-situ  
245 preservation of fossil remains. Although the biotic data obtained for the different deglaciation episodes  
246 share many similarities, fossil assemblages and successive developments appear to be influenced by  
247 actual lake depth. Overall deepest lake conditions are reconstructed for early MIS 5a, and shallowest  
248 waters during the early Holocene. This can be expected due to continuing infilling of the Sokli basin  
249 and progressive lowering of the glacial lake outlet channels resulting from erosion. Below, the biotic  
250 records will be used to reconstruct in detail the aquatic and telmatic ecosystem composition in and along  
251 the shore of the Sokli Ice Lake during its evolution.

252

#### 253 4.2.1 Glacial Lake Stages 1 and 2

254

255 The fossil content and diversity of fossil remains (expressed by e.g. Hill's  $N_2$  values for diatoms and  
256 chironomids) increase, both during the early Holocene (Fig. 4A) and early MIS 3 deglaciation episodes  
257 (Fig. 4B), with increasing distance of the ice-margin from the coring-site (Lake Stages 1-2 in Fig. 3).  
258 The overall meagre fossil content of the Lake Stage 1-2 sediments of early MIS 5a age (Fig. 4C) most  
259 probably result from deep water conditions (suggested by relatively high planktonic diatom percentages)  
260 and long distance to shore. Furthermore, C/N values (only measured for the Holocene (Fig. 4A))  
261 gradually drop during Lake Stages 1-2 suggesting an increased in-lake productivity. It is less-likely that  
262 the low C/N values in the latter part of Lake Stage 2 result from a decreased contribution of higher plants  
263 since seeds of *Carex* and high pollen percentages of Cyperaceae and *Salix* indicate an expansion in the  
264 wetland zone close to the coring-sites. In general, high C/N ratios ( $> 10$ ) indicate input of organic matter  
265 from higher plants (littoral/wetland/terrestrial), and low ratios ( $< 10$ ) reflect input from phytoplankton  
266 (Wetzel, 2001).

267 Turbulent waters and turbid water conditions occur close to ice-margins and significantly decrease  
268 primary productivity (Henley et al., 2000). Also, the fossil record in the coarse-grained ice-marginal  
269 sediment at the base of the studied sequences (Lake Stage 1) can be expected to be diluted due to high  
270 sedimentation rates (Risberg et al., 1999) and plant/animal remains to be more easily destroyed in this  
271 high-energy environment.

272 Chironomids that are recorded among the first colonizers of the Sokli Ice Lake during all three  
273 time-slices (early Holocene, MIS 3 and 5a) are the deep-water taxa *Heterotrissocladius maeaeri*-type,  
274 *Tanytarsus lugens*-type and *Procladius*. *H. maeaeri*-type is reported to have dominated the deep waters  
275 of the late-glacial Baltic Ice Lake in southern Finland (Fig. 1; Luoto et al., 2010). Diatoms were mostly  
276 Fragilariaceae and *Aulacoseira*. Fragilariaceae is a group of taxa that is characterized as pioneering and  
277 opportunistic, since they colonize quickly, have high reproduction rates and are more adaptive to a  
278 changing environment (Risberg et al., 1996; Lotter et al., 1999). The encountered planktonic  
279 *Aulacoseira* species (*A. ambigua*, *A. subarctica*, *A. alpigena*) are all heavily silicified and most probably  
280 reflect a large influx of Si into the glacial lake and enhanced levels of turbulence. A high Si-level is

281 further suggested by findings of *Tetracyclus glans* (Michel et al., 2006). Statoblasts of the bryozoan  
282 *Fredericella indica* (Økland and Økland, 2001) and high abundances of pooid phytoliths and Tertiary  
283 diatoms in the sediment indicate wave action and significant shore erosion. Tertiary diatoms are  
284 exemplified by species like *Pliocaenicus costatus*, a relict freshwater taxon from the Pliocene, and  
285 *Paralia sulcata* of marine-origin, their robust structure allowing for multiple phases of re-working  
286 (Helmens et al., 20009, and references therein). The pioneering algal *Botrycoccus braunii*, algal type  
287 T.225 (van Geel et al., 1989) and the cladoceran *Daphnia* (water-flea; recorded by ephippia) are  
288 additionally well-represented among the first lake colonizers.

289 Occurrences of the diatom taxa *Epithemia adnata* and *Rhopalodia gibba* in the upper part of the  
290 Lake Stage 2 sediment of early Holocene age suggest pH values exclusively above 7 (van Dam et al.,  
291 1994). High abundances of ostracod remains indicate enhanced Ca-concentrations. This lake water  
292 chemistry mostly probably resulted from rapid leaching of carbonates and other soluble minerals from  
293 surface soils shortly after deglaciation (Engstrom et al., 2000). An enhanced level of nutrients in the  
294 glacial lake during the latter part of Lake Stage 2 in early MIS 3, and particularly during early MIS 5a,  
295 is further suggested by relatively high percentage values of *Stephanodiscus* species (*S. medius*, *S.*  
296 *minutulus*, *S. neoastraea*, *S. alpinus*).

297 Although the recording of the bryozoan *F. indica* suggests stony shores with sparse aquatic  
298 vegetation (Økland and Økland, 2005), macrofossils, pooid phytoliths and pollen indicate the presence  
299 of the herbs *Rorippa palustris*, *Filipendula* and grasses (e.g. *Glyceria lithuanica*) as well as *Empetrum*  
300 *nigrum*, *Carex*, *Juncus* *Equisetum* and *Sphagnum* in local shore or wetland habitats (i.e. telmatic habitat)  
301 along the glacial lake shoreline.

302

#### 303 4.2.2 Glacial Lake Stage 3

304

305 The local establishment of diverse aquatic and telmatic communities is recorded in the silty glacial lake  
306 sediments of Lake Stage 3. This stage followed the opening of a new spillway which resulted in a drastic  
307 decrease in water depth and size of the Sokli Ice Lake (Fig. 3). The margin of the Fennoscandian Ice  
308 Sheet during Stage 3 was situated at a distance of less than 10 km from the coring-site (i.e. before

309 deglaciation of the Nuortti canyon in Fig. 3). The amount of macrofossils in lake sediments is strongly  
310 influenced by taphonomic factors, e.g. with a close location of the shoreline sharply increasing  
311 macrofossil abundances at the coring-site (Hannon and Gaillard, 1997; Väiliranta, 2006), and  
312 macrofossils are distinctly more abundant in the Lake Stage 3 sediment, particularly in the early  
313 Holocene. Below, the Lake Stage 3 aquatic and telmatic ecosystems will be discussed separately for the  
314 three deglaciation episodes.

315 Fossil remains in the early Holocene deposit (Fig. 4A) reflect a shallow lake and even further  
316 shallowing, accompanied by an extension in the wetland zone (macrofossils of *Carex*, *B. nana*,  
317 *Vaccinium*, *Typha* and *Ranunculus* sect. *Batrachium*), higher up in the Stage 3 silts. Characteristic is the  
318 high representation in the macrofossil record of the shallow-water plant *Callitriche*  
319 *cophocarpa/hermaphroditica*, the bryozoan *Plumatella repens* and *Daphnia*. Also, various littoral  
320 chironomid taxa (*Polypedilum nubeculosum*-, *Microtendipes pedellus*-, *Dicrotendipes nervosus*-type)  
321 are recorded. A large variety of *Potamogeton* species (*P. berchtoldii*, *P. filiformis*, *P. friesii*, *P.*  
322 *obtusifolius*, *P. pectinatus*), aquatic Nymphaeaceae (*Nymphaea*, *Nuphar*), and the bryozoan *Cristatella*  
323 *mucedo* are additionally well-represented in the upper part of the silt deposit. The pioneering, narrow-  
324 leaved *Potamogeton* species thrive in shallow (0.1-1.5 m) water depths. *C. mucedo* (if produced locally)  
325 suggest more coloured water with less wave action than *F. indica* (Økland and Økland, 2005).  
326 Simultaneously to the trend in lake shallowing, rising nutrient levels are suggested by increasing  
327 abundances of the chironomid taxon *Endochironomus albipennis*-type (Moller Pillot, 2009). The  
328 occurrences of *C. mucedo*, Nymphaeaceae and *Typha* indicate boreal conditions. In contrast to these rich  
329 biotic communities, diatom assemblages in the early Holocene Stage 3 sediment are largely dominated  
330 by *Staurosira construens* var. *venter*, showing peak values reaching > 85%. This might reflect the littoral  
331 and periphytic (growing on plants) habitat preferences of this diatom species and/or tolerance to poor  
332 light conditions (Bigler et al., 2003), the latter possibly resulting from erosion of the adjacent esker.

333 Shallowing of the glacial lake and accompanying rising nutrient levels are recorded in detail in  
334 the thick early MIS 3 silty deposit. Initially, *F. indica* (Økland and Økland, 2005) and the chironomid  
335 taxon *Ablabesmyia* (Vallenduuk et al., 2007) suggest moderate nutrient availability. Subsequently,  
336 nutrient levels increase as indicated by high percentage values for *E. albipennis*-type combined with

337 occurrences of *P. friessi* and the macro-algae *Nitella*. The appearance of the cladoceran *Simocephalus*  
338 in the upper sandy part of the deposit suggests shallow, open water, eutrophic conditions and dense  
339 stands of vegetation (van Geel et al., 1983). Strongly reduced water depths during deposition of the  
340 upper sandy silts of Stage 3 are further indicated by enhanced values for the shallow-water algae  
341 *Spirogyra* and the littoral diatom taxon *Staurosirella pinnata*, as well as a distinct extension in the  
342 wetland zone. The latter is recorded by high abundances of bryophyte leaves, seeds of *Carex*, *Juncus*,  
343 *B. nana* and *E. nigrum*, increased percentage values for Cyperaceae pollen and a near continuous  
344 registration of hyphopodia of *Clasterosporium caricinum*, i.e. a fungus that parasitizes on *Carex* (van  
345 Geel et al., 1983).

346 Fossils in the silty Glacial Lake Stage 3 sediment of early MIS 5a age include macro-remains of  
347 *P. repens*, *Callitriche* (*C. cophocarpa*, *C. hermaphroditica* *C. hamulata*), narrow-leaved *Potamogeton*  
348 (including *P. compressus*) and the shallow-water plant *Ceratophyllum*. Characteristic is the combined  
349 appearance of *Tanytarsus pallidicornis*- and *Psectrocladius sordidellus*-type, i.e. chironomid taxa  
350 presently common in the sublittoral and littoral zones of boreal lakes mostly found living among  
351 *Phragmites* stands (Luoto, 2010). Nutrient availability was overall high as indicated by occurrences of  
352 *E. albipennis*-type, *Chironomus plumosus*-type (Brooks et al., 2001) and a high representation of  
353 *Stephanodiscus* among the diatoms. The enhanced nutrient level, and high chironomid species diversity  
354 (high Hill's N<sub>2</sub> values), might be related to inflow of running water close to the coring-site as also has  
355 been recorded in the MIS 5d-c deposit in the Sokli basin (Engels et al, 2010). Inflow of water is indicated  
356 by relatively high abundances of stream-inhabiting chironomid taxa (e.g. *Eukiefferiella* and  
357 *Rheocricotopus*) as well as the type of lamination of the silty deposit, i.e. laminae of varying grain-sizes  
358 (clay to fine sand) and thicknesses. Similar as during the early Holocene and MIS 3, a strong reduction  
359 in water depth during deposition of the uppermost sandy part of the MIS 5a silt deposit is reflected in  
360 the fossil record (littoral *M. pedellus*-type and *S. pinnata*; macrofossils of *Carex*, *Salix*, *B. nana*).

361 Interestingly, the diatom *Aulacoseira granulata* var. *angustissima* appears with peak values in the  
362 upper sandy sediment of Lake Stage 3 both during early MIS 5a and MIS 3 and, together with increased  
363 representations of the diatom taxa *Staurosirella leptostauron*, *Diploneis elliptica* and *Amphora libyca*,  
364 reflect alkaline waters with high Ca-concentrations (Gómez et al., 1995; Jones et al., 2004). This lake

365 water chemistry was recorded in the early Holocene glacial lake sequence already during the relatively  
366 shallow-water Lake Stage 2. It is possible that the glacial lake volume during early MIS 3 and 5a became  
367 only reduced enough during Lake Stage 3 in order for alkaline water conditions to be established.

368

#### 369 4.2.3 Early Loitsana and Sokli lakes

370

371 The early Holocene sandy gyttja deposited in Lake Loitsana contains relatively high fossil abundances  
372 of the chironomid taxa *Cricotopus intersectus*- and *Cricotopus cylindraceus*-type. These are typical  
373 littoral taxa with an affinity for plants (Brooks et al., 2007; Luoto, 2010) and their occurrences might be  
374 related to an influx of insect and plant remains (e.g. *Equisetum* tissue) by running water. Particularly  
375 stream-inhabiting chironomids (e.g. *Rheotanytarsus*, *Eukiefferiella*, *Rheocricotopus*), but also  
376 rheophilic diatom taxa (e.g. *Meridion circulare*, *Amphora pediculus*), indicate inflow of running water  
377 nearby the coring-site. *Corynocera ambigua*, which has a complex ecology (Brodersen and Lindegaard,  
378 1999), dominates the chironomid assemblage in the MIS 5a lake deposit in the Sokli basin. Nutrient  
379 levels were high both in the early Loitsana and Sokli lakes (*E. albipennis*-type, *C. plumosus*-type,  
380 *Stephanodiscus parvus*). This might have resulted from morphometric eutrophication (Hofmann 1998),  
381 i.e. a condition created by a reduction in lake volume. In our study, the latter occurred when the Sokli  
382 Ice Lake fully drained and lacustrine sedimentation continued in small depressions within the Sokli  
383 basin.

384

#### 385 4.3 MIS 3 ice-sheet dynamics

386

387 The glacial lake sequences of early Holocene, MIS 3 and MIS 5a age reveal remarkably similar  
388 assemblages and successions in aquatic and telmatic biota, however, some additional, special features  
389 are shown by the early MIS 3 glacial lake deposit. This most probably relates to different levels of ice-  
390 sheet dynamics upon deglaciation. First, the glacial lake sediment of early MIS 3 age is interlayered  
391 throughout the sequence with distinct clay laminae suggesting a continuing influence of the ice-sheet on  
392 lake deposition, even in the case of a distal position. Secondly, the sediment and fossil record seems to

393 register a glacier re-advance phase which intersected the shallow Glacial Lake Stage 3. The latter is  
394 demonstrated by the re-occurrence of deep water (increases in tycho-planktonic diatom taxa and the  
395 deep-water chironomid taxon *T. lugens*-type), a large lake-size (*Aulacoseira islandica*; Serieyssol et al.,  
396 2009), oligotrophic lake water conditions (*Cymbella aspera*; Krammer and Lange-Bertalot, 1986) and  
397 significant shore erosion (> 50 % of phytoliths) (Fig. 4B). Re-advances of the ice-margin are commonly  
398 recorded in glacial lake sediment sequences and are not necessarily climate-driven (Carrivick and  
399 Tweed, 2013). Striking also is the co-occurrence of peak abundances of *Procladius* with a large variety  
400 of algae (*B. braunii*, *Pediastrum*, *Tetraedon* cf. *minimum*, *Zygnema*, *Spirogyra*, T.225; high diatom-  
401 inferred Hill's N<sub>2</sub>) in the finely laminated clayey sediment at the end of Stage 2. This might reflect  
402 density stratification, a process typical for glacial lakes (Carrivick and Tweed, 2013, and references  
403 therein) in which sediment-loaded cold meltwater extends below warmer and more transparent river-fed  
404 water. The river-fed water probably favoured a diverse algal community. *Procladius*, which is abundant  
405 during this phase, is a ubiquitous chironomid taxon with a wide environmental tolerance including turbid  
406 water inflows (Greffard et al., 2012) and is found in lakes with a large variability in limnological  
407 conditions, such as hypolimnetic oxygen availability and water temperature (Brodersen et al., 2004). In  
408 addition to these features, the till bed, and overlying glacio-fluvial gravels and sands, at the base of the  
409 early MIS 3 deglacial sediment sequence shows large chunks of organic debris reworked from the  
410 Eemian Interglacial (MIS 5e) gyttja in the Sokli basin (Helmens et al., 2000). All these features together  
411 suggest active, warm-based ice in the course of early MIS 3 deglaciation.

412

#### 413 **4.4 Terrestrial vegetation in the ice-marginal environment during deglaciation**

414

415 Pollen and macrofossil remains found in the glacial lake sediments further allow a detailed  
416 reconstruction of terrestrial vegetation during the different episodes of deglaciation. Numerous seeds of  
417 birch trees and pollen percentage values for *Betula pubescens/pendula* of up to 80 % in the shallow  
418 Glacial Lake Stage 3 silts of early Holocene age (Fig. 4A) indicate the presence of birch forest on the  
419 land adjacent to the Sokli Ice Lake (Shala et al., 2017). Birch trees already bordered the glacial lake  
420 during Stage 2 (macrofossils present) and possibly Stage 1, i.e. in close proximity to the ice-margin.

421 Macrofossils are overall scarce in the latter sediments, probably due to the large lake size and long  
422 distance to shore. Moreover, the large contribution of grasses to the pollen sum, which for a major part  
423 might be derived from local shore habitats (high abundances of phytoliths), inhibits a precise pollen-  
424 based reconstruction of the terrestrial vegetation during Lake Stages 1 and 2.

425 *B. pubescens/pendula* and *Pinus* are represented by pollen values of 30-50 % and 20 %, respectively,  
426 in the glacial lake sediments of early MIS 3 and MIS 5a age (Fig. 4B-C). Whereas *B. pubescens/pendula*  
427 dominates the pollen assemblages in modern calibration samples from the sub-arctic birch forest in  
428 northern Fennoscandia, birch pollen percentages drop at the expense of pine pollen north of the birch  
429 tree-line (Aario, 1940; Prentice, 1978). The increase in representation of pine relative to birch occurs  
430 as the forest thins out, and *Pinus* pollen with its greater dispersal capability is being blown in from  
431 the south. Therefore, the modern calibration data suggests the presence of tundra vegetation on recently  
432 deglaciated terrain during early MIS 3 and 5a. According to Bos et al. (2009), the tundra vegetation  
433 during early MIS 3 appears to have been remarkably similar to the present-day tundra in northern  
434 Fennoscandia. Seeds of tree birch in the early MIS 5a glacial lake sediment indicate the presence of  
435 birch trees in the tundra vegetation, and *B. pubescens/pendula* pollen percentage values rising to 80 %  
436 in the overlying gyttja deposit registers the local establishment of birch forest. Pine forest with larch  
437 is recorded in the younger part of the MIS 5a gyttja deposit (N. Kuosmanen, unpublished data). Pollen  
438 and macrofossil data from interstadial organic-bearing sediment in northern Finland, dated to MIS 3,  
439 register birch forest and the local presence of pine trees (Sarala et al., 2016).

440 *Betula* spp. are fast immigrants due to the advantage of an abundant production of wind-dispersed  
441 fruits, rapid reproductive rates, fast growth rates and a young reproductive-maturity age (Birks, 1986).  
442 Furthermore, rapid migration of trees over the deglaciated land was probably facilitated by the northern  
443 location of glacial plant refugia (Väliranta et al., 2011; Tsuda et al., 2016). The existence of boreal tree  
444 populations in northerly small pockets of environmentally favorable conditions, in some cases close to  
445 the edge of the LGM ice-sheet, is being suggested by evidence from e.g. macrofossil charcoal assemblages  
446 and ancient DNA (Willis and van Andel, 2004; Parducci et al., 2012). However, while aquatic biota in  
447 the glacial lake show a mostly boreal character (see section 4.2), vegetation on land was sub-arctic in  
448 nature probably due to time needed for soil forming processes (Väliranta et al., 2015).



449 Interestingly also, our study shows that forested conditions were achieved in the ice-marginal  
450 environment during the early Holocene, whereas more open tundra vegetation seems to have  
451 characterized this environment during early MIS 3 and 5a. This might be the result of the relatively slow  
452 melting of the large LGM ice-sheet, allowing tree birch to migrate in congruent pace with the retreating  
453 ice-margin.

454

#### 455 **4.5 Climate during deglaciation**

456

457 There is overwhelming evidence in the fossil record from the early Holocene, MIS 3 and MIS 5a glacial  
458 lake sediments at Sokli for warm summers during deglaciation. This result is in accordance with high  
459 Milankovitch-forced summer insolation at high latitudes during all three time intervals (Berger and  
460 Loutre, 1991; Fig. 2).

461 Table 2 summarizes estimates of mean July air temperature ( $T_{jul}$ ) for the three deglaciation  
462 episodes based on macrofossil and chironomid remains. Aquatic and telmatic plant indicator taxa  
463 identified in the macrofossil records are given in the table together with (in parenthesis) lowest required  
464  $T_{jul}$  for current species presence in Finland (following Väiliranta et al., 2015; Shala et al., 2017).  
465 Additionally, chironomid taxa with warm temperature optima (i.e. warmer-than-today) are indicated;  
466 the optima are based on the distribution patterns of chironomid taxa in the Finnish chironomid-  
467 temperature calibration data-set (Luoto et al., 2014a, 2014b). The presence of these warm-indicating  
468 plant and chironomid taxa (Table 2), as well as the quantitative chironomid-inferred  $T_{jul}$  records for early  
469 MIS 3 (Engels et al., 2008), the early Holocene (Shala et al., 2017), and early MIS 5a (T. Luoto,  
470 unpublished data), indicate mean July air temperatures similar (13 °C) or higher-than-today during all  
471 three deglaciation episodes. Highest  $T_{jul}$  (ca. 15-15.5 °C) are inferred for the early Holocene deglacial  
472 episode (Table 2; Shala et al., 2017). Furthermore, the aquatic and littoral plant macrofossil assemblages  
473 encountered in the glacial lake sediments consistently suggest boreal conditions in the ice-marginal  
474 environment (Bos et al., 2009; Väiliranta et al., 2015; Shala et al., 2017; this study).

475 Warm summers in Fennoscandia during the early Holocene are also reported in e.g. Kullman  
476 (1998), Luoto et al. (2014b) and Paus and Haugland (2017), and during early MIS 3 in Sarala et al.

477 (2016). The latter is inferred from findings of fossil remains of the aquatic taxa *Sagittaria sagitifolia*  
478 (14.4 °C), *Nymphaea* and *Callitriche* in sediment dated to MIS 3. Quantitatively inferred climate  
479 parameters based on fossil remains predating the LGM are scarce, not only in Fennoscandia but also on  
480 the European mainland. Warm summers as well as warm winters are recorded by plant macrofossils and  
481 insect remains at Oerel (northern Germany) during early MIS 5a (Behre et al., 2005).

482

#### 483 **4.6 Comparisons with studies in North America**

484

485 One of the few detailed fossil analyses, including pollen, NPP's and macrofossils, on glacial lake  
486 sediments in North America is presented in Boyd et al. (2003). This study focuses on the terminal stages  
487 of Glacial Lake Hind, i.e. one of several interconnected proglacial lakes that formed across the Canadian  
488 prairies in front of the retreating margin of the Laurentide Ice Sheet in the early Holocene. Glacial Lake  
489 Hind discharged eastwards into Glacial Lake Agassiz (Fig. 1). Boyd et al. (2003) report a diverse  
490 macrofossil assemblage in glacial lake clays and silts, representing a high diversity of emergent and  
491 aquatic plants (e.g. various *Potamogeton* species such as narrow-leaved *P. filiformis*), and including  
492 abundant needles of *Picea* and a bud scale of *Populus*. Peak values for pollen of the wetland plant *Typha*  
493 in gyttja directly overlying the Glacial Lake Hind sediment indicates substantial postglacial warming  
494 (Boyd et al., 2003).

495 Early Holocene warm summers in response to Milankovitch-forced insolation have earlier been  
496 reported by Ritchie et al. (1983) based on pollen and macrofossil remains of *Typha* from 12 high-latitude  
497 sites in northern Canada dated to ca. 6-12 kyr BP and clustering at 10.5 kyr BP. The results from Canada  
498 are very similar to those obtained in the present study at Sokli indicating diverse aquatic biota, local  
499 presence of trees and warm summers along the retreating margin of the large N Hemisphere ice-sheets  
500 during deglaciation.

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#### 505 **4.7 Glacial lake sediments and carbon storage**

506

507 Although glacial lakes covered large parts of northern Europe and North America, contemporaneous  
508 glacial lakes generally were relatively small, confined to the deeper parts in the landscape. The glacial  
509 lakes went through complex evolutions as the topography changed and spill-ways opened-up along the  
510 retreating ice-margins (e.g. Dyke, 2004; Boyd et al., 2007; Johansson, 2007; Carrivick and Tweed,  
511 2013). This is exemplified by the study of Jansson (2003) which mapped a total of 26 glacial lakes (65  
512 sub-stages) in Labrador/Ungava, eastern Canada, during the early Holocene retreat of the Laurentide Ice  
513 Sheet.

514 Within the glacial lakes, protected bays were common features and, in combination with inflow  
515 of rivers draining unglaciated terrain, allowed for aquatic and telmatic biota to flourish in the lakes and  
516 along their shores (Boyd et al., 2003; Boyd, 2007; Bos et al., 2009; Helmens et al., 2009; Shala et al.,  
517 2014b; this study). Diverse biota was promoted by enhanced nutrient levels in the recently deglaciated  
518 terrain combined with insolation-induced warm summers. A close proximity to glacial plant refugia (e.g.  
519 Willis and van Andel, 2004) further facilitated the establishment of trees or open forest in the ice-  
520 marginal environment. Abundant macrofossils in the shallow glacial lake sediment of early Holocene  
521 age at Sokli, combined with the presence of birch forest in the catchment, led to an organic carbon  
522 content (LOI) in the sediment rising to 20 % (Fig. 4A; Shala et al., 2014a, 2014b). The sediment was  
523 found interlayered with thin organic laminae at a site close to the former glacial lake shore (Saunavuotso  
524 in Fig. 2B). This shows that postglacial carbon storage started within the glacial lake sediment, i.e. prior  
525 to the accumulation of peat that was initiated over large parts of central and eastern Canada (Dredge and  
526 Cowan, 1989) and northern Europe (e.g. Oksanen et al., 2001) due to the presence of the relatively  
527 impermeable glacial lake silts and clays. Although more regional studies are needed, the present study  
528 at Sokli reveals that aquatic biota and terrestrial vegetation quickly adapted to new environmental  
529 conditions during N-Hemisphere deglaciation and that glacial lake ecosystems, and the carbon stored in  
530 their sediments, should not be neglected in earth system modeling.

531

532

## 533 **5 Conclusions**

534

535 Our study shows that glacial lake sediments at the base of interglacial and interstadial deposits can  
536 provide important information on past environmental and climate conditions. These sediments have  
537 rarely been included in proxy-based palaeo-environmental studies, with the consequence that  
538 environmental and climate conditions during deglaciation of the large N-Hemisphere ice-sheets have  
539 largely remained unknown. Deglacial stages such as the LGM-Holocene transition are important  
540 testbeds in assessing climate sensitivity and the role of individual forcings and feedbacks in climate  
541 change (Shakun et al., 2012). In large parts of the N Hemisphere, the earliest evidence of the postglacial  
542 terrestrial and aquatic environments and climate are stored in glacial lake sediments. Our finding of  
543 distinct indicators of warm and productive conditions in the ice-marginal environment, immediately  
544 following deglaciation, is thus relevant for assessing the climate response and rates of ecosystem change  
545 during the critical glacial-interglacial transitions. Furthermore, the spatial extent and depth of glacial  
546 lake sediments, and their geochemistry, should be carefully assessed to quantify the carbon storage in  
547 glacial lake systems. Finally, paleo-environmental studies based on glacial lake sediments will benefit  
548 from studies in present-day ice-marginal environments, which, however, are currently scarce (Carrivick  
549 and Tweed, 2013).

550

## 551 **Author Contributions**

552

553 K.H. conceived the study and wrote the manuscript, with input from all authors. K.H., C.K., J.S.S., S.S.,  
554 J.A.A.B., S.E., N.K., T.P.L. and M.V. performed laboratory analyses and, together with J.R. and J.W.,  
555 made palaeo-environmental interpretations. A.O. made the glacial lake maps in Figure 1.

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561

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804 Table 1. AMS radiocarbon ages for the MIS 5a lake (gyttja) deposit at Sokli.

805

Borehole	Depth (m)	Material	Age (yr BP)	Laboratory No.
Sokli 2/2010	9.4	<i>Macrofossils of terrestrial plants (mostly seeds of birch)</i>	>47000	Poz-56337
Sokli 2/2010	9	<i>Macrofossils of terrestrial plants (mostly wood)</i>	>51000	Poz-56338

806

807

808 Table 2. Mean July air temperature ( $T_{jul}$ ) estimates based on macrofossils of aquatic/telmatic plants  
 809 (minimum  $T_{jul}$  values) and chironomids (optimum  $T_{jul}$  values) for the different deglaciation episodes at  
 810 Sokli. Present-day  $T_{jul}$  at Sokli is 13 °C.

811

Deglaciation episode	Minimum $T_{jul}$ (macrofossils)	Optimum $T_{jul}$ (chironomids)
Early Holocene	<i>Typha</i> (15.7 °C) <i>Glyceria lithuanica</i> (15.7 °C) <i>Callitriche cophocarpa/ hermaphroditica</i> (13.7-14.0 °C) narrow-leaved <i>Potamogeton</i> spp. such as <i>P. friesii</i> (13.6 °C) <i>Nymphaea</i> (13.5 °C)	<i>Cricotopus cylindraceus</i> -type (15.4 °C) <i>Cricotopus intersectus</i> -type (15.1 °C) <i>Chironomus anthracinus</i> -type (14.7 °C) <i>Polypedilum nubeculosum</i> -type (14.5 °C) <i>Microtendipes pedellus</i> -type (13.4 °C) <i>Psectrocladius sordidellus</i> -type (13.1 °C)
Early MIS 3	<i>Callitriche hermaphroditica</i> (14.0 °C) narrow-leaved <i>Potamogeton</i> spp. ( <i>P. friesii</i> ; 13.6 °C)	<i>Chironomus anthracinus</i> -type (14.7 °C) <i>Cladotanytarsus mancus</i> -type (14.4 °C) <i>Polypedilum nubeculosum</i> -type (14.5 °C) <i>Microtendipes pedellus</i> -type (13.4 °C)
Early MIS 5a	<i>Ceratophyllum</i> (14.1 °C) <i>Callitriche hermaphroditica</i> (14.0 °C) <i>Callitriche cophocarpa</i> (13.7 °C) narrow-leaved <i>Potamogeton</i> spp. ( <i>P. compressus</i> , <i>P. berchtoldii</i> ; 13.6 °C)	<i>Chironomus anthracinus</i> -type (14.7 °C) <i>Cladotanytarsus mancus</i> -type (14.4 °C) <i>Polypedilum nubeculosum</i> -type (14.5 °C) <i>Microtendipes pedellus</i> -type (13.4 °C) <i>Psectrocladius sordidellus</i> -type (13.1 °C)

812 **Figure captions**

813

814 Fig. 1. The extent of glacial lakes in northern Europe (**A**) and North America (**B**) during the last  
815 deglaciation around 10 kyr cal BP. **A** shows the maximum extent of the Ancylus Lake in the Baltic Sea  
816 basin at 10 kyr cal BP (Björck, 1995). The other lakes are time-transgressive. The extent of Ice Lakes  
817 along the western margin of the Fennoscandian Ice Sheet, and the glacial retreat chronology, are  
818 according to Stroeven et al. (2016, and references therein), whereas Ice Lakes along the northern and  
819 eastern margin of the ice-sheet follow Johansson (2007) and Ojala et al. (2013, and references therein),  
820 respectively. The Sokli Ice Lake which is the subject of this study is highlighted. **B** shows the total  
821 cumulative area covered by Glacial Lake Agassiz in North America in the time interval 13-8 kyr cal BP  
822 (Teller and Leverington, 2004).

823

824 Fig. 2. **A** New composite lithological column for the Late Quaternary sedimentary sequence in the Sokli  
825 basin with to the right the local stratigraphy according to Helmens (2014). The location within the Sokli  
826 basin of boreholes which are indicated along the left side of the lithological log is given in **B**. The  
827 correlation of the Sokli record with the marine global oxygen-isotope stack (Lisiecki and Raymo, 2005)  
828 and high-latitude July insolation (Berger and Loutre, 1991) (**C**) is based on stratigraphy and absolute  
829 dating evidence ( $^{14}\text{C}$ , OSL, TL and IRSL dates); absolute dates are indicated to the left of the lithological  
830 column in **A** (see text). Deglacial sediment intervals that are the subject of the present study, dated to  
831 the early Holocene, early MIS 3 and early MIS 5a, are highlighted by pink boxes.

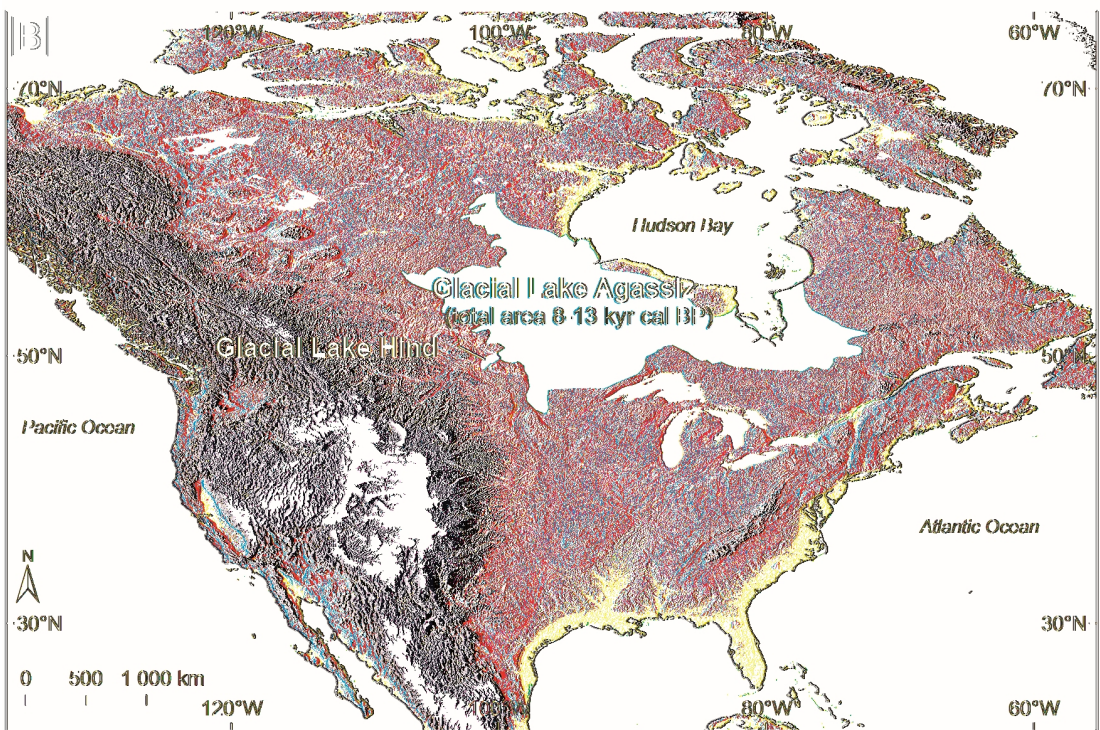
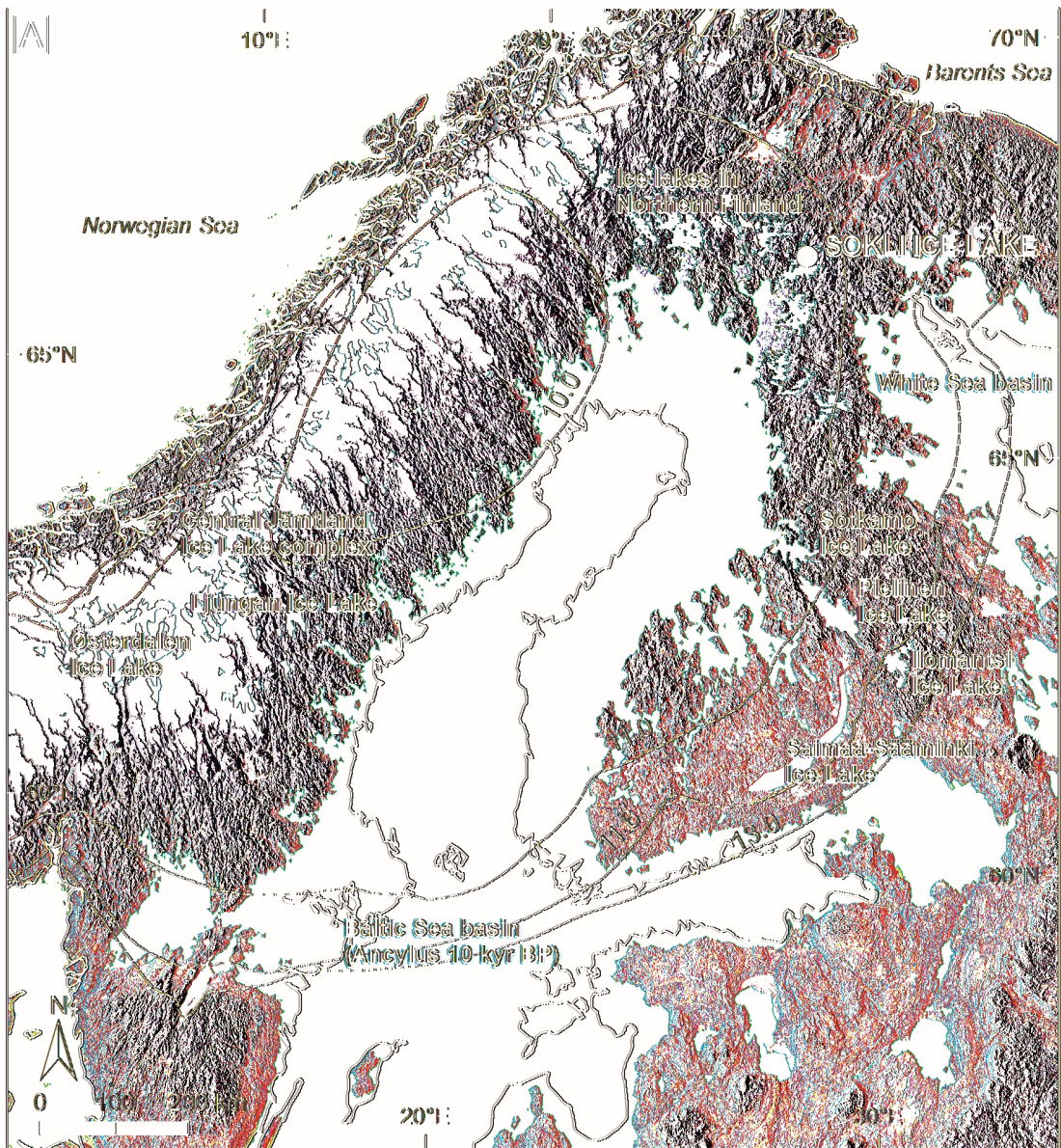
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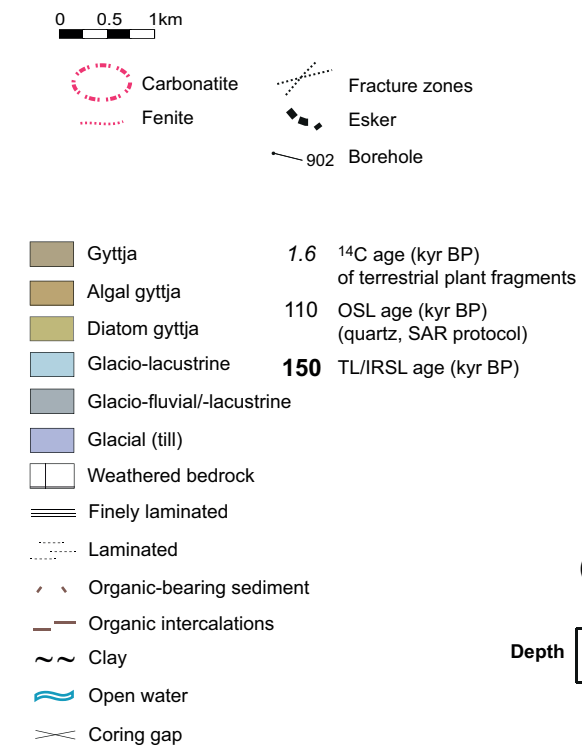
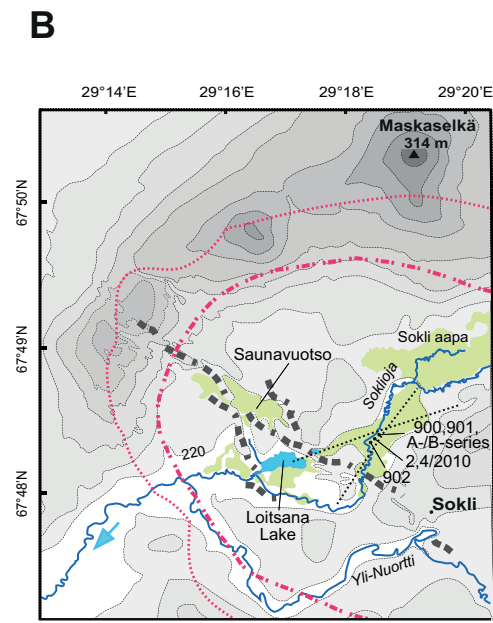
833 Fig. 3. Evolution of the later stages of the Sokli Ice Lake in the early Holocene; cross-sections show  
834 maximum water depths in Lake Loitsana (Shala et al., 2014a). Note that elevations are modern values.  
835 A similar glacial lake evolution occurred during the early MIS 3 and 5a deglaciation episodes. Lake  
836 Stage 1: deglaciation of the study site. Lake Stage 2: maximum spatial extent of the glacial lake. Lake  
837 Stage 3: partial drainage resulting in a smaller and more shallow glacial lake.

838

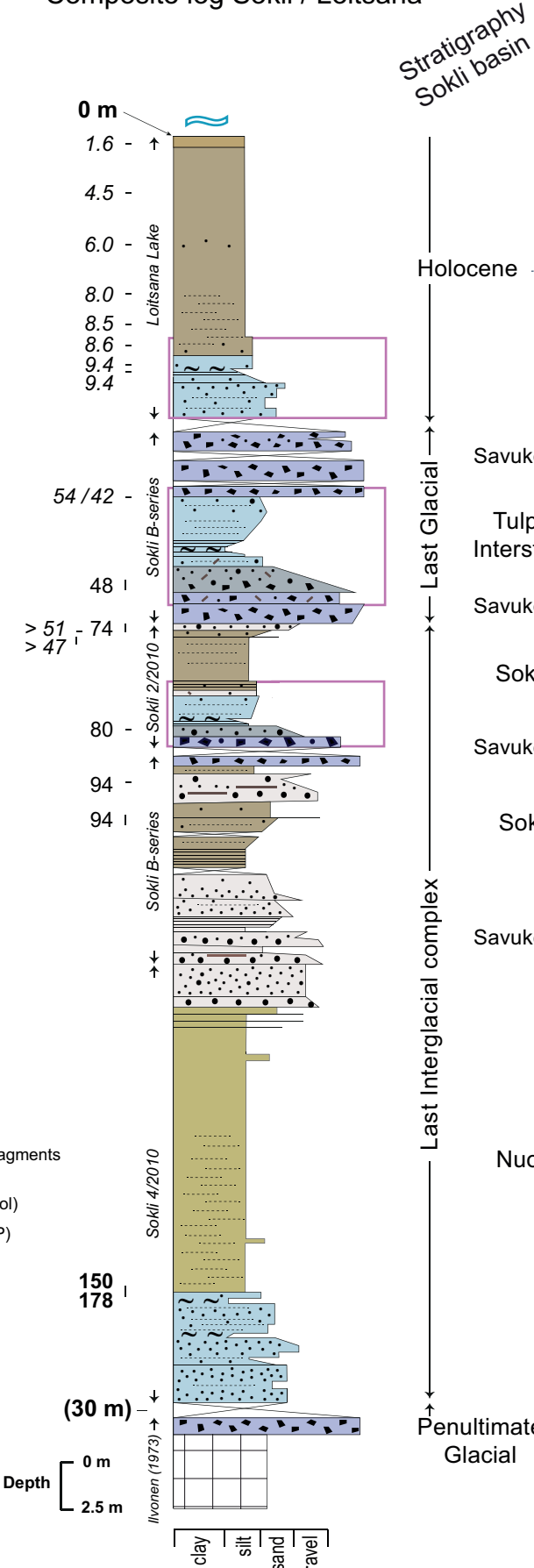
839 Fig. 4. Selection of proxy data for deglacial sediments in the Sokli basin dated to the early Holocene  
840 (A), early MIS 3 (B) and early MIS 5a (C). For legends see C. Phytoliths and Tertiary diatoms are  
841 calculated as percentages of the total sum of siliceous microfossils. Percentage values of pollen, spores  
842 and NPP's are based on the total sum of pollen of terrestrial plants. Based on A Shala et al. (2014a,  
843 2014b, 2017); B Engels et al. (2008), Bos et al. (2009), and Helmens et al. (2009); C this study.



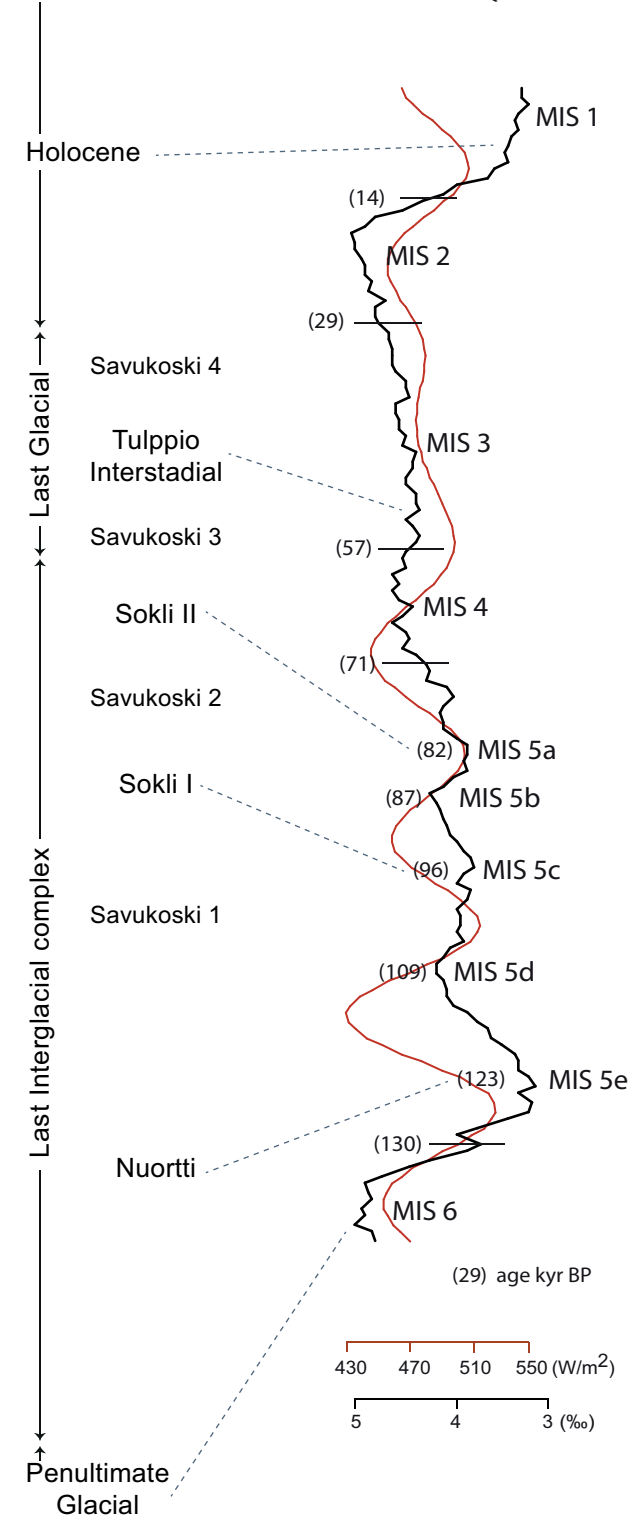




**A**  
Composite log Sokli / Loitsana



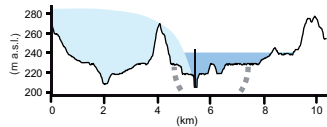
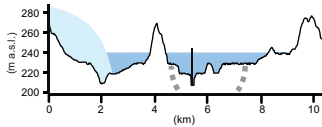
**C**  
Insolation July 65° N  
Benthic  $\delta^{18}O$  global stack (LR04)



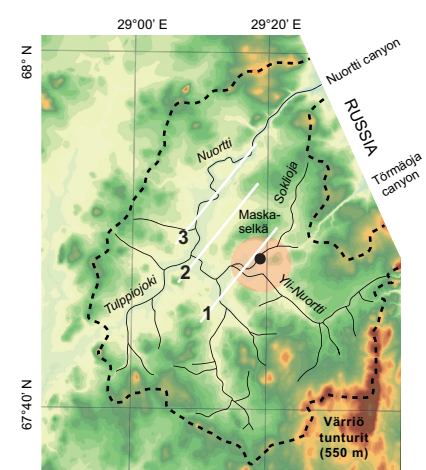
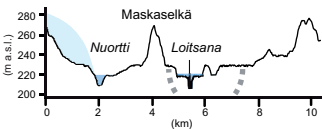
Maximum extension of glacial lake  
**LAKE STAGE 2**



Deglaciation of coring-site  
**LAKE STAGE 1**



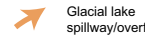
Glacial lake drainage  
**LAKE STAGE 3**



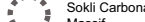
Retreating margin of the Fennoscandian Ice Sheet



Subglacial meltwater channel mouth



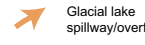
Esker



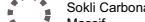
Sokli Ice Lake



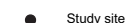
Glacial lake spillway/overflow



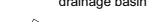
Sokli Carbonatite Massif



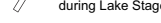
Study site



Nuortti River drainage basin



Position of ice front during Lake Stage 1



Sokli Carbonatite Massif



A (early Holocene)

