

1 **Regional Holocene climate and landscape changes recorded in the large subarctic lake**

2 **Torneträsk, N Fennoscandia**

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14 **Abstract**

15 Understanding the response of sensitive Arctic and subarctic landscapes to climate change is
16 essential to determine the risks of ongoing and projected climate warming. However, these
17 responses will not be uniform in terms of timing and magnitude across the landscape because of
18 site-specific differences in ecosystem susceptibility to climate forcing. Here we present a multi-
19 proxy analysis of a sediment record from the 330-km² lake Torneträsk to assess the sensitivity of
20 the Fennoscandian subarctic landscape to climate change over the past ~9500 years. By
21 comparing responses of this large-lake system to past climatic and environmental changes with
22 those in small lakes in its catchment, we assessed when the magnitude of change was sufficient to
23 affect an entire region rather than only specific sub-catchments that may be more sensitive to
24 localized environmental changes such as, e.g., tree-line dynamics. Our results show three periods
25 of regional landscape alteration with distinct change in sediment composition: i) landscape
26 development following deglaciation and through the Holocene Thermal Maximum, ~9500-3400
27 cal yr BP; ii) increased soil erosion during the Little Ice Age (LIA); and iii) rapid change during
28 the past century coincident with ongoing climate change. The gradual landscape development led
29 to successive changes in the lake sediment composition over several millennia, whereas climate
30 cooling during the late Holocene caused a rather abrupt shift occurring within ~100 years.
31 However, this shift at the onset of the LIA (~750 cal yr BP) occurred >2000 years later than the
32 first indications for climate cooling recorded in small lakes in the Torneträsk catchment,
33 suggesting that a critical ecosystem threshold was not crossed until the LIA. In contrast, the
34 ongoing response to recent climate change was immediate, emphasizing the unprecedented scale
35 of ongoing climate changes in subarctic Fennoscandia.

36

37 Keywords: Inorganic geochemistry; Soil erosion; Climate change; Oxygen and silicon isotopes;
38 Holocene; Scandinavia.

39

40 **1. Introduction**

41 Landscapes in the Arctic and subarctic are undergoing extensive transformations because of
42 ongoing climate change. In these landscapes, climate change leads to pronounced changes in land
43 cover and vegetation composition (e.g., Zhang et al., 2013) as well as to the degradation of
44 permafrost, which alters hydrology and slope stability (e.g., Hinzman et al., 2005). These
45 dynamics in turn cause climate feedbacks, particularly by affecting carbon cycling and release of
46 greenhouse gases (Pearson et al., 2013; Schuur et al., 2015). However, the scale of this
47 transformation is not uniform across northern landscapes (e.g., Elmendorf et al., 2012; Xu et al.,
48 2013).

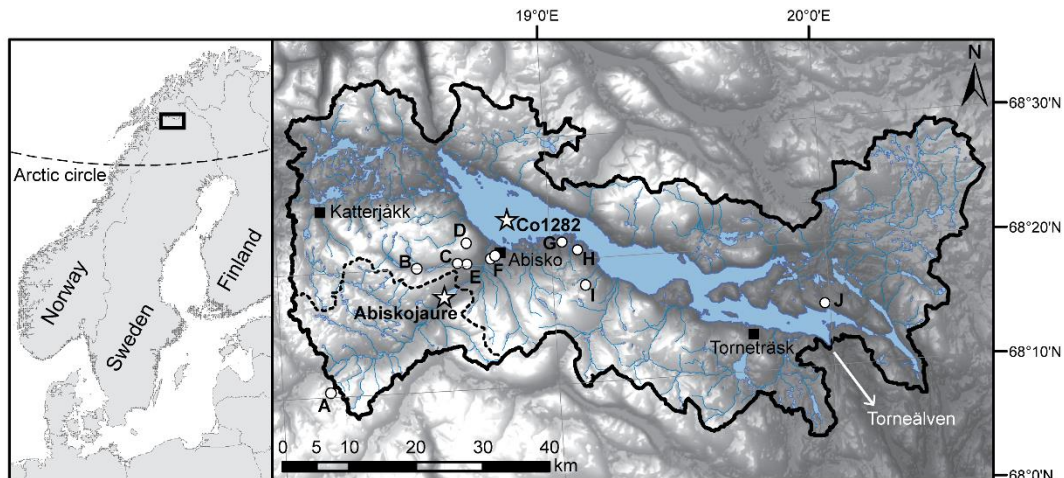
49 Lakes can respond rapidly to changes in climate forcing with regard to their physical (e.g., water
50 temperature, thermal stratification, ice cover duration), chemical (e.g., oxygen levels, carbon and
51 nutrient cycling), and biological (e.g., phenology, food web structure) characteristics (Williamson
52 et al., 2009). In addition to immediate in-lake responses, there are also responses in the aquatic
53 and surrounding terrestrial systems that occur over centuries to millennia. Lakes integrate
54 information on these long-term changes and may archive them in their sediments. Responses to
55 long-term changes, such as variations in vegetation cover, soil development or catchment erosion,
56 are non-linear because of complex forcing-response relationships. Furthermore, lakes may show
57 variable responses to climate changes because of differences in characteristics; for example,
58 small, shallow lakes with a low heat capacity may respond rapidly to small changes and thus

59 present higher-frequency noise in limnological and paleolimnological data (Adrian et al., 2009).
60 With increased lake and watershed size, initial small-scale changes in the watershed are recorded
61 at different scales in these cascading systems and potentially with time lags (cf., Dearing and
62 Jones, 2003). This lower sensitivity of larger drainage basins to local changes allows the
63 determination of the timing and character of important environmental tipping points, i.e., changes
64 that are of sufficient magnitude to affect an entire region rather than only specific areas of a
65 watershed that are particularly sensitive to climate/environmental changes, e.g., tree-line lakes.

66 In northernmost Sweden in the subarctic catchment of the large lake Torneträsk (“Torne Lake” in
67 English; surface area: 330 km²; watershed: 3350 km²; Fig. 1), the observed climate change during
68 the past century has thus far led to a mean annual temperature increase of 2.5°C, a shortened (~40
69 days) lake ice cover period (Callaghan et al., 2013), a reduced permafrost thickness (Åkerman
70 and Johansson, 2008), and an increase of wet areas in mires due to permafrost degradation
71 (Malmer et al., 2005). The Torneträsk area has been subject to numerous studies in different
72 disciplines focusing on a modern process understanding, including carbon cycling (e.g.,
73 Christensen et al., 2007; Karlsson et al., 2010), permafrost dynamics (e.g., Åkerman and
74 Johansson, 2008; Kokfelt et al., 2009), and sediment transport (e.g., Beylich et al., 2006). In
75 addition to these studies on recent processes, sediment records from several small lakes in the
76 catchment have been studied to reconstruct Holocene climate and environmental change with a
77 focus on vegetation development (e.g., Barnekow, 1999; 2000), quantitative climate
78 reconstructions (e.g., Bigler et al., 2002; 2003), sediment transport and erosion (e.g., Snowball
79 and Sandgren, 1996; Rubensdotter and Rosqvist, 2003) and changes in atmospheric circulation
80 patterns (e.g., Hammarlund et al., 2002; Rosqvist et al., 2007; Shemesh et al., 2001) (Fig. 1).

81 Because of the complexity and multitude of studies of small lakes conducted in the Torneträsk
82 catchment, a study of the lake itself represents an ideal site for investigating the integrated
83 regional climatic and environmental development in northernmost Sweden since the retreat of the
84 Fennoscandian ice sheet. Studies of Torneträsk's sediment record are thus far scarce and have
85 only focused on a general sedimentological description of the lacustrine sediment (Andrén et al.,
86 2002), recent changes in the input of terrestrial organic matter (OM) (Vonk et al., 2012) and on
87 depositional patterns and subaquatic topography (Vogel et al., 2013).

88 Using a combination of geochemical, isotopic and biological proxies, the aims of this multi-
89 proxy study are twofold. One aim was to survey the surface sediment geochemistry in order to
90 assess spatial variability, how this variability relates to catchment influences, and how signals
91 might vary across this 70-km-long lake. The second and main aim was to analyze a complete
92 sediment sequence from Torneträsk in order to determine the regional climate and environmental
93 development in northernmost Sweden during the Holocene and to assess i) how responses to
94 environmental change are recorded in the large lake system compared to small lakes that have
95 been studied within its catchment, and ii) to what extent the ongoing climate change has already
96 affected the lake ecosystem, and how the scale of any identified changes compares to changes
97 that occurred during past periods of distinct climate change such as the Holocene Thermal
98 Maximum (HTM) or the Little Ice Age (LIA).



99

100 Figure 1. (left) Map over Fennoscandia showing the location of the study region. (right)
 101 Elevation map over the watershed of Torneträsk (black contour line). Dashed contour line defines
 102 the watershed of Abiskojaure, filled squares indicate the location of three weather stations
 103 (Swedish Meteorological and Hydrological Institute) across the Torneträsk catchment, while
 104 open stars mark the sampling sites in Torneträsk (core Co1282) and Abiskojaure. Open circles
 105 indicate locations of paleolimnological studies on smaller lakes from the area (A: Vuolep
 106 Allakasjaure, B: Kårsa valley lakes, C: Pikkujärvi, D: Lake Njulla, E: Lake Tibetanus, F: Vuolep
 107 Njakajaure and Badsjön, G: Villasjön and Inre Harrsjön, H: Vuoskkujávri, I: Lake 850/865, J:
 108 Lake Latteluokta. The white arrow highlights the outlet of Torneträsk, Torneälven, in the SE.

109

110 2. Regional setting

111 Torneträsk (68° 29'-68° 11' N, 20° 01'-18° 36' E; 341 m a.s.l.) is 70 km long (NW-SE),
 112 maximally 10 km wide (SW-NE), has a surface area of 330 km² and a watershed of 3350 km²
 113 (Fig. 1). The lake has an average depth of 53 m and a maximum depth of 168 m (based on
 114 bathymetric measurements in 1920/1921, Abisko Research Station; Fig. 2), with an estimated

115 water residence time of c. 8.5 years (Swedish Meteorological and Hydrological Institute (SMHI);
116 <https://vattenwebb.smhi.se/>). The watershed is drained by several smaller streams and rivers of
117 which the Abiskojärva, entering the lake west of the village of Abisko, is the largest inlet with an
118 average discharge of $14 \text{ m}^3 \cdot \text{s}^{-1}$ (Fig. 1). Torneälven (“Torne River” in English) drains Torneträsk
119 to the SE with an average discharge of $65 \text{ m}^3 \cdot \text{s}^{-1}$ (1999–2013; SMHI; <https://vattenwebb.smhi.se/>)
120 (Fig. 1). The lake is (ultra)oligotrophic ($\text{TN} = 195 \pm 25 \text{ } \mu\text{g} \cdot \text{l}^{-1}$, $\text{TP} = 3 \pm 2 \text{ } \mu\text{g} \cdot \text{l}^{-1}$, $\text{TOC} = 1.3 \pm 0.5$
121 $\text{mg} \cdot \text{l}^{-1}$), circumneutral ($\text{pH} = 7.3 \pm 0.1$) (environmental monitoring data (MVM);
122 <http://www.slu.se/miljodata-MVM>), and $\delta^{18}\text{O}$ and $\delta^2\text{H}$ of water samples from Torneträsk were –
123 $12.7/-12.8\text{‰}$ and $-92.6/-93.0\text{‰}$, respectively in August 1999 (Shemesh et al., 2001).

124 The complex bedrock geology in the Torneträsk area consists of Archean plutonites (mainly
125 granite and syenite) and their metamorphic products that are overthrust by Caledonian
126 metasediments and amphibolites (Swedish Geological Survey (SGU) database;
127 maps2.sgu.se/kartgenerator/maporder_sv.html). During the retreat of the Weichselian ice sheet,
128 the NW-SE trending Torneträsk depression hosted one/multiple ice-dammed lakes with water
129 levels up to 250 m higher than today as suggested by paleoshoreline and paleodelta deposits
130 (Gretener and Stromquist, 1981; Melander 1977). The area was largely deglaciated around 9500
131 cal yr BP (glaciers occupy ~0.5% of the watershed today (SMHI; <https://vattenwebb.smhi.se/>)),
132 leading to the drainage of the precursor lakes and the likely establishment of the present
133 shoreline. This approximate minimum deglaciation age is suggested by several radiocarbon-dated
134 sediments from smaller lakes within Torneträsk’s catchment (e.g., Barnekow et al., 1998; Bigler
135 et al., 2003; Shemesh et al., 2001) as well as by cosmogenic radionuclide dating of glacially
136 scoured surfaces and erratics from the area (Stroeven et al., 2002). Bedrock is exposed or only
137 covered by thin soils (<1m) in large parts of the catchment (~43%; SMHI;

138 <https://vattenwebb.smhi.se/>), particularly at higher elevations in the mountains overlooking
139 western Torneträsk, whereas glacial, glaciofluvial and fluvial deposits occur in confined areas at
140 lower elevations. In lower lying areas, mires have developed in local depressions (Melander
141 1977; SGU database; maps2.sgu.se/kartgenerator/maporder_sv.html).

142 The climate in the Torneträsk area is generally oceanic but has a pronounced oceanic-continental
143 gradient from W to E that is enhanced by the strong orographic effect of the Scandes Mountains
144 (Table 1). Ice cover on the lake usually occurs from December/January to May/June. The
145 Torneträsk area is located in the zone of discontinuous permafrost and catchment vegetation
146 consists of alpine tundra above the present tree-line at 600–800 m a.s.l. and subalpine birch to
147 northern boreal pine-birch forests below tree-line, with a total forest cover of ~18% across the
148 watershed (SMHI; <https://vattenwebb.smhi.se/>). Pine has a continuous distribution in the SE part
149 of the Torneträsk catchment, whereas pine occurs only sporadically in the western part of the
150 catchment below 450 m a.s.l. Dwarf shrubs dominate the field layer vegetation together with
151 grasses, sedges and herbs (Barnekow and Sandgren, 2001).

152 In addition to sediments from Torneträsk, we also recovered and analyzed a sediment sequence
153 from Abiskojaure (68° 18' N, 18° 36' E; 488 m a.s.l.; Fig. 1) to provide a more comprehensive
154 view on changes in catchment dynamics and aid in constraining the chronology of core Co1282
155 by cross-correlation (see 4.2.2 Chronology). Abiskojaure's watershed, located in the SW
156 watershed of Torneträsk, is one of the main sub-watersheds (368 km²) and drains via
157 Abiskojákka into the lake. The lake has a residence time of c. 0.2 years, a surface area of 2.8 km²,
158 an average and max. depth of 16.4 m and 35 m, respectively (SMHI;
159 <https://vattenwebb.smhi.se/>). Birch forest occurs along the river corridor up to ~700 m a.s.l. and
160 covers <10% of the watershed, while most of the watershed consists of sparsely vegetated alpine

161 tundra with thin soils (<1m) or unvegetated areas with exposed bedrock (~78% of the watershed).
 162 Glaciers cover about 1% of Abiskojaure's watershed (SMHI; <https://vattenwebb.smhi.se/>; SGU
 163 database; maps2.sgu.se/kartgenerator/maporder_sv.html).
 164 Earlier paleoecological studies in the Torneträsk watershed did not identify distinguishable
 165 environmental impacts in response to human activities, which were likely minor and limited to
 166 fishing and hunting up to the 17th century when more intensive reindeer husbandry started to
 167 develop (Emanuelsson, 1987). During the past century, infrastructure developments in the area
 168 included railroad (built AD 1898-1902) and highway (built AD 1980-1984) construction along
 169 the southern shore of the lake.

170
 171 Table 1. Meteorological data for three weather stations across the Torneträsk watershed (West to
 172 East) for the period 1961-1990
 173 (<http://www.smhi.se/klimatdata/meteorologi/temperatur/dataserier-med-normalvarden-1.7354>).
 174 Winter is defined as the period October to April and summer as the period May to September.

Station		Katterjåkk	Abisko	Torneträsk
Latitude		68° 25'N	68° 21'N	68° 13'N
Longitude		18° 10'E	18° 49'E	19° 43'E
Altitude (m)		500	388	393
Precipitation (mm)	Annual	844	304	476
	Summer	340	157	268
	Winter	504	147	208
Temperature (°C)	Annual	-1.7	-0.8	-1.0
	Summer	+6.4	+7.5	+7.9
	Winter	-7.5	-6.8	-7.4

175

176 3. Material and methods

177 3.1 Sampling

178 Surface samples and a composite sediment core (Co1282) from Torneträsk were recovered from
179 the ice-covered lake in April 2012. The surface samples were collected from 43 sites across the
180 lake using a gravity corer (Fig. 2). The uppermost 2 cm were subsampled on site and stored in a
181 freezer. Core Co1282 was sampled in the western lake basin at a water depth of 130 m (68°
182 23'40"N, 18° 50'55"E; Fig. 2). The sampling site was selected based on hydroacoustic data, which
183 indicated the presence of acoustically stratified, horizontally continuous and extensive high and
184 low amplitude reflections that are indicative of continuous hemi-pelagic sedimentation (Vogel et
185 al., 2013) (Fig. S1). We recovered a 18.2-cm-long gravity core and a 181.2-cm-long percussion
186 piston core at the coring site Co1282 using UWITEC equipment. The overlapping cores were
187 correlated based on macroscopic lithological features such as red iron oxide and black manganese
188 oxide layers and supported by geochemistry (major element concentrations, Si/Al and Pb/Ti
189 ratios), which yielded a total composite core length of 188.2 cm. The sediment sequence from
190 Abiskojaure was recovered in April 2016 using a gravity and piston corer and yielded a
191 composite core length of 217.5 cm. Cores were stored at 4°C until subsampling in the laboratory.

192 3.2. Age-depth modeling

193 Several techniques were employed to date the composite core from Torneträsk, including ^{210}Pb ,
194 accelerator mass spectroscopy (AMS) radiocarbon, optically stimulated luminescence (OSL)
195 dating, and, after difficulties with radiocarbon and OSL dating of the Torneträsk sediment, cross-
196 correlating significant geochemical changes in the record to those recorded in the AMS
197 macrofossil radiocarbon-dated sediment record from Abiskojaure. Near-surface sediments were
198 radiometrically dated by analyzing ^{210}Pb , ^{226}Ra , ^{137}Cs and ^{241}Am by direct gamma assay in the
199 Environmental Radiometric Facility at University College London. The resulting ^{210}Pb

200 chronology was calculated using the constant rate of ^{210}Pb supply (CRS) dating model (Appleby,
201 2001).

202 Terrestrial plant macrofossils are absent in core Co1282; thus AMS radiocarbon ages were
203 determined on six bulk sediment samples at the LARA radiocarbon dating facilities of the
204 University of Bern. Three plant macrofossils from the sediment record from Abiskojaure yielded
205 sufficient C for AMS radiocarbon dating (Beta Analytic INC) (Table 2). All ages are reported as
206 calibrated ages BP, i.e., before AD 1950 (cal yr BP). OSL dating was conducted on unexposed
207 fractions of 10 cm half-core rounds in the OSL laboratory of the Institute of Geological Sciences
208 at the University of Bern. Only polymineral fine grains were measured as the quartz fraction did
209 not yield a luminescence signal. Both sample preparation and measurement were carried out as
210 reported in Lowick et al. (2015), as was dose rate determination.

211 We used the IntCal13 calibration curve (Reimer et al., 2013) for calibration of radiocarbon ages
212 and the Bayesian age-depth modeling software Bacon 2.2. (Blaauw and Christen, 2011) to
213 determine the age-depth relationship in the sediment cores. The technique estimates accumulation
214 rates by using Markov Chain Monte Carlo simulations and models age uncertainties for each
215 sample interval in the core (Blaauw and Christen, 2011).

216 3.3. Geochemistry

217 Prior to analyses, all sediment samples were freeze-dried and ground using a swing mill for
218 samples from Abiskojaure and core Co1282 and a planetary mill for the surface samples,
219 respectively. Major and trace element geochemistry was measured on 0.2 g sample material by
220 wavelength dispersive X-ray fluorescence using a Bruker S8 Tiger spectrometer equipped with
221 an Rh anticathode X-ray tube. Precision was within $\pm 3\%$ for all elements used in this study
222 except for Zr ($\pm 6\%$), and accuracy was within $\pm 5\%$ except for Na ($\pm 10\%$), Mg ($\pm 8\%$) and Zr

223 ($\pm 6\%$) (Rydberg, 2014). To infer compositional changes in the minerogenic fraction related to the
224 degree of weathering, we used the K/Al ratio and the more comprehensive ratio
225 $(\text{Na}_2\text{O}+\text{MgO}+\text{K}_2\text{O}+\text{CaO})/\text{TiO}_2$. These ratios can provide information about the weathering
226 degree of the detrital siliciclastics because the comparatively more mobile alkali and alkaline
227 earth elements are preferentially leached from aluminosilicates compared with immobile Al or Ti
228 during chemical weathering (cf., Kauppila and Salonen, 1997; Parker, 1970; Roy et al. 2008).
229 The ratio $(\text{Na}_2\text{O}+\text{MgO}+\text{K}_2\text{O}+\text{CaO})/\text{TiO}_2$ is hereafter referred to as the Chemical Index (CI),
230 where lower values would indicate a more intensive chemical weathering. However, changes in
231 these ratios must be interpreted with caution because they can also be influenced by, e.g., changes
232 in grain-size or source of the deposited material (Boyle 2001). To assess potential concurrent
233 changes in grain-size or source material and their potential influence on the CI, we used K/Rb
234 and Zr/Ti ratios. K/Rb can provide information about grain-size changes because K is enriched in
235 feldspars, which occur primarily in the coarser grain fraction, while Rb is common in micas and
236 clays that are more abundant in the fine fraction. Lower ratios indicate finer-grained and higher
237 ratios coarser-grained material (Kylander et al. 2013). Zr/Ti varies depending on the type of
238 parent material, and has thus been used to identify possible changes in the source of material
239 when assuming equal mobility of both elements during low-temperature weathering (Fabel et al.
240 2006; Kylander et al. 2013; Law et al. 1991; Reynolds et al. 2004).

241 Biogenic silica (bSi) in the sediment of Torneträsk was determined by Fourier transform infrared
242 (FTIR) spectroscopy using a Bruker Vertex 70 equipped with a MCT (mercury-cadmium-
243 telluride) detector, a KBr beam splitter, and a HTS-XT accessory unit (multi-sampler). 11 mg of
244 sample material was mixed with 0.5 g of oven-dried spectroscopic grade potassium bromide
245 (KBr) for the analysis. Quantifications are derived from a PLSR calibration model based on

246 synthetic sediment mixtures with defined bSi content. For more specific details about the method,
247 analytical procedure and instrument setup see Meyer-Jacob et al. (2014a). Total carbon (TC),
248 total nitrogen (TN) (from which we calculated atomic C/N- ratios), $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ were analyzed
249 using a Flash EA 2000 elemental analyzer coupled to an isotope ratio mass spectrometer (Thermo
250 Fisher Scientific). Samples were not treated with HCl prior to elemental analysis because FTIR
251 analyses showed that the samples did not contain inorganic carbon/carbonates, which have very
252 distinct FTIR spectral characteristics (cf., Rosén et al. 2011, Vogel et al. 2008). TC is therefore
253 considered as equivalent to total organic carbon (TOC). Element and bSi concentrations are
254 expressed as percentage by weight. Accumulation rates of bSi (AR_{bSi} ; $\text{g}\cdot\text{m}^{-2}\cdot\text{yr}^{-1}$) were calculated
255 according to Eq. (1):

$$256 \quad \text{AR}_{\text{bSi}} = \text{SR} \times \text{DBD} \times \% \text{bSi} \times 10^2, \quad (1)$$

257 where SR is the sedimentation rate ($\text{cm}\cdot\text{yr}^{-1}$), DBD is the dry bulk density ($\text{g}\cdot\text{cm}^{-3}$), and %bSi is
258 the bSi concentration of the sample.

259 Si/Al ratios, indicative of relative changes in bSi content (Peinerud et al. 2001), were used to
260 facilitate the cross-correlation of the records from Torneträsk and Abiskojaure in which bSi was
261 not quantified. In Torneträsk the correlation between bSi and Si/Al ratios is $R^2=0.83$.

262 3.4. Diatom analysis

263 Diatom slides were prepared using standard methods including digestion with 30% H_2O_2 as
264 described in Battarbee et al. (2001). Rinsed diatom samples were dried on coverslips and
265 permanently mounted onto microscope slides using Naphrax. Diatom identification followed
266 mainly Krammer and Lange-Bertalot (1986-91) and the guidelines of the Surface Water

267 Acidification Programme (SWAP) (Stevenson et al., 1991). For stratigraphical display, diatom
268 counts are expressed as relative abundances.

269 3.5. Oxygen and silicon isotope analyses of diatom silica

270 Diatom samples for isotope analyses were purified following the cleaning stages described in
271 Morley et al. (2004), which include removal of OM, sieving, differential settling and density
272 separation. Prior to analyses, diatom sample purity was assured by visual inspection via SEM and
273 by micro XRF measurements. $\delta^{18}\text{O}$ and $\delta^{30}\text{Si}$ were analyzed using a step-wise fluorination
274 method. The outer hydrous layer of the diatoms was removed in a pre-fluorination stage using a
275 BrF_5 reagent at low temperature. This was followed by a full reaction at high temperature to
276 liberate oxygen (that was converted to CO_2 and measured for $\delta^{18}\text{O}_{\text{diatom}}$) and silicon (collected as
277 SiF_4) isotopes that were measured using a MAT 253 dual-inlet mass spectrometer. $\delta^{18}\text{O}$ values
278 were converted to the VSMOW scale and $\delta^{30}\text{Si}$ values were converted to the NBS28 scale, both
279 using the within-run laboratory standard BFC_{mod} .

280 3.6. Interpolation of spatial variability

281 We used inverse distance weighted (IDW) interpolation in the ArcGis 10.2.1 software to
282 interpolate and visualize the spatial variability of TOC, C/N ratios, bSi and the CI across
283 Torneträsk.

284

285 **4. Results and discussion**

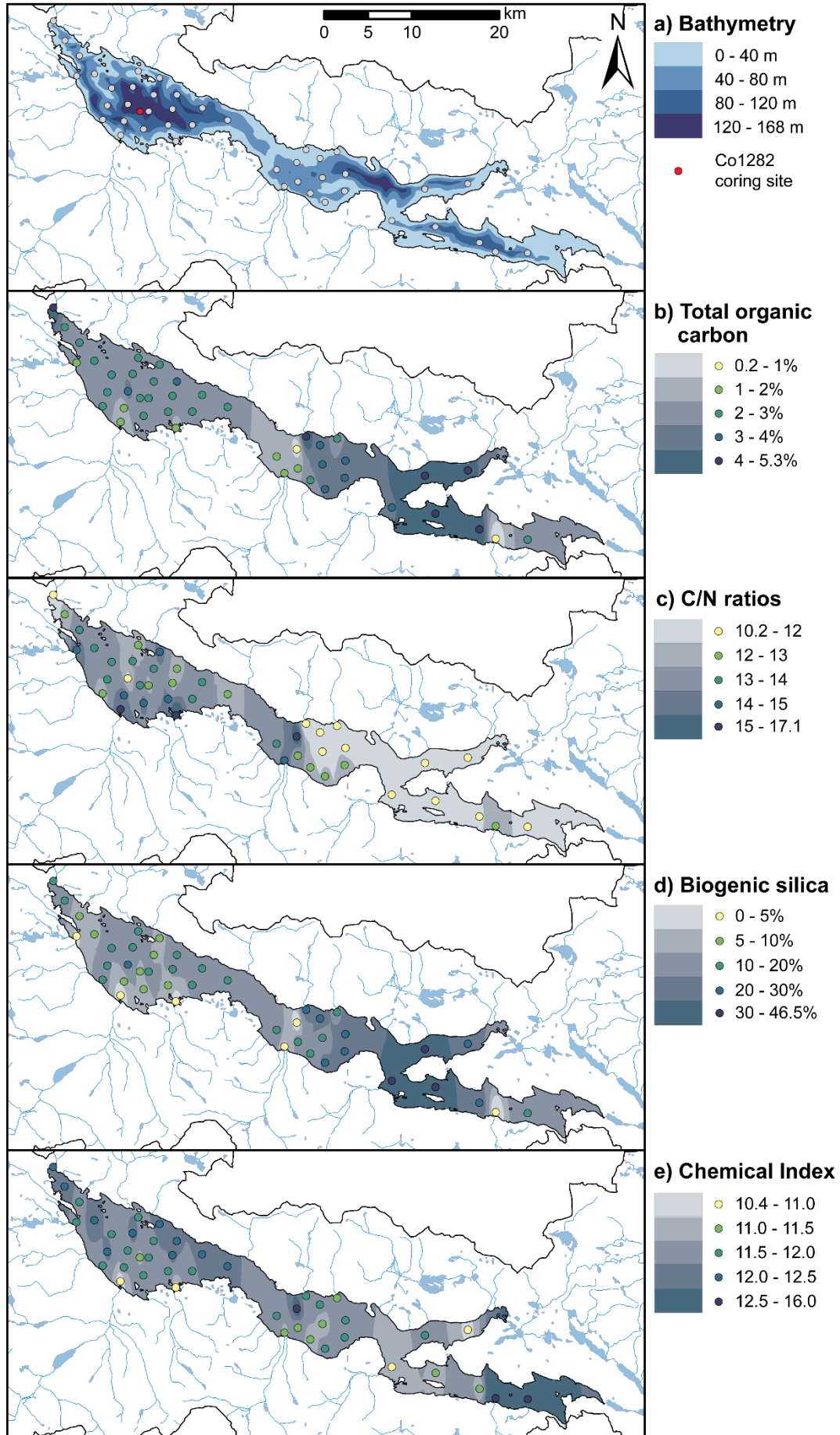
286 4.1. Spatial variability of recent sediment composition

287 TOC, atomic C/N ratios, bSi and the CI are shown in Figure 2 to exemplify the spatial variability
288 in the surface sediments of Torneträsk. TOC concentrations are generally low, ranging from 0.2

289 to 5.3% (mean: 2.7%). C/N ratios vary from 10.2 to 17.1 (mean: 12.9), which suggests both
290 aquatic and terrestrial OM (Meyers and Ishiwatari, 1993). bSi and the CI exhibit strong spatial
291 variability with values in the ranges of 0–46.5% (mean: 14.6%) and 10.4–16.0 (mean: 11.9),
292 respectively. Variations in the surface (i.e., recent) sediment composition are mainly controlled
293 by i) the proximity to riverine input, ii) differences in catchment characteristics that result mainly
294 from the contrasting landscapes in the western and eastern parts of the catchment, and iii) the
295 heterogeneous bedrock geology. Close to river inlets such as Abiskoječåkka, terrestrial-derived OM
296 explains elevated C/N ratios and an increased riverine input of clastic material causes a dilution
297 of TOC and bSi concentrations. Hydroacoustic measurements in Tornetråsk indicate that
298 sediment accumulation is up to 8 times higher in proximity to major river inlets compared to rates
299 in distal areas; for example, the sediment thickness is >8 m near the Abiskoječåkka inlet but less
300 than 2 m in more distal parts of the western basin (Vogel et al., 2013).

301 In addition to proximity to river inlets, TOC, C/N ratios and bSi show differences from W to E
302 across Tornetråsk. When divided into a western (n = 29) and an eastern (n = 15) subset, mean
303 TOC, C/N ratios and bSi values differ significantly between lake basins with 2.4 versus 3.4%
304 TOC, C/N ratios of 13.5 versus 11.7, and 10.1 versus 23.3% bSi in the western and eastern half,
305 respectively. This pattern can be explained by the likely different style and intensity of erosion
306 and weathering in the western compared to the eastern watershed of Tornetråsk. In the western
307 watershed, higher relief energies, large areas of exposed bedrock and sparser vegetation cover
308 would promote the erosion of chemically less-matured substrates. Similarly, low rates of
309 microbially catalyzed weathering processes and OM turnover in soils would limit the release of
310 nutrients and dissolved load. In contrast, the eastern catchment, with lower relief energies, thicker
311 soils, a denser/more productive vegetation cover and more extensive peatlands, would favor an

312 enhanced release of nutrients/dissolved load. This is supported by several studies that show
313 increased silicate weathering rates and inorganic P release with increasing vegetation cover in
314 previously glaciated regions (Anderson et al., 2000; Humborg et al., 2004; Moulton et al., 2000).
315 These differences in catchment characteristics may explain the contrasting patterns of sediment
316 composition indicated by the proxies for aquatic production (bSi, TOC) and organic matter
317 source (i.e., C/N). The CI does not show a distinct spatial pattern in Torneträsk (Fig. 2), because
318 the mineral composition within the lake basin reflects not only the degree of weathering but also
319 the heterogeneity of the bedrock geology in the Torneträsk area.



321 Figure 2a) Bathymetric map of Torneträsk with 40 m contour intervals based on bathymetric
322 measurements from 1920/1921. b-e) Spatial variability of total organic carbon (TOC), C/N ratios,
323 biogenic silica (bSi), and the chemical index (CI) in the surface sediment (0-2 cm) of Torneträsk.
324 Color-coded dots indicate site-specific values, whereas contour lines were interpolated using
325 IDW interpolation in ArcGis 10.2.1 and indicate interpolated ranges.

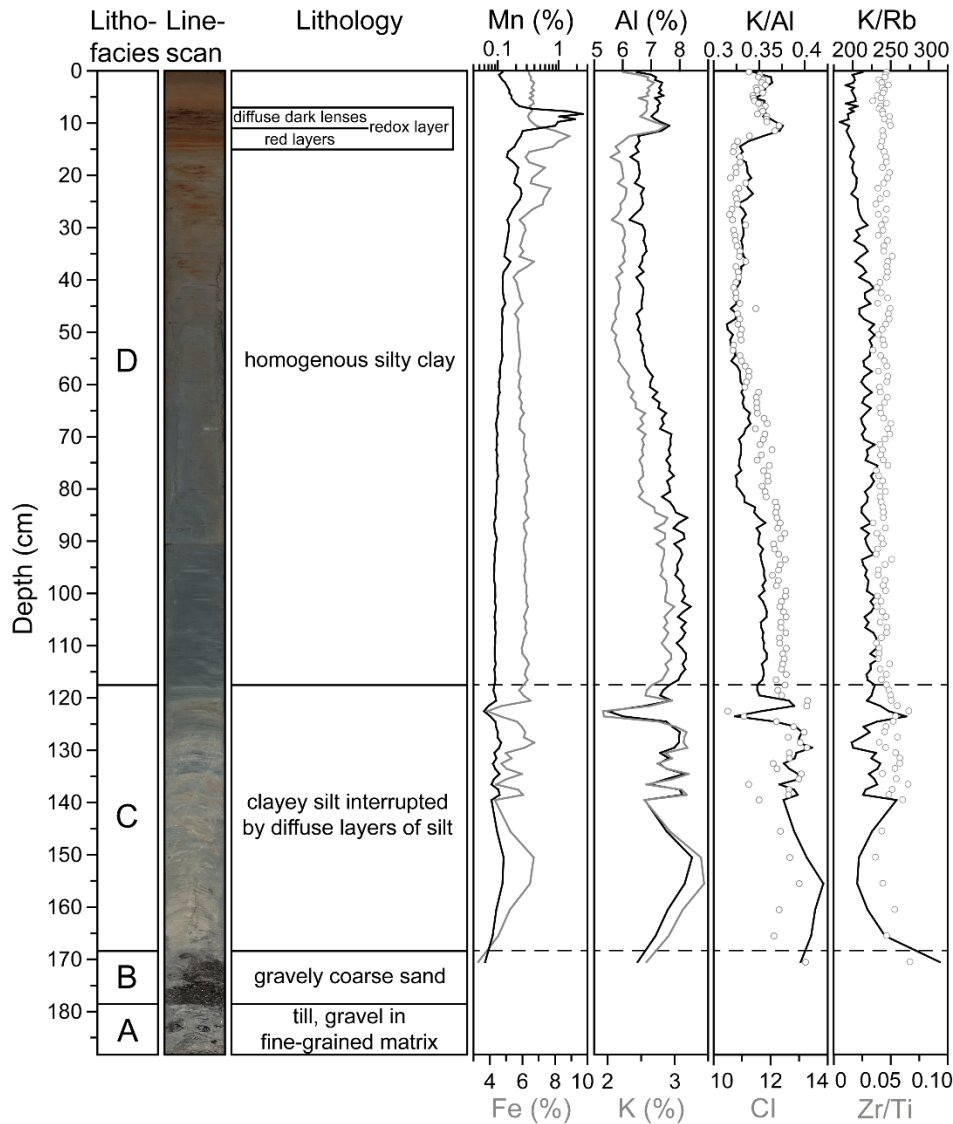
326

327 4.2 Long-term variations (core Co1282)

328 4.2.1 Lithology

329 The 188.2-cm-long sediment core Co1282 comprises a record of the depositional history in the
330 western Torneträsk basin from the end of the Weichselian glaciation to the onset of lacustrine
331 sedimentation until the present. The sequence can be subdivided into four lithofacies. The basal
332 lithofacies A (188–178 cm blf; Fig. 3) is characterized by a grey, massive and dense matrix-
333 supported diamicton with gravel and sand in a clay matrix. Due to its specific characteristics,
334 particularly the high density of the material in combination with the glacial setting, we
335 interpret the diamicton as a basal moraine deposit/lodgement till. The diamicton is overlain by
336 lithofacies B (178–168 cm blf; Fig. 3), which is composed of unconsolidated and poorly sorted
337 gravely-coarse sand and thus likely of glacio-fluvial origin. The following lithofacies C (168–117
338 cm blf; Fig. 3) appears thinly bedded and consists of light-grey clayey silt that is interrupted by
339 normally graded layers of slightly coarser material (silt). This facies was likely deposited in a
340 distal proglacial environment during the ice-dammed lake phase in the Torneträsk basin. The
341 rhythmical lamination of the fine-grained sediments would result from variations in the
342 sedimentation regime (suspension settling) due to fluctuations in intensity and sediment load of
343 meltwater discharge coming from the retreating ice sheet (Carrivick and Tweed, 2013).

344 Lithofacies D (117–0 cm blf; Fig. 3) is characterized by homogenous silty clay with a color
345 change from dark grey at the bottom to brown-grey towards the surface. Deposition of lithofacies
346 D commenced after the retreat of the Fennoscandian ice sheet from the area and represents the
347 continuous hemipelagic lacustrine sedimentation in Torneträsk during the Holocene. Between
348 15–7 cm blf, distinct horizons occur that consist of Fe-rich (up to 8.9%) reddish layers overlain
349 by diffuse dark lenses rich in Mn (up to 2.5%; Fig. 3). Similar layers have been reported for
350 sediments of other large, oligotrophic lakes, such as Lake Baikal where they have been ascribed
351 to enrichment of redox sensitive elements at the redox front in low sedimentation rate settings
352 (Och et al., 2012).



353
 354 Figure 3. Lithofacies, color linescan, lithological description, and concentrations of Mn, Fe, Al,
 355 and K, as well as the Chemical Index (CI) and K/Al, K/Rb, and Zr/Ti ratios for core Co1282,
 356 recovered from the western basin of Torneträsk.

357

358 4.2.2 Chronology

359 The chronology of core Co1282 from Torneträsk is constrained by ^{210}Pb dating as well as cross-
 360 correlating significant geochemical changes to those in the AMS radiocarbon-dated sediment

361 record from Abiskojaure (Figs. 4 and 5). Total ^{210}Pb activities in the sediment reached
362 equilibrium with the background (supported) ^{210}Pb activity at 1.5–2 cm, which indicates a low
363 sedimentation rate of $\sim 0.015 \text{ cm}\cdot\text{yr}^{-1}$ in the near-surface sediments (Fig. 4a, b).

364 Radiocarbon measurements on bulk OC in core Co1282 suggest a large and variable reservoir/old
365 carbon effect for bulk OC accumulated in the sediments of Torneträsk (Table 2). The surface
366 sample (0–1 cm) and a near-surface sample (1.5–2 cm), which are within the ^{210}Pb dating range,
367 yield ^{14}C ages of 2772 ± 19 and 4533 ± 22 yrs BP, respectively (Table 2). The large and variable
368 reservoir effect is corroborated by radiocarbon dates determined in another core from the western
369 basin of Torneträsk, which also range between 3580 and 5330 yrs BP in the investigated
370 uppermost 18 cm of the core (Vonk et al., 2012). For the luminescence dating, gamma
371 spectrometric measurements revealed a strong disequilibria in the U-Ra series for all samples,
372 and suggest a later enrichment in U after deposition. In light of this, minimum and maximum
373 ages were calculated using the dose rates determined using the activities of ^{238}U and ^{226}Ra ,
374 respectively (Table S1), and both sets of ages dramatically overestimate the expected ages for the
375 core. This suggests that the luminescence signal was not fully zeroed prior to burial. The presence
376 of a significant residual signal, together with the problem of disequilibrium over time in the
377 radionuclides, makes it impossible to obtain reliable OSL ages.

378 Because of the large age overestimates obtained from both radiocarbon and luminescence dating,
379 these results could not be used to constrain a robust age-depth relationship in core Co1282.

380 Instead, to constrain the chronology we cross-correlated significant geochemical changes in core
381 Co1282 with those clearly observed in the core from Abiskojaure for which a chronology was
382 established based on AMS radiocarbon dated woody plant macrofossils (twig fragments) (Table
383 2, Fig. S2). Four sections with similar patterns of change across both sediment sequences can be

384 identified: 1) high CI values (>12) and low Si/Al ratios in deeper sediments: 82-117 cm in
385 Torneträsk and >184 cm in Abiskojaure; 2) declining CI from >12 to ~ 11 , increasing bSi trend:
386 82-56 cm/184-104 cm (Abiskojaure's sediment composition is more variable during this period,
387 reflecting the coring location in close proximity to the inlet); 3) low CI (<11), high Si/Al: 56-12
388 cm/104-29 cm; and 4) in recent sediment a shift to higher CI (~ 11 or higher) and drop in Si/Al
389 12-0 cm/29-0 cm (Fig. 4). Only three macrofossils with sufficient C could be extracted from the
390 sediment core of Abiskojaure; however these macrofossils originate from sample levels at or near
391 the pronounced changes in geochemical properties recorded in Abiskojaure's sediments and thus
392 provide satisfactory age-depth control for the timing of major changes in the sediment
393 composition of Torneträsk. Similar geochemical changes in both cores suggest that the recorded
394 dynamics in core Co1282 are representative for the western Torneträsk catchment and/or
395 dominated by the riverine sediment input from Abiskojäkka, which drains Abiskojaure. Because
396 of the potential time lag that may exist for when the changes were recorded in Torneträsk and the
397 upstream-located Abiskojaure, inferred ages for core Co1282 based on the cross-correlation to
398 Abiskojaure should be considered as minimum ages.

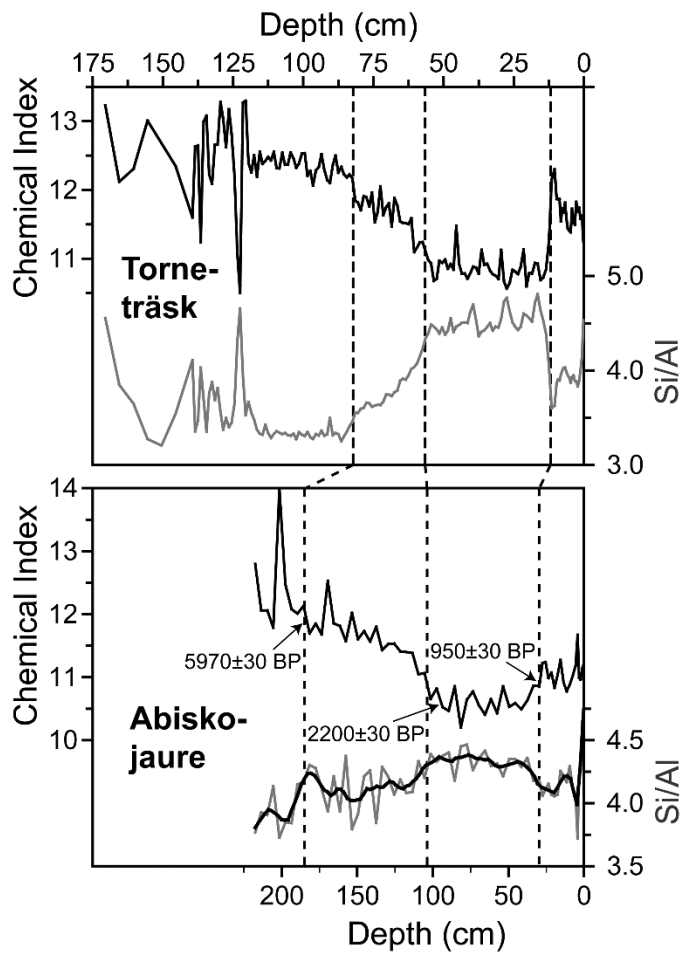
399 We constrained the basal age for the lacustrine sediments (117 cm depth) in core Co1282 based
400 on radiocarbon-dated sediment records from small lakes in this area and cosmogenic exposure
401 ages of glacially-scoured surfaces and erratics, which indicate a deglaciation age around
402 9500 ± 250 cal yr BP (e.g., Barnekow et al., 1998; Bigler et al., 2003; Shemesh et al., 2001;
403 Stroeven et al. 2002) (Fig. 4). This would imply an average Holocene sedimentation rate of 0.01
404 cm yr^{-1} , which is consistent with the recent sedimentation rate indicated by ^{210}Pb dating.

405

406 Table 2. Calibrated radiocarbon ages for the sediment sequences from Abiskojaure and
 407 Torneträsk.

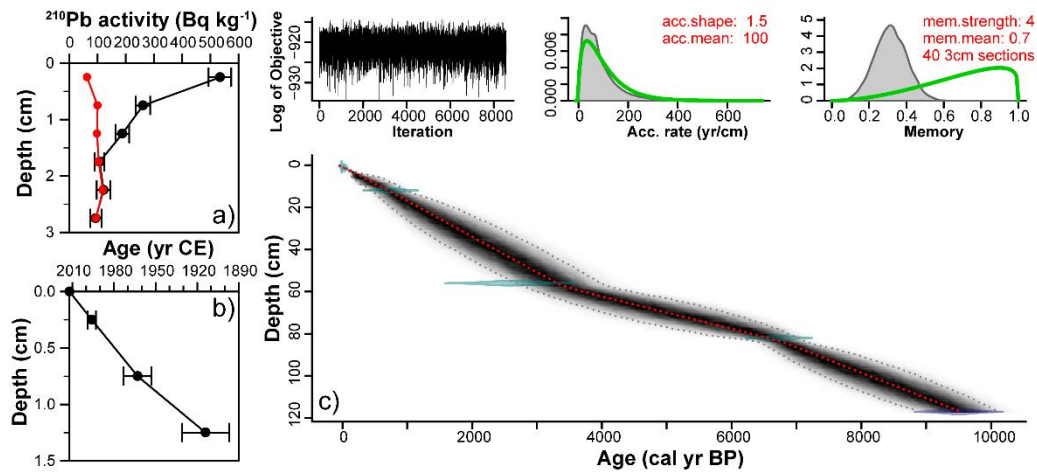
Laboratory ID	Sample ID	Composite depth (cm)	¹⁴ C age (yr BP)	Calibrated age range (2σ) (cal yr BP)	δ ¹³ C	Material
Beta-440907	AJ2016In_K32-33	32.5	950 ± 30	930–790	–26.8	Woody plant macrofossil
Beta-440908	AJ2016In_P85.5	95	2200 ± 30	2320–2130	–27.4	Woody plant macrofossil
Beta-442326	AJ2016In_P175	184.5	5970 ± 30	6885–6735	–28.0	Woody plant macrofossil
2234.1.1	Co1282-1_0-1	0.5	2772 ± 19	2926–2795	–25.4	Org C
2633.1.1	Co1282-1_1.5-2	1.75	4533 ± 22	5310–5055	–32.0	Org C
2235.1.1	Co1282-1_16-17	16.5	4867 ± 23	5646–5588	–24.5	Org C
2236.1.1	Co1282-3I_29-30	36.5	6290 ± 23	7263–7171	–24.1	Org C
2237.1.1	Co1282-3I_53-54	60.5	8752 ± 28	9891–9613	–22.6	Org C
2238.1.1	Co1282-3III_93-94	100.5	15914 ± 57	19413–18988	–18.2	Org C

408



409

410 Fig. 4. Correlation between core Co1282 from the western Torneträsk basin and the core from
 411 Abiskojaure using the chemical index and Si/Al ratios (note differences in x- and y-axis scaling
 412 and range). Radiocarbon dates are given as ^{14}C ages (yr BP) and dashed lines indicate sections of
 413 similar change used for correlating the two cores.



414

415 Figure 5. Chronology of core Co1282 from Torneträsk. a) Total (black) and supported (red) ^{210}Pb
 416 concentrations versus depth. b) Radiometric chronology showing the CRS model ^{210}Pb dates. c)
 417 Bayesian age-depth model based on ^{210}Pb -, dates inferred from the cross-correlation to the
 418 radiocarbon-dated sequence from Abiskojaure (transparent green) and deglaciation age of the
 419 area (transparent blue). 'Best' model based on the weighted mean age for each depth (red dashed
 420 line) and 95% confidence intervals (grey dashed lines). Greyscale intensity indicates likelihood of
 421 calendar ages. Upper panel (left to right) shows the Markov Chain Monte Carlo iterations (log of
 422 objective), the prior (green curves) and posterior (grey histograms) distributions for the
 423 accumulation rate and memory, respectively.

424

425 4.2.3 Holocene climatic and environmental development

426 Ice-dammed lake phase (168–117 cm blf, lithofacies C)

427 The proglacial sediments that accumulated in Torneträsk during its ice-dammed lake phase are
 428 characterized by a highly variable geochemistry compared to the subsequent lacustrine
 429 sediments. For example, major element concentrations fluctuate frequently between minimum
 430 and maximum values of 3.3–6.7% for Fe, 1.9–3.4% for K, and 5.5–8.4% for Al (Fig. 3). The high

431 variability, corresponding to the rhythmical bedding/lamination of the proglacial sediments, is
432 likely related to variations in grain-size, indicated by the K/Rb ratio ranging from 198 to 270, as a
433 result of the varying intensity and sediment load of the meltwater discharge from the retreating
434 ice sheet (Carrivick and Tweed, 2013).

435

436 Initial landscape development after deglaciation (117–84 cm blf; lithofacies D; ~9500–6600 cal
437 yr BP)

438 The initial hemi-pelagic lacustrine sediments deposited after the deglaciation of the area show
439 only minor compositional variations. Major element concentrations are stable with values around
440 2.9% for K and 8.0% for Al. Elemental ratios of K/Al, K/Rb, and Zr/Ti as well as the CI remain
441 constant throughout this period with values around 0.35, 221, 0.042 and 12.4, respectively (Figs.
442 3 and 6). bSi concentrations are continuously low (<0.1%), which is consistent with a very low
443 diatom abundance that in turn shows a stable diatom community composition (Fig. 7). Proxies
444 associated with OM exhibit slight increases from 0.4 to 0.7% for TOC, from 0.03 to 0.05% for
445 TN, from ~16 up to 18 for C/N ratios, and from +2.7 to +3.8‰ for $\delta^{15}\text{N}$, whereas $\delta^{13}\text{C}$ decreases
446 from -22.1 to -23.2‰ (Fig. 6). In the corresponding sediment section from Abiskojaure, the CI is
447 also high and relatively stable (~12) except for three samples associated with two clay layers
448 (Fig. 4).

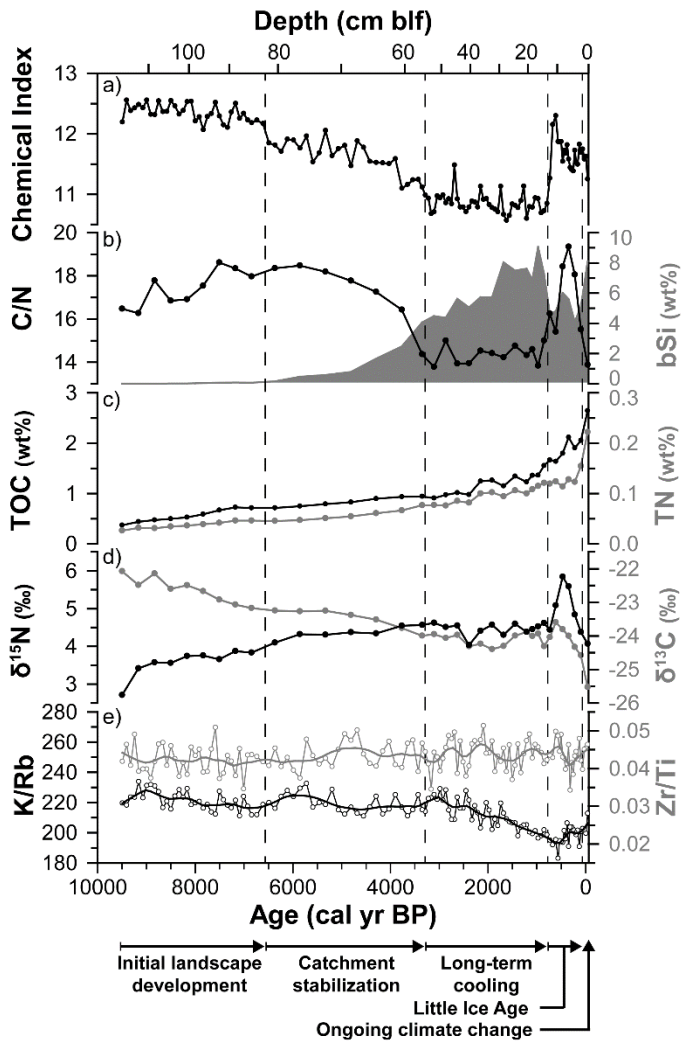
449 The constant major element concentrations, elemental ratios, and CI indicate that there are no
450 qualitative changes in the composition of the minerogenic matter input from the catchment during
451 the early Holocene. For example, the high CI is similar to the mean value of the proglacial
452 sediments (12.5), which shows that the minerogenic material deposited during this period is
453 essentially chemically unaltered. These data indicate a homogeneous sediment source, relatively

454 low chemical weathering rates and the presence of easily erodible and chemically unaltered
455 glacial deposits in the Torneträsk area. Low weathering rates would limit the release of DSi
456 and other nutrients into the aquatic system, and explain the low bSi concentrations and low
457 diatom abundance in this part of the record.

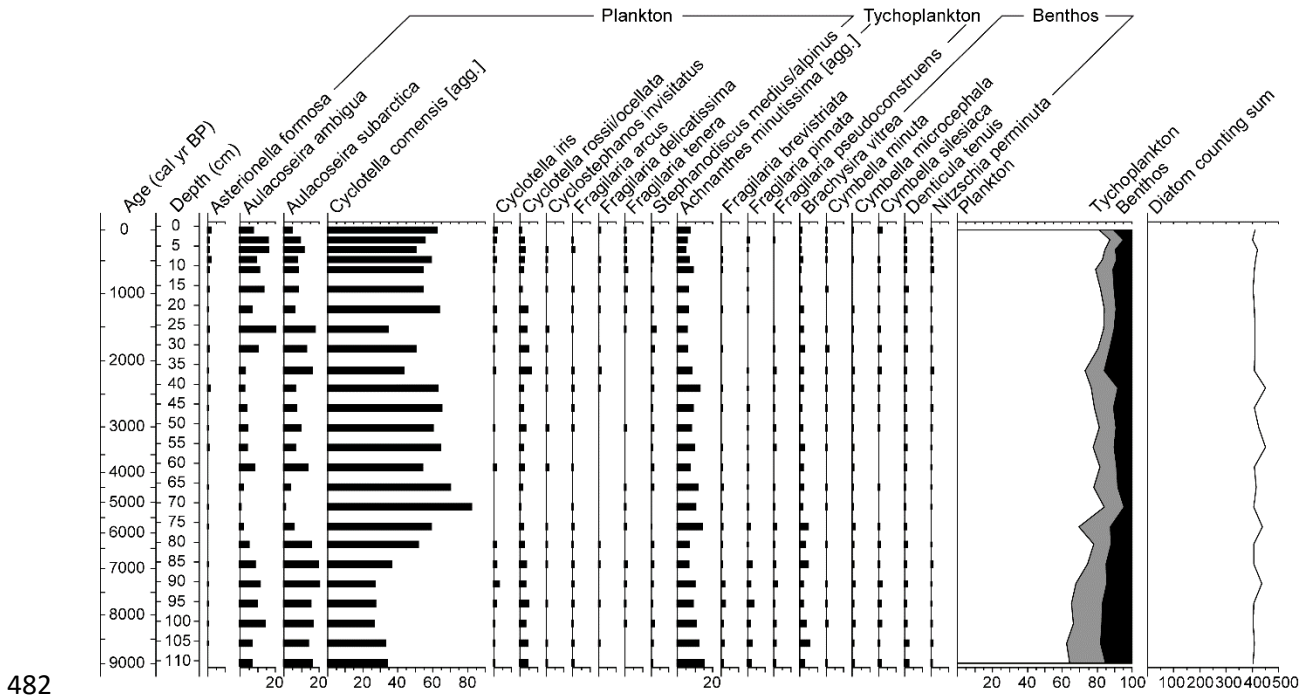
458 The slight increase in C/N indicates a gradually increasing export of terrestrial OM into the lake,
459 which would result from the gradual development of vegetation cover and soil formation in the
460 catchment. This initial landscape development would similarly explain the observed trends for
461 TOC, TN, $\delta^{15}\text{N}$ and $\delta^{13}\text{C}$. Conclusive interpretations of the $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ trends are complicated
462 by the complex nature of bulk OM isotope signals (Meyers and Ishiwatari, 1993; Meyers, 2003)
463 and the low magnitude of change of $\sim 2\text{‰}$ throughout the early and mid-Holocene; however
464 Hammarlund (1993) and Hammarlund et al. (1997) suggested that early Holocene declines in
465 $\delta^{13}\text{C}$ were the result of the transition from a deglaciated, vegetation-free landscape to tundra and
466 finally forest vegetation during the initial landscape development. The authors proposed that the
467 development of vegetation and soils enhanced soil respiration, which led to an increased export
468 of ^{13}C -depleted carbon dioxide to the aquatic system and subsequently to a ^{13}C -depletion in
469 phytoplankton.

470 Altered soil biogeochemical processes in response to a gradual development of vegetation cover
471 would likewise account for the increase of $\delta^{15}\text{N}$ in core Co1282. In general, bulk OM $\delta^{15}\text{N}$
472 increases in soils with depth and OM age, which has been shown for soil profiles in the
473 Torneträsk area (Makarov et al., 2008). During decomposition, the heavier ^{15}N is gradually
474 enriched in the residual OM by isotopic fractionation (Högberg, 1997). Correspondingly, the
475 increasing $\delta^{15}\text{N}$ trend in core Co1282 would result from enhanced soil OM decomposition and

476 increased leaching of ¹⁵N-enriched nitrate in the course of progressing soil and vegetation
 477 development in the Torneträsk catchment.



478
 479 Figure 6. a) Chemical Index (CI), b) biogenic silica (bSi) and C/N ratios, c) total organic carbon
 480 (TOC) and nitrogen (TN), d) $\delta^{13}\text{C}$, and $\delta^{15}\text{N}$, and e) K/Rb and Zr/Ti ratios versus inferred
 481 sediment ages for core Co1282 from Torneträsk over the Holocene.



482

483 Figure 7. Relative abundance of selected diatom taxa for core Co1282 from Torneträsk over the
 484 Holocene. The scale at the base of each panel is given in percentages.

485

486 Mid-Holocene thermal maximum and catchment stabilization (84–57 cm blf; lithofacies D;
 487 ~6600–3400 cal yr BP)

488 Compared to the relatively uniform composition of the early Holocene sediments, the sediment
 489 composition gradually changed between ~6600 and ~3400 cal yr BP. Approximately 3000 years
 490 after deglaciation, concentrations of K and Al started to progressively decline from ~2.9 to ~2.2%
 491 and from ~8.0 to 6.8%, respectively. Simultaneously, K/Al ratios and CI decreased from ~0.35 to
 492 ~0.33 and from ~12.2 to 11.2, while K/Rb and Zr/Ti ratios remained stable with values
 493 fluctuating around 219 and 0.043, respectively (Figs. 3 and 6). bSi levels slowly increased from
 494 <0.1 to ~2%, whereas C/N ratios decreased from ~18 to 14.4. Other proxies associated with OM
 495 follow their early Holocene trends and continuously increased from 0.7 to 1.0% for TOC, from

496 0.05 to 0.08% for TN, and from +3.8 to +4.5‰ for $\delta^{15}\text{N}$, while $\delta^{13}\text{C}$ continued to decrease from –
497 23.2 to –24.0‰ (Fig. 6). The CI in the sediment record from Abiskojaure likewise declined
498 during this period from ~12 to ~11.4 (Fig. 4).

499 The distinct declines in K/Al ratios and CI indicate a qualitative change in the sediment
500 composition that is likely driven by the depletion of mobile alkali and alkaline earth elements
501 through chemical weathering of mineral matter in catchment soils because elemental ratios
502 indicative for changes in grain-size (K/Rb) and source area (Zr/Ti) remained unchanged during
503 this period. Studies from Lake Kilpisjärvi, northern Finland (Kauppila and Salonen, 1997), and
504 from NW America (Whitehead et al., 1989) demonstrated that increased chemical weathering
505 rates (and a loss of more mobile elements) followed the establishment of denser forests in
506 previously glaciated areas. We therefore assume that the qualitative geochemical changes in core
507 Co1282 are likewise driven by the widespread establishment of denser pine-birch forests, which
508 commenced at ~7600 cal yr BP and reached maximum density during the HTM at ~6300 cal yr
509 BP in the western Torneträsk catchment (Barnekow, 2000; Barnekow and Sandgren, 2001). The
510 climate optimum during the HTM (~8000–4800 cal yr BP), with temperatures 1.5–2.0 °C warmer
511 in northern Fennoscandia than at present (Barnekow, 2000; Barnekow and Sandgren, 2001,
512 Seppä et al., 2009) (Fig. 8), also had an impact on in-lake processes and the diatom community
513 composition in Torneträsk. The diatom assemblage shows a gradual shift with an increased
514 abundance of *Cyclotella comensis* and decrease in *Aulacoseira ambigua* and *subarctica* (Fig. 7)
515 likely in response to elevated temperatures during this period. Warming and earlier ice break-up
516 may enhance the strength and duration of thermal stratification, which favors more buoyant
517 *Cyclotella* taxa over heavier *Aulacoseira* taxa (Rühland et al., 2015). Bigler et al. (2006) found a
518 similar shift in the diatom community in the small lake Vuolep Njakajaure (Fig. 1 F); however,

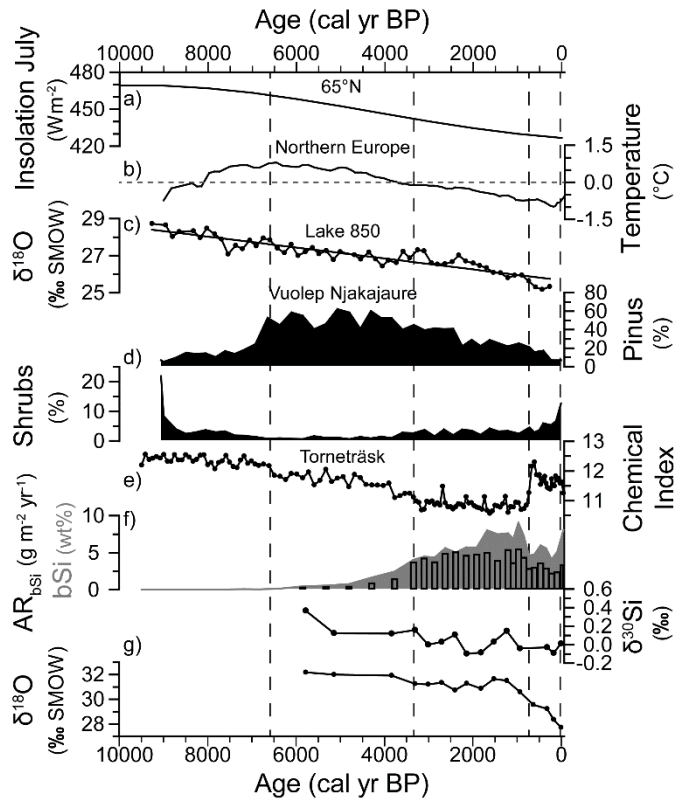
519 this shift occurred ~1000 years earlier, which indicates the slower (delayed) response to climate
520 change in the large and deep Torneträsk as also shown for other (sub)arctic lakes (Rühland et al.,
521 2015 and references therein).

522 The preceding catchment transformation from bare bedrock and glacial sediment deposits
523 after deglaciation to a forested catchment supporting soil formation during the mid-Holocene is
524 corroborated by the consistent trends of $\delta^{15}\text{N}$ and $\delta^{13}\text{C}$, assuming that biogeochemical processes
525 in soils primarily control trends in stable C and N isotopes in the Torneträsk sediment during the
526 early lake development (cf., Hammarlund, 1993; Hammarlund et al., 1997; Wolfe et al., 1999).

527 Rather than indicating a decreased input of terrestrial OM into the lake, the decline of C/N ratios
528 together with the concurrent increase of bSi levels shows an increased importance of aquatic OM
529 production. This onset of notable aquatic production was facilitated by the increased export of
530 nutrients from the catchment following the vegetation development in the catchment and
531 subsequent intensification of chemical weathering. Likewise in Lake Kilpisjärvi, northern
532 Finland, initial bSi concentrations were low and did not significantly increase until the expansion
533 of pine and establishment of denser vegetation (Kauppila and Salonen, 1997). This is consistent
534 with watershed studies from northern Sweden, which show that fluxes of DSi in rivers increase
535 with increasing vegetation cover in the catchment (Humborg et al., 2004).

536 Changes in the availability of DSi can be traced using Si isotope analyses of diatom silica. In core
537 Co1282, we could extract sufficient material for diatom isotope analyses from 76 cm blf (~6000
538 cal yr BP) upwards when bSi concentrations were $\geq 0.2\%$. $\delta^{30}\text{Si}_{\text{diatom}}$ decrease from +0.37 to 0‰
539 between c. 6000 and 3000 cal yr BP, followed by lower values fluctuating around 0‰ (min: –
540 0.10, max: +0.15‰) during the remainder of the late Holocene until present (Fig. 8). Diatoms
541 preferentially incorporate ^{28}Si compared to ^{30}Si (and ^{29}Si), and consequently with enhanced

542 productivity and utilization of DSi, $\delta^{30}\text{Si}$ ($^{30}\text{Si}/^{28}\text{Si}$) progressively increases in the DSi pool. This
543 productivity signal is recorded in the $\delta^{30}\text{Si}$ of diatoms because diatoms then incorporate DSi from
544 this progressively ^{30}Si -enriched pool. Variations in the DSi supply can also alter the $\delta^{30}\text{Si}$ signal
545 of diatoms by affecting the biological DSi demand (Leng et al., 2009; Swann et al., 2010). In the
546 sediments of Torneträsk, the initial $\delta^{30}\text{Si}_{\text{diatom}}$ decline indicates increased DSi utilization during
547 the earlier lake development. Because diatom productivity was very low during this period, we
548 interpret the increased utilization as an indication for a higher DSi demand that resulted from the
549 initially low weathering rates and thus low DSi supply. This suggests that diatom productivity in
550 Torneträsk was Si limited during the early lake ontogeny, which delayed the onset of notable
551 diatom production by ~3000 years after deglaciation until vegetation development facilitated Si
552 export. Alternatively, an increase in weathering rates could also contribute to the initial $\delta^{30}\text{Si}$
553 decline; for example, Opfergelt and Delmelle (2012) found that $\delta^{30}\text{Si}$ decreased in soil profiles
554 with increasing weathering degree.



555

556 Figure 8. a) July Insolation at 65°N (Berger and Loutre, 1991), b) LOESS smoothed (span: 0.05)
 557 stacked summary curve of the pollen-based temperature variability for Northern Europe (Seppä et
 558 al., 2009), c) $\delta^{18}\text{O}_{\text{diatom}}$ for Lake 850 (Fig. 1 I, Shemesh et al., 2001), percentage terrestrial pollen
 559 from d) *Pinus sylvestris* and shrub vegetation in the sediment record of Vuolep Njakajaure (Fig. 1
 560 F; Barnekow, 2000) compared to e) the Chemical Index (CI), f) biogenic silica (bSi)
 561 concentrations and accumulation rates (AR_{bSi}), g) diatom $\delta^{18}\text{O}$ and $\delta^{30}\text{Si}$ in the sediment core
 562 Co1282 from Torneträsk over the past 6000 and 10000 years, respectively.

563

564 Late Holocene climate cooling (57–13 cm blf; lithofacies D; ~3400–750 cal yr BP)

565 Following the long-term vegetation expansion and catchment stabilization during the mid-
 566 Holocene, most organic and inorganic geochemical properties in core Co1282 show no major

567 variations between ~3400 and 750 cal yr BP. Major element concentrations, K/Al and Zr/Ti
568 ratios as well as the CI fluctuate around 2.2% for K, 6.6% for Al, 0.33 for K/Al, 0.044 for Zr/Ti,
569 and 10.9 for CI (Fig. 3). C/N ratios, $\delta^{13}\text{C}$, and $\delta^{15}\text{N}$ are relatively stable during this period with
570 values around 14.4, -24.1‰ , and $+4.5\text{‰}$, respectively, while TOC and TN concentrations
571 continued to increase slightly from 1.0 to 1.6% and from 0.08 to 0.12%, respectively (Fig. 6). In
572 contrast, bSi increased from 2 to above 7% over the whole period, whereas K/Rb ratios started to
573 decline around 2200 cal yr BP from values around 220 during the early and late Holocene to
574 ~200 by ~750 cal yr BP. Fe and Mn co-vary during this part of the sediment record and layers
575 enriched in Fe and Mn are most likely buried relicts of the active redox horizon that continually
576 moves upward with progressing sediment accumulation (Och et al., 2012) (Fig. 3). The CI in the
577 corresponding sediment section from Abiskojaure reached its lowest values during the Holocene
578 and stabilized around 10.7 (Fig. 4).

579 The relatively constant sediment geochemistry between ~3400 and 750 cal yr BP indicates no
580 larger changes in the composition of the catchment-derived sediment delivered to the coring site
581 in the western lake basin. In comparison, studies of smaller lakes at higher altitudes in the
582 catchment suggest increased soil erosion during the late Holocene from 3200–2600 cal yr BP
583 onwards, respectively (Berglund et al., 1996; Snowball and Sandgren, 1996; Rubensdotter and
584 Rosqvist, 2003). For the late Holocene, several paleolimnological studies in the Torneträsk area
585 indicate a climate cooling that led to a tree-line retreat, which would expose previously forested
586 soils and thus facilitate soil erosion. Pollen and macrofossil records indicate a lowering of the
587 pine tree line by ~175 m and also of mountain birch already from ~4500 cal yr BP, which is
588 assumed to represent a decline of the growing season temperature of $\sim 1.5\text{--}2^\circ\text{C}$ compared to
589 today. This tree-line retreat ultimately led to the establishment of today's more sparse subalpine

590 birch forests and an increased presence of dwarf shrubs, grasses and sedges across the landscape
591 also at lower altitudes (Barnekow, 1999, 2000) (Fig. 8). Temperature reconstructions based on
592 diatoms, chironomids and pollen likewise indicate a pronounced cooling during the late Holocene
593 by $\sim 1\text{-}2^\circ\text{C}$, but depending on the site and the chosen biological indicator the timing of the cooling
594 onset varies between ~ 3500 and 1900 cal yr BP (Bigler et al., 2002; 2006). These landscape
595 dynamics agree well with the proposed neoglacial expansions of the Kårsa glacier (Fig. 1 B)
596 around 3300 cal yr BP (Berglund et al., 1996; Snowball and Sandgren, 1996) and of the
597 Kalanvare glacier, located just outside the Torneträsk catchment (Fig. 1 A), at c. 4400 , 3000 ,
598 2000 , and after 1200 cal yr BP (Rosqvist et al., 2004). Increased glacial activity would increase
599 production of finer-grained material (rock flour), which would ultimately be accumulated in
600 Torneträsk. Declining K/Rb ratios in core Co1282 indicate such an increased contribution of
601 finer-grained material from ~ 2200 cal yr BP onwards. This trend is corroborated by grain-size
602 analyses from a short core that was recovered close to the coring site of core Co1282, which
603 shows an increasing clay content over the corresponding depth interval (Vonk et al. 2012).
604 The continuous increase of bSi concentrations during this period would suggest a progressive
605 improvement of environmental conditions for aquatic production but accumulation rates of bSi
606 reveal instead that diatom production reached a somewhat stable level with AR_{bSi} around 4.5 ± 1
607 $\text{g} \cdot \text{m}^{-2} \cdot \text{yr}^{-1}$ after the initial increase (Fig. 8). The overall low AR_{bSi} in Torneträsk is of the same
608 order of magnitude as those reconstructed for other large lakes such as Lake Baikal (Qiu et al.,
609 1993) or Lake El'gygytgyn (Meyer-Jacob et al., 2014b) in Siberia. Regardless of changes in bSi
610 content and flux there was no significant change in the diatom assemblage (Fig. 7).

611

612 Soil erosion during the Little Ice Age (13–1 cm blf; lithofacies D; $\sim 750\text{-}50$ cal yr BP)

613 The relatively stable sediment geochemistry during the late Holocene was interrupted by a
614 pronounced shift in the sediment composition commencing at $\sim 750 \pm 275$ cal yr BP
615 contemporaneous with the beginning of the LIA (Matskovsky and Helama, 2014; Melvin et al.,
616 2012; Seppä et al., 2009). Major element concentrations, K/Al ratios, and the CI shifted to higher
617 values of $\sim 2.6\%$ for K, $\sim 7.2\%$ for Al, ~ 0.36 for K/Al, and ~ 11.7 for the CI, while Zr/Ti and K/Rb
618 show no major changes and remain around 0.043 and 200, respectively (Fig. 3). bSi
619 concentrations dropped to $\sim 5\%$, whereas C/N ratios and $\delta^{15}\text{N}$ increased to maximum values of
620 19.4 and $+5.8\%$. $\delta^{13}\text{C}$ slightly increases only initially to -23.6% and drops to Holocene
621 minimum values of -25.5% at the sediment surface, while TOC and TN follow their Holocene
622 trend and reach maximum values of 2.6 and 0.22% in the most recent sediment (Fig. 6). In the
623 corresponding sediment from Abiskojaure, the CI also shifted to significantly higher values
624 (~ 11.1) (Fig. 4).

625 Elevated K/Al ratios and CI, together with constant Zr/Ti and K/Rb ratios, indicate a stable grain-
626 size distribution and sediment source but that the minerogenic material deposited in Torneträsk
627 from ~ 750 cal yr BP onwards is less-intensively weathered compared to sediment that
628 accumulated since the beginning of the late Holocene cooling. Less-weathered material would
629 come from deeper mineral soil layers, and together with an increase in C/N ratios indicating more
630 terrestrial OM, this signifies enhanced soil erosion in the Torneträsk area. An enhanced erosion
631 of catchment soils is also supported by the simultaneous increase of $\delta^{15}\text{N}$ in the sediment because
632 OM from deeper soils is enriched in ^{15}N (Högberg, 1997); for example, Makarov et al. (2008)
633 showed that deeper soil horizons (>10 cm soil depth) in the Abisko area exhibit $\delta^{15}\text{N}$ of $>+5\%$.
634 During the LIA, temperatures were ~ 1 °C cooler than at present in Fennoscandia with coldest
635 tree-ring inferred summer temperatures during the 13th through 19th centuries for the Torneträsk

636 area (Matskovsky and Helama, 2014; Melvin et al., 2012; Seppä et al., 2009) (Figs. 8 and 9).
637 Paleoecological temperature reconstructions from smaller lakes in the Torneträsk watershed
638 likewise suggest a ~ 1 °C cooling over the last millennia (e.g., Bigler et al., 2002, 2003;
639 Hammarlund et al., 2002). The pronounced cooling during this period is emphasized by the
640 formation of permafrost in the area after 700 cal yr BP (Kokfelt et al., 2010; Rydberg et al.,
641 2010). Neo-glacial activity also increased over the last millennium until the end of the LIA
642 (Rosqvist et al., 2004; Snowball and Sandgren, 1996), which is reflected in the Torneträsk record
643 by a higher input of finer-grained material indicated by the lowest K/Rb ratios recorded for the
644 Holocene during this period (Fig. 6). However, increased glacial activity alone cannot explain the
645 change in sediment composition during the LIA because, while increased glacial erosion would
646 have exposed less-weathered material, it would not have altered the quality of the catchment-
647 derived OM.

648 Carbonate and diatom silica $\delta^{18}\text{O}$ from several small-lake records in the study region indicate
649 changes in atmospheric circulation patterns over the Holocene, and particularly during the late
650 Holocene (e.g., Rosqvist et al., 2004, 2007; Shemesh et al., 2001). In the sediment sequence from
651 Torneträsk, $\delta^{18}\text{O}$ in diatom silica declined by 4.5‰ during the past ~ 6000 years. The $\delta^{18}\text{O}$ decline
652 was relatively slow between ~ 6000 and 750 cal yr BP with a total decrease of ~ 1.5 ‰, but
653 accelerated substantially during the past ~ 750 years when $\delta^{18}\text{O}$ declined by an additional ~ 3 ‰
654 (Fig. 8). The successive long-term decline over the Holocene has been ascribed to changes in the
655 relative contribution of different air masses to the area, changing from a maritime climate with a
656 strong zonal airflow from W/SW to a more continental climate with a dominant meridional
657 airflow from N/NE (Hammarlund et al., 2002; Shemesh et al., 2001).

658 Compared to the slight decline (1.5‰) over most of the Holocene, the strongly amplified decline
659 (3‰) in diatom $\delta^{18}\text{O}$ during the past ~750 years requires a more significant change in climate.
660 The main factors controlling diatom $\delta^{18}\text{O}$ are lake-water temperature and the $\delta^{18}\text{O}$ composition of
661 the ambient lake-water, which in turn depends on the $\delta^{18}\text{O}$ composition of the local precipitation
662 and evaporation. Several studies from smaller lakes in the Torneträsk area have suggested that
663 temperature could not be the sole driver of the observed $\delta^{18}\text{O}$ changes and that a 1°C temperature
664 decline, as indicated for the LIA, alone would have caused a $\delta^{18}\text{O}$ decline of <1‰ (e.g.,
665 Hammarlund et al., 2002; Rosqvist et al., 2013; Shemesh et al., 2001). Reduced evaporation
666 could amplify the $\delta^{18}\text{O}$ decline, but lake-water surveys show that lakes in the area are generally
667 little affected by evaporation (Jonsson et al., 2009); furthermore, lake-water $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values
668 from Torneträsk itself (Shemesh et al., 2001) fall on the local meteoric water line established by
669 Jonsson et al. (2009), which indicates that the lake is – at least at present – unaffected by
670 evaporation.

671 Changes in the $\delta^{18}\text{O}$ of precipitation is likely the most critical factor in the Torneträsk watershed
672 and could have contributed to the pronounced $\delta^{18}\text{O}$ decline during the past ~750 years in two
673 ways; one is an even stronger dominance of ^{18}O -depleted air masses, while the other is a relative
674 increase in winter precipitation, which is more ^{18}O -depleted compared to summer precipitation
675 due to the temperature difference during condensation (Dansgaard 1964). Each of these has been
676 proposed by different paleoecological studies in the area. Rosqvist et al. (2004; 2007) suggested
677 that lake-water $\delta^{18}\text{O}$ declined during periods dominated by dry and cool arctic air masses from N,
678 leading to cooler summers. For the LIA, Loader et al. (2013) inferred an increased contribution of
679 arctic air leading to cool and sunny summers from tree-ring based temperature and sunshine
680 reconstructions for the Torneträsk area (Fig. 9).

681 Alternatively, Jonsson et al. (2010) suggested that increased winter precipitation and elevated
682 winter/summer precipitation ratios led to a $\delta^{18}\text{O}$ decline in a high-altitude alpine lake record in
683 the Torneträsk catchment. However, precipitation reconstructions for the study region exhibit
684 mixed results (Hammarlund et al., 2002; Seppä and Birks, 2001). Lower summer temperatures as
685 well as increased winter precipitation could both contribute to neo-glaciation, which has been
686 observed during the LIA in Fennoscandia as well as specifically in the Torneträsk area (Karlén
687 1988; Nesje 2009; Rosqvist et al., 2004; Snowball and Sandgren, 1996).

688 Amplified climate change during the LIA is corroborated in the Torneträsk record by the
689 simultaneous decline of AR_{bSi} (Fig. 8). Diatom production is sensitive to changes in the length of
690 ice cover (Lotter and Bigler, 2000), which depends on the temperature (Magnuson et al., 2000) as
691 well as the amount and timing of snowfall. Monitoring data since the early 20th century indicate
692 that the ice-cover duration of Torneträsk is primarily driven by temperature. From 1913 to 2006,
693 the mean ice-cover duration of Torneträsk decreased by ~40 days. Simultaneously, mean annual
694 air temperatures have increased by 2.5°C at the Abisko research station, whereas snow depth
695 initially increased but then declined since the 1980's (Callaghan et al., 2013). The increasing
696 length of the open water season over the past 100 years, together with increasing temperatures,
697 would explain the recent increase in bSi and AR_{bSi} (Figs. 6 and 8). The overall good agreement
698 between bSi variations in the sediment record and tree-ring based changes in temperature and
699 atmospheric circulation emphasizes the climate sensitivity of diatom production in Torneträsk
700 during the past millennium (Fig. 9), which, however, did not significantly alter the diatom
701 assemblage composition (Fig. 7).

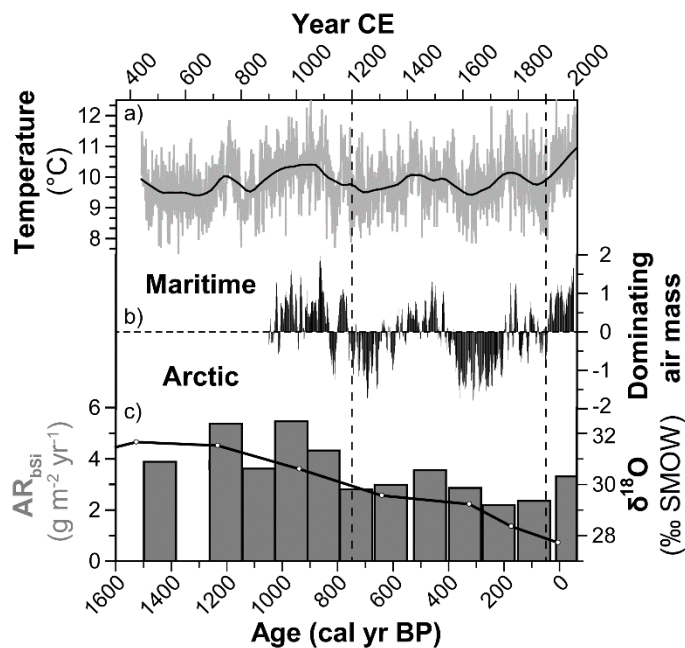
702 Remarkably, the pronounced shift in sediment composition in Torneträsk did not occur until the
703 start of the LIA despite an already ongoing climate cooling for more than 2000 years. In addition,

704 the abrupt nature of these changes in the sediment record indicates either a rather sudden change
705 of environmental conditions with the onset of the LIA or alternatively that environmental
706 conditions crossed a critical threshold in the Torneträsk watershed when temperatures further
707 declined during the LIA. Climate reconstruction for the Torneträsk region indicate that the scale
708 of climate change during the LIA was not unprecedented during the late Holocene. For example,
709 no qualitative changes in the deposited sediments occurred in response to the gradual 1-2°C
710 decline since the HTM or in response to a temperature decline of equal magnitude to the LIA
711 during a cold period ~1400 cal yr BP (Matskovsky and Helama, 2014; Seppä et al., 2009) (Figs. 8
712 and 9).

713 Sediment records from smaller lakes in the Torneträsk catchment indicate increased mobilization
714 of minerogenic matter already from 3400-2600 cal yr BP onwards, following tree-line retreat as
715 well as increasing glacial activity during the late Holocene (Berglund et al., 1996; Rubensdotter
716 and Rosqvist, 2003; Snowball and Sandgren, 1996). Similarly, increased soil erosion has been
717 reported for higher elevation lakes in other parts of the Scandinavian Mountains following the
718 late Holocene tree-line retreat (Barnett et al, 2001; Hammarlund et al. 2004). In contrast,
719 increased erosion of catchment soils is not recorded in sediments of Torneträsk's western basin
720 before the onset of the LIA when pollen records from the area show an increasingly open
721 landscape, indicated by the further increased presence of dwarf shrubs, grasses and sedges
722 (Barnekow, 1999, 2000; Bigler et al., 2002) (Fig. 8). The increasingly sparse vegetation cover
723 would have further increased the susceptibility of soils to erosion, which are generally thin (~1
724 m), poorly developed, and have relatively shallow organic horizons (<15 cm) above the modern
725 tree-line in the Torneträsk area (Darmody et al. 2004; Makarov et al. 2008). This is supported by
726 radiocarbon-dated buried soils from the catchment that suggest highest deposition rates of

727 upslope eroded material during the LIA (Darmody et al. 2004). We propose that, following the
 728 tree-line retreat, catchment destabilization and subsequent soil mobilization occurred first mainly
 729 locally at higher elevations. With diminishing vegetation cover, the Torneträsk catchment
 730 gradually became more sensitive to soil erosion through, e.g., rainfall events and snowmelt run-
 731 off, and crossed a critical threshold around ~750 cal yr BP. From that time onwards, previously
 732 forested soils in large parts of the catchment were increasingly exposed to erosion and less-
 733 weathered material was also transported into Torneträsk's western basin.

734



735

736 Figure 9. a) Summer (May to August) temperature reconstructions for the Torneträsk area based
 737 tree ring maximum late-wood-density (Melvin et al., 2012) and b) dominant atmospheric
 738 circulation based on departures between tree ring inferred temperature (X-ray density) and
 739 sunshine (stable carbon isotope ratio), where negative values indicate a dominance of cold-sunny
 740 weather (Arctic air dominant) and positive values indicate a dominance of warm-cloudy
 741 conditions (maritime air dominant) (Loader et al., 2013), compared to c) biogenic silica

742 accumulation rates (AR_{bsi}) and diatom- $\delta^{18}O$ of core Co1282 from Torneträsk for the period 1600
743 to -58 cal yr BP.

744

745 Ongoing environmental change following climate warming (1–0 cm blf; lithofacies D; 50 cal yr
746 BP to present)

747 During the past century, K and Al concentrations have returned to pre-LIA values, and K/Al
748 ratios and CI started to decline (Fig. 6). Together with the accompanying decline of C/N ratios
749 and $\delta^{15}N$, which both indicate a reduced soil-derived OM input to the lake, these data suggest a
750 stabilization of the catchment that has reduced soil erosion and terrestrial carbon export in the
751 Torneträsk area. This return to a sediment composition more similar to pre-750 cal yr BP can
752 only be linked to ongoing climate warming in the area, with a mean annual air temperature
753 increase of 2.5°C during the last century, that is also reflected in the recent increase in diatom
754 production (Figs. 8 and 9). Increased soil stabilization is corroborated by vegetation studies from
755 the Abisko area showing increases in vegetation growth rates and distributional range during the
756 past century (Callaghan et al., 2013, and references therein).

757 It is difficult to interpret the recent return to sedimentation in the western basin similar to that
758 prior to ~750 cal yr BP strictly from a climate perspective because of the increased human impact
759 in the area during the past centuries, including infrastructure development, increased reindeer
760 husbandry and more recently tourism. However, were these important, major environmental
761 disturbances such as the railroad (built AD 1898-1902) and highway (built AD 1980-1984)
762 construction along the southern shore of the lake would have caused opposite trends by
763 increasing soil disturbance and erosion, and have thus not led to discernable changes in the
764 sediment composition of Torneträsk's central western basin.

765

766 **5. Conclusions**

767 Although the dating of the sediment record from Torneträsk relies on a combination of directly
768 determined and inferred ages that we could transfer from a dated sediment record from a lake
769 located in one of Torneträsk's main subcatchments, the sediment record of this 330 km² lake
770 provides valuable insights into regional landscape-scale changes over the past c. 9500 yr. The
771 initial landscape development and accompanying lake ontogeny were the dominating influences
772 on sediment composition in Torneträsk after deglaciation. During the early to mid-Holocene,
773 gradual processes associated with the development of soils and vegetation drove the element
774 cycling in the aquatic and terrestrial environment and led to an initial nutrient (Si) limitation in
775 Torneträsk preventing notable diatom production until ~6600 cal yr BP. With progressing
776 ecosystem stabilization, the lake became more climate-sensitive during the late Holocene when
777 diatom production responded to climate variations (tree-ring inferred temperature) during the last
778 millennia, and a drop in diatom $\delta^{18}\text{O}$ suggests a larger shift in atmospheric circulation and/or
779 precipitation patterns.

780 In contrast to the successive change in the sediment composition following landscape
781 development, climate cooling during the late Holocene led to a pronounced shift in Torneträsk's
782 sediment composition that occurred rather abruptly (~100 years) but not before the onset of the
783 LIA around ~750 cal yr BP. This shift marking increased soil erosion was delayed by >2000
784 years compared to the first indications (tree-line retreat, increased erosion) for climate cooling
785 that are recorded in small lakes located in the Torneträsk catchment. The rather abrupt and
786 delayed compositional change in Torneträsk suggests a non-linear threshold response to climate
787 forcing. Following further climate cooling during the LIA, larger parts of the Torneträsk

788 watershed became susceptible to soil erosion, and from ~750 cal yr BP onwards environmental
789 changes initially only registered in higher located sub-catchments more sensitive to tree-line
790 dynamics were also registered in Torneträsk.

791 The second more rapid ecosystem change recorded in the sediment record of Torneträsk is related
792 to the ongoing climate change that thus far has led to a return to pre-LIA conditions, which
793 suggests catchment stabilization related to increased vegetation cover. In comparison to the
794 previously recorded signals – gradual response to landscape development and abrupt but lagged
795 response to climate cooling during the late Holocene, the lake ecosystem response to ongoing
796 climate change is rather immediate, emphasizing the unprecedented scale of climate change in
797 subarctic Fennoscandia during the past century.

798

799 **Acknowledgements**

800 We thank Florian Boxberg, Gerard Rocher Ros, Cristian Gudasz and Rolf Zale for their help in
801 recovering sediment from Torneträsk and Abiskojaure. We also thank Peter Rosén, Thomas
802 Westin and the Climate Impacts Research Centre (CIRC) in Abisko for their support as well as
803 the Swedish Research Council and the “Stiftelsen J C Kempes Stipendiefond” for financial
804 support. Fieldwork was supported by the EU INTERACT transnational access program. Hilary
805 Sloane at the British Geological Survey undertook the O and Si isotope analysis of the diatom
806 silica. We also thank the anonymous reviewers for their helpful comments and suggestions.

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