Regional Holocene climate and landscape changes recorded in the large subarctic lake Torneträsk, N Fennoscandia

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Abstract

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

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

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

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

In northernmost Sweden in the subarctic catchment of the large lake Torneträsk (“Torne Lake” in English; surface area: 330 km²; watershed: 3350 km²; Fig. 1), the observed climate change during the past century has thus far led to a mean annual temperature increase of 2.5°C, a shortened (~40 days) lake ice cover period (Callaghan et al., 2013), a reduced permafrost thickness (Åkerman and Johansson, 2008), and an increase of wet areas in mires due to permafrost degradation (Malmer et al., 2005). The Torneträsk area has been subject to numerous studies in different disciplines focusing on a modern process understanding, including carbon cycling (e.g., Christensen et al., 2007; Karlsson et al., 2010), permafrost dynamics (e.g., Åkerman and Johansson, 2008; Kokfelt et al., 2009), and sediment transport (e.g., Beylich et al., 2006). In addition to these studies on recent processes, sediment records from several small lakes in the catchment have been studied to reconstruct Holocene climate and environmental change with a focus on vegetation development (e.g., Barnekow, 1999; 2000), quantitative climate reconstructions (e.g., Bigler et al., 2002; 2003), sediment transport and erosion (e.g., Snowball and Sandgren, 1996; Rubensdotter and Rosqvist, 2003) and changes in atmospheric circulation patterns (e.g., Hammarlund et al., 2002; Rosqvist et al., 2007; Shemesh et al., 2001) (Fig. 1).
Because of the complexity and multitude of studies of small lakes conducted in the Torneträsk catchment, a study of the lake itself represents an ideal site for investigating the integrated regional climatic and environmental development in northernmost Sweden since the retreat of the Fennoscandian ice sheet. Studies of Torneträsk’s sediment record are thus far scarce and have only focused on a general sedimentological description of the lacustrine sediment (Andrén et al., 2002), recent changes in the input of terrestrial organic matter (OM) (Vonk et al., 2012) and on depositional patterns and subaquatic topography (Vogel et al., 2013).

Using a combination of geochemical, isotopic and biological proxies, the aims of this multi-proxy study are twofold. One aim was to survey the surface sediment geochemistry in order to assess spatial variability, how this variability relates to catchment influences, and how signals might vary across this 70-km-long lake. The second and main aim was to analyze a complete sediment sequence from Torneträsk in order to determine the regional climate and environmental development in northernmost Sweden during the Holocene and to assess i) how responses to environmental change are recorded in the large lake system compared to small lakes that have been studied within its catchment, and ii) to what extent the ongoing climate change has already affected the lake ecosystem, and how the scale of any identified changes compares to changes that occurred during past periods of distinct climate change such as the Holocene Thermal Maximum (HTM) or the Little Ice Age (LIA).
Figure 1. (left) Map over Fennoscandia showing the location of the study region. (right) Elevation map over the watershed of Torneträsk (black contour line). Dashed contour line defines the watershed of Abiskojaure, filled squares indicate the location of three weather stations (Swedish Meteorological and Hydrological Institute) across the Torneträsk catchment, while open stars mark the sampling sites in Torneträsk (core Co1282) and Abiskojaure. Open circles indicate locations of paleolimnological studies on smaller lakes from the area (A: Vuolep Allakasjaure, B: Kårsa valley lakes, C: Pikkujärvi, D: Lake Njulla, E: Lake Tibetanus, F: Vuolep Njakajaure and Badsjön, G: Villasjön and Inre Harrsjön, H: Vuoskkujávri, I: Lake 850/865, J: Lake Latteluokta. The white arrow highlights the outlet of Torneträsk, Torneälven, in the SE.

2. Regional setting

Torneträsk (68° 29'-68° 11' N, 20° 01'-18° 36' E; 341 m a.s.l.) is 70 km long (NW-SE), maximally 10 km wide (SW-NE), has a surface area of 330 km² and a watershed of 3350 km² (Fig. 1). The lake has an average depth of 53 m and a maximum depth of 168 m (based on bathymetric measurements in 1920/1921, Abisko Research Station; Fig. 2), with an estimated
water residence time of c. 8.5 years (Swedish Meteorological and Hydrological Institute (SMHI); https://vattenwebb.smhi.se/). The watershed is drained by several smaller streams and rivers of which the Abiskojäkka, entering the lake west of the village of Abisko, is the largest inlet with an average discharge of 14 m$^3\cdot$s$^{-1}$ (Fig. 1). Torneälven (“Torne River” in English) drains Torneträsk to the SE with an average discharge of 65 m$^3\cdot$s$^{-1}$ (1999–2013; SMHI; https://vattenwebb.smhi.se/) (Fig. 1). The lake is (ultra)oligotrophic (TN = 195±25 µg·l$^{-1}$, TP = 3±2 µg·l$^{-1}$, TOC = 1.3±0.5 mg·l$^{-1}$), circumneutral (pH = 7.3±0.1) (environmental monitoring data (MVM); http://www.slu.se/miljodata-MVM), and δ$^{18}$O and δ$^2$H of water samples from Torneträsk were –12.7/–12.8‰ and –92.6/–93.0‰, respectively in August 1999 (Shemesh et al., 2001).

The complex bedrock geology in the Torneträsk area consists of Archean plutonites (mainly granite and syenite) and their metamorphic products that are overthrusted by Caledonian metasediments and amphibolites (Swedish Geological Survey (SGU) database; maps2.sgu.se/kartgenerator/maporder_sv.html). During the retreat of the Weichselian ice sheet, the NW-SE trending Torneträsk depression hosted one/multiple ice-dammed lakes with water levels up to 250 m higher than today as suggested by paleoshoreline and paleodelta deposits (Gretener and Stromquist, 1981; Melander 1977). The area was largely deglaciated around 9500 cal yr BP (glaciers occupy ~0.5% of the watershed today (SMHI; https://vattenwebb.smhi.se/)), leading to the drainage of the precursor lakes and the likely establishment of the present shoreline. This approximate minimum deglaciation age is suggested by several radiocarbon-dated sediments from smaller lakes within Torneträsk’s catchment (e.g., Barnekow et al., 1998; Bigler et al., 2003; Shemesh et al., 2001) as well as by cosmogenic radionuclide dating of glacially scoured surfaces and erratics from the area (Stroeven et al., 2002). Bedrock is exposed or only covered by thin soils (<1m) in large parts of the catchment (~43%; SMHI;
https://vattenwebb.smhi.se/), particularly at higher elevations in the mountains overlooking western Torneträsk, whereas glacial, glaciofluvial and fluvial deposits occur in confined areas at lower elevations. In lower lying areas, mires have developed in local depressions (Melander 1977; SGU database; maps2.sgu.se/kartgenerator/maporder_sv.html).

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

In addition to sediments from Torneträsk, we also recovered and analyzed a sediment sequence from Abiskojaure (68° 18' N, 18° 36' E; 488 m a.s.l.; Fig. 1) to provide a more comprehensive view on changes in catchment dynamics and aid in constraining the chronology of core Co1282 by cross-correlation (see 4.2.2 Chronology). Abiskojaure’s watershed, located in the SW watershed of Torneträsk, is one of the main sub-watersheds (368 km²) and drains via Abiskojåkka into the lake. The lake has a residence time of c. 0.2 years, a surface area of 2.8 km², an average and max. depth of 16.4 m and 35 m, respectively (SMHI; https://vattenwebb.smhi.se/). Birch forest occurs along the river corridor up to ~700 m a.s.l. and covers <10% of the watershed, while most of the watershed consists of sparsely vegetated alpine
tundra with thin soils (<1m) or unvegetated areas with exposed bedrock (~78% of the watershed).

Glaciers cover about 1% of Abiskojaure’s watershed (SMHI; https://vattenwebb.smhi.se; SGU database; maps2.sgu.se/kartgenerator/maporder_sv.html).

Earlier paleoecological studies in the Torneträsk watershed did not identify distinguishable environmental impacts in response to human activities, which were likely minor and limited to fishing and hunting up to the 17th century when more intensive reindeer husbandry started to develop (Emanuelsson, 1987). During the past century, infrastructure developments in the area included railroad (built AD 1898-1902) and highway (built AD 1980-1984) construction along the southern shore of the lake.

Table 1. Meteorological data for three weather stations across the Torneträsk watershed (West to East) for the period 1961-1990 (http://www.smhi.se/klimatdata/meteorologi/temperatur/dataserier-med-normalvarden-1.7354).

Winter is defined as the period October to April and summer as the period May to September.

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<th>Torneträsk</th>
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3. Material and methods

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

3.2. Age-depth modeling

Several techniques were employed to date the composite core from Torneträsk, including $^{210}\text{Pb}$, accelerator mass spectroscopy (AMS) radiocarbon, optically stimulated luminescence (OSL) dating, and, after difficulties with radiocarbon and OSL dating of the Torneträsk sediment, cross-correlating significant geochemical changes in the record to those recorded in the AMS macrofossil radiocarbon-dated sediment record from Abiskojaure. Near-surface sediments were radiometrically dated by analyzing $^{210}\text{Pb}$, $^{226}\text{Ra}$, $^{137}\text{Cs}$ and $^{241}\text{Am}$ by direct gamma assay in the Environmental Radiometric Facility at University College London. The resulting $^{210}\text{Pb}$
chronology was calculated using the constant rate of $^{210}\text{Pb}$ supply (CRS) dating model (Appleby, 2001).

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

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

3.3. Geochemistry

Prior to analyses, all sediment samples were freeze-dried and ground using a swing mill for samples from Abiskojaure and core Co1282 and a planetary mill for the surface samples, respectively. Major and trace element geochemistry was measured on 0.2 g sample material by wavelength dispersive X-ray fluorescence using a Bruker S8 Tiger spectrometer equipped with an Rh anticathode X-ray tube. Precision was within ±3% for all elements used in this study except for Zr (±6%), and accuracy was within ±5% except for Na (±10%), Mg (±8%) and Zr
(±6%) (Rydberg, 2014). To infer compositional changes in the minerogenic fraction related to the
degree of weathering, we used the K/Al ratio and the more comprehensive ratio
(Na₂O+MgO+K₂O+CaO)/TiO₂. These ratios can provide information about the weathering
degree of the detrital siliciclastics because the comparatively more mobile alkali and alkaline
earth elements are preferentially leached from aluminosilicates compared with immobile Al or Ti
during chemical weathering (cf., Kauppila and Salonen, 1997; Parker, 1970; Roy et al. 2008).
The ratio (Na₂O+MgO+K₂O+CaO)/TiO₂ is hereafter referred to as the Chemical Index (CI),
where lower values would indicate a more intensive chemical weathering. However, changes in
these ratios must be interpreted with caution because they can also be influenced by, e.g., changes
in grain-size or source of the deposited material (Boyle 2001). To assess potential concurrent
changes in grain-size or source material and their potential influence on the CI, we used K/Rb
and Zr/Ti ratios. K/Rb can provide information about grain-size changes because K is enriched in
feldspars, which occur primarily in the coarser grain fraction, while Rb is common in micas and
clays that are more abundant in the fine fraction. Lower ratios indicate finer-grained and higher
ratios coarser-grained material (Kylander et al. 2013). Zr/Ti varies depending on the type of
parent material, and has thus been used to identify possible changes in the source of material
when assuming equal mobility of both elements during low-temperature weathering (Fabel et al.

Biogenic silica (bSi) in the sediment of Torneträsk was determined by Fourier transform infrared
(FTIR) spectroscopy using a Bruker Vertex 70 equipped with a MCT (mercury-cadmium-
telluride) detector, a KBr beam splitter, and a HTS-XT accessory unit (multi-sampler). 11 mg of
sample material was mixed with 0.5 g of oven-dried spectroscopic grade potassium bromide
(KBr) for the analysis. Quantifications are derived from a PLSR calibration model based on
synthetic sediment mixtures with defined bSi content. For more specific details about the method, analytical procedure and instrument setup see Meyer-Jacob et al. (2014a). Total carbon (TC), total nitrogen (TN) (from which we calculated atomic C/N- ratios), $\delta^{13}C$ and $\delta^{15}N$ were analyzed using a Flash EA 2000 elemental analyzer coupled to an isotope ratio mass spectrometer (Thermo Fisher Scientific). Samples were not treated with HCl prior to elemental analysis because FTIR analyses showed that the samples did not contain inorganic carbon/carbonates, which have very distinct FTIR spectral characteristics (cf., Rosén et al. 2011, Vogel et al. 2008). TC is therefore considered as equivalent to total organic carbon (TOC). Element and bSi concentrations are expressed as percentage by weight. Accumulation rates of bSi ($AR_{bSi}$; g·m⁻²·yr⁻¹) were calculated according to Eq. (1):

$$AR_{bSi} = SR \times DBD \times %bSi \times 10^2,$$

(1)

where SR is the sedimentation rate (cm·yr⁻¹), DBD is the dry bulk density (g·cm⁻³), and %bSi is the bSi concentration of the sample.

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

3.4. Diatom analysis

Diatom slides were prepared using standard methods including digestion with 30% H₂O₂ as described in Battarbee et al. (2001). Rinsed diatom samples were dried on coverslips and permanently mounted onto microscope slides using Naphrax. Diatom identification followed mainly Krammer and Lange-Bertalot (1986-91) and the guidelines of the Surface Water
Acidification Programme (SWAP) (Stevenson et al., 1991). For stratigraphical display, diatom counts are expressed as relative abundances.

3.5. Oxygen and silicon isotope analyses of diatom silica

Diatom samples for isotope analyses were purified following the cleaning stages described in Morley et al. (2004), which include removal of OM, sieving, differential settling and density separation. Prior to analyses, diatom sample purity was assured by visual inspection via SEM and by micro XRF measurements. $\delta^{18}$O and $\delta^{30}$Si were analyzed using a step-wise fluorination method. The outer hydrous layer of the diatoms was removed in a pre-fluorination stage using a BrF$_5$ reagent at low temperature. This was followed by a full reaction at high temperature to liberate oxygen (that was converted to CO$_2$ and measured for $\delta^{18}$O$_{diatom}$) and silicon (collected as SiF$_4$) isotopes that were measured using a MAT 253 dual-inlet mass spectrometer. $\delta^{18}$O values were converted to the VSMOW scale and $\delta^{30}$Si values were converted to the NBS28 scale, both using the within-run laboratory standard BFC$_{mod}$.

3.6. Interpolation of spatial variability

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

4. Results and discussion

4.1. Spatial variability of recent sediment composition

TOC, atomic C/N ratios, bSi and the CI are shown in Figure 2 to exemplify the spatial variability in the surface sediments of Torneträsk. TOC concentrations are generally low, ranging from 0.2
C/N ratios vary from 10.2 to 17.1 (mean: 12.9), which suggests both aquatic and terrestrial OM (Meyers and Ishiwatari, 1993). bSi and the CI exhibit strong spatial variability with values in the ranges of 0–46.5% (mean: 14.6%) and 10.4–16.0 (mean: 11.9), respectively. Variations in the surface (i.e., recent) sediment composition are mainly controlled by i) the proximity to riverine input, ii) differences in catchment characteristics that result mainly from the contrasting landscapes in the western and eastern parts of the catchment, and iii) the heterogeneous bedrock geology. Close to river inlets such as Abiskojåkka, terrestrial-derived OM explains elevated C/N ratios and an increased riverine input of clastic material causes a dilution of TOC and bSi concentrations. Hydroacoustic measurements in Torneträsk indicate that sediment accumulation is up to 8 times higher in proximity to major river inlets compared to rates in distal areas; for example, the sediment thickness is >8 m near the Abiskojåkka inlet but less than 2 m in more distal parts of the western basin (Vogel et al., 2013).

In addition to proximity to river inlets, TOC, C/N ratios and bSi show differences from W to E across Torneträsk. When divided into a western (n = 29) and an eastern (n = 15) subset, mean TOC, C/N ratios and bSi values differ significantly between lake basins with 2.4 versus 3.4% TOC, C/N ratios of 13.5 versus 11.7, and 10.1 versus 23.3% bSi in the western and eastern half, respectively. This pattern can be explained by the likely different style and intensity of erosion and weathering in the western compared to the eastern watershed of Torneträsk. In the western watershed, higher relief energies, large areas of exposed bedrock and sparser vegetation cover would promote the erosion of chemically less-matured substrates. Similarly, low rates of microbially catalyzed weathering processes and OM turnover in soils would limit the release of nutrients and dissolved load. In contrast, the eastern catchment, with lower relief energies, thicker soils, a denser/more productive vegetation cover and more extensive peatlands, would favor an
enhanced release of nutrients/dissolved load. This is supported by several studies that show increased silicate weathering rates and inorganic P release with increasing vegetation cover in previously glaciated regions (Anderson et al., 2000; Humborg et al., 2004; Moulton et al., 2000). These differences in catchment characteristics may explain the contrasting patterns of sediment composition indicated by the proxies for aquatic production (bSi, TOC) and organic matter source (i.e., C/N). The CI does not show a distinct spatial pattern in Torneträsk (Fig. 2), because the mineral composition within the lake basin reflects not only the degree of weathering but also the heterogeneity of the bedrock geology in the Torneträsk area.
Figure 2a) Bathymetric map of Torneträsk with 40 m contour intervals based on bathymetric measurements from 1920/1921. b-e) Spatial variability of total organic carbon (TOC), C/N ratios, biogenic silica (bSi), and the chemical index (CI) in the surface sediment (0-2 cm) of Torneträsk. Color-coded dots indicate site-specific values, whereas contour lines were interpolated using IDW interpolation in ArcGis 10.2.1 and indicate interpolated ranges.

4.2 Long-term variations (core Co1282)

4.2.1 Lithology

The 188.2-cm-long sediment core Co1282 comprises a record of the depositional history in the western Torneträsk basin from the end of the Weichselian glaciation to the onset of lacustrine sedimentation until the present. The sequence can be subdivided into four lithofacies. The basal lithofacies A (188–178 cm blf; Fig. 3) is characterized by a grey, massive and dense matrix-supported diamicton with gravel and sand in a clay matrix. Due to its specific characteristics, particularly the high density of the material in combination with the glacigenic setting, we interpret the diamicton as a basal moraine deposit/lodgement till. The diamicton is overlain by lithofacies B (178–168 cm blf; Fig. 3), which is composed of unconsolidated and poorly sorted gravely-coarse sand and thus likely of glacio-fluvial origin. The following lithofacies C (168–117 cm blf; Fig. 3) appears thinly bedded and consists of light-grey clayey silt that is interrupted by normally graded layers of slightly coarser material (silt). This facies was likely deposited in a distal proglacial environment during the ice-dammed lake phase in the Torneträsk basin. The rhythmical lamination of the fine-grained sediments would result from variations in the sedimentation regime (suspension settling) due to fluctuations in intensity and sediment load of meltwater discharge coming from the retreating ice sheet (Carrivick and Tweed, 2013).
Lithofacies D (117–0 cm blf; Fig. 3) is characterized by homogenous silty clay with a color change from dark grey at the bottom to brown-grey towards the surface. Deposition of lithofacies D commenced after the retreat of the Fennoscandian ice sheet from the area and represents the continuous hemipelagic lacustrine sedimentation in Torneträsk during the Holocene. Between 15–7 cm blf, distinct horizons occur that consist of Fe-rich (up to 8.9%) reddish layers overlain by diffuse dark lenses rich in Mn (up to 2.5%; Fig. 3). Similar layers have been reported for sediments of other large, oligotrophic lakes, such as Lake Baikal where they have been ascribed to enrichment of redox sensitive elements at the redox front in low sedimentation rate settings (Och et al., 2012).
Figure 3. Lithofacies, color linescan, lithological description, and concentrations of Mn, Fe, Al, and K, as well as the Chemical Index (CI) and K/Al, K/Rb, and Zr/Ti ratios for core Co1282, recovered from the western basin of Torneträsk.

4.2.2 Chronology

The chronology of core Co1282 from Torneträsk is constrained by $^{210}$Pb dating as well as cross-correlating significant geochemical changes to those in the AMS radiocarbon-dated sediment
record from Abiskojaure (Figs. 4 and 5). Total \(^{210}\)Pb activities in the sediment reached equilibrium with the background (supported) \(^{210}\)Pb activity at 1.5–2 cm, which indicates a low sedimentation rate of \(-0.015\) cm·yr\(^{-1}\) in the near-surface sediments (Fig. 4a, b).

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

Because of the large age overestimates obtained from both radiocarbon and luminescence dating, these results could not be used to constrain a robust age-depth relationship in core Co1282. Instead, to constrain the chronology we cross-correlated significant geochemical changes in core Co1282 with those clearly observed in the core from Abiskojaure for which a chronology was established based on AMS radiocarbon dated woody plant macrofossils (twig fragments) (Table 2, Fig. S2). Four sections with similar patterns of change across both sediment sequences can be
identified: 1) high CI values (>12) and low Si/Al ratios in deeper sediments: 82-117 cm in Torneträsk and >184 cm in Abiskojaure; 2) declining CI from >12 to ~11, increasing bSi trend: 82-56 cm/184-104 cm (Abiskojaure’s sediment composition is more variable during this period, reflecting the coring location in close proximity to the inlet); 3) low CI (<11), high Si/Al: 56-12 cm/104-29 cm; and 4) in recent sediment a shift to higher CI (~11 or higher) and drop in Si/Al 12-0 cm/29-0 cm (Fig. 4). Only three macrofossils with sufficient C could be extracted from the sediment core of Abiskojaure; however these macrofossils originate from sample levels at or near the pronounced changes in geochemical properties recorded in Abiskojaure’s sediments and thus provide satisfactory age-depth control for the timing of major changes in the sediment composition of Torneträsk. Similar geochemical changes in both cores suggest that the recorded dynamics in core Co1282 are representative for the western Torneträsk catchment and/or dominated by the riverine sediment input from Abiskojäkka, which drains Abiskojaure. Because of the potential time lag that may exist for when the changes were recorded in Torneträsk and the upstream-located Abiskojaure, inferred ages for core Co1282 based on the cross-correlation to Abiskojaure should be considered as minimum ages.

We constrained the basal age for the lacustrine sediments (117 cm depth) in core Co1282 based on radiocarbon-dated sediment records from small lakes in this area and cosmogenic exposure ages of glacially-scoured surfaces and erratics, which indicate a deglaciation age around 9500±250 cal yr BP (e.g., Barnekow et al., 1998; Bigler et al., 2003; Shemesh et al., 2001; Stroeven et al. 2002) (Fig. 4). This would imply an average Holocene sedimentation rate of 0.01 cm yr⁻¹, which is consistent with the recent sedimentation rate indicated by ²¹⁰Pb dating.
Table 2. Calibrated radiocarbon ages for the sediment sequences from Abiskojaure and Torneträsk.

<table>
<thead>
<tr>
<th>Laboratory ID</th>
<th>Sample ID</th>
<th>Composite depth (cm)</th>
<th>(^{14}\text{C} ) age (yr BP)</th>
<th>Calibrated age range (2(\sigma)) (cal yr BP)</th>
<th>(\delta^{13}\text{C} )</th>
<th>Material</th>
</tr>
</thead>
<tbody>
<tr>
<td>Beta-440907</td>
<td>AJ2016In_K32-33</td>
<td>32.5</td>
<td>950 ± 30</td>
<td>930–790</td>
<td>–26.8</td>
<td>Woody plant macrofossil</td>
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<tr>
<td>Beta-440908</td>
<td>AJ2016In_P85.5</td>
<td>95</td>
<td>2200 ± 30</td>
<td>2320–2130</td>
<td>–27.4</td>
<td>Woody plant macrofossil</td>
</tr>
<tr>
<td>Beta-442326</td>
<td>AJ2016In_P175</td>
<td>184.5</td>
<td>5970 ± 30</td>
<td>6885–6735</td>
<td>–28.0</td>
<td>Woody plant macrofossil</td>
</tr>
<tr>
<td>2234.1.1</td>
<td>Co1282-1_0-1</td>
<td>0.5</td>
<td>2772 ± 19</td>
<td>2926–2795</td>
<td>–25.4</td>
<td>Org C</td>
</tr>
<tr>
<td>2633.1.1</td>
<td>Co1282-1_1.5-2</td>
<td>1.75</td>
<td>4533 ± 22</td>
<td>5310–5055</td>
<td>–32.0</td>
<td>Org C</td>
</tr>
<tr>
<td>2235.1.1</td>
<td>Co1282-1_16-17</td>
<td>16.5</td>
<td>4867 ± 23</td>
<td>5646–5588</td>
<td>–24.5</td>
<td>Org C</td>
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<tr>
<td>2236.1.1</td>
<td>Co1282-3I_29-30</td>
<td>36.5</td>
<td>6290 ± 23</td>
<td>7263–7171</td>
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<td>2237.1.1</td>
<td>Co1282-3I_53-54</td>
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<td>8752 ± 28</td>
<td>9891–9613</td>
<td>–22.6</td>
<td>Org C</td>
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<tr>
<td>2238.1.1</td>
<td>Co1282-3III_93-94</td>
<td>100.5</td>
<td>15914 ± 57</td>
<td>19413–18988</td>
<td>–18.2</td>
<td>Org C</td>
</tr>
</tbody>
</table>
Fig. 4. Correlation between core Co1282 from the western Torneträsk basin and the core from Abiskojaure using the chemical index and Si/Al ratios (note differences in x- and y-axis scaling and range). Radiocarbon dates are given as $^{14}$C ages (yr BP) and dashed lines indicate sections of similar change used for correlating the two cores.
Figure 5. Chronology of core Co1282 from Torneträsk. a) Total (black) and supported (red) $^{210}\text{Pb}$ concentrations versus depth. b) Radiometric chronology showing the CRS model $^{210}\text{Pb}$ dates. c) Bayesian age-depth model based on $^{210}\text{Pb}$-, dates inferred from the cross-correlation to the radiocarbon-dated sequence from Abiskojaure (transparent green) and deglaciation age of the area (transparent blue). 'Best' model based on the weighted mean age for each depth (red dashed line) and 95% confidence intervals (grey dashed lines). Greyscale intensity indicates likelihood of calendar ages. Upper panel (left to right) shows the Markov Chain Monte Carlo iterations (log of objective), the prior (green curves) and posterior (grey histograms) distributions for the accumulation rate and memory, respectively.

4.2.3 Holocene climatic and environmental development

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

The proglacial sediments that accumulated in Torneträsk during its ice-dammed lake phase are characterized by a highly variable geochemistry compared to the subsequent lacustrine sediments. For example, major element concentrations fluctuate frequently between minimum 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
variability, corresponding to the rhythmical bedding/lamination of the proglacial sediments, is likely related to variations in grain-size, indicated by the K/Rb ratio ranging from 198 to 270, as a result of the varying intensity and sediment load of the meltwater discharge from the retreating ice sheet (Carrivick and Tweed, 2013).

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

The initial hemi-pelagic lacustrine sediments deposited after the deglaciation of the area show only minor compositional variations. Major element concentrations are stable with values around 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 constant throughout this period with values around 0.35, 221, 0.042 and 12.4, respectively (Figs. 3 and 6). bSi concentrations are continuously low (<0.1%), which is consistent with a very low diatom abundance that in turn shows a stable diatom community composition (Fig. 7). Proxies associated with OM exhibit slight increases from 0.4 to 0.7% for TOC, from 0.03 to 0.05% for TN, from ~16 up to 18 for C/N ratios, and from +2.7 to +3.8‰ for δ¹⁵N, whereas δ¹³C decreases from –22.1 to –23.2‰ (Fig. 6). In the corresponding sediment section from Abiskojaure, the CI is also high and relatively stable (~12) except for three samples associated with two clay layers (Fig. 4).

The constant major element concentrations, elemental ratios, and CI indicate that there are no qualitative changes in the composition of the minerogenic matter input from the catchment during the early Holocene. For example, the high CI is similar to the mean value of the proglacial sediments (12.5), which shows that the minerogenic material deposited during this period is essentially chemically unaltered. These data indicate a homogeneous sediment source, relatively
low chemical weathering rates and the presence of easily erodible and chemically unaltered glacigenic deposits in the Torneträsk area. Low weathering rates would limit the release of DSi and other nutrients into the aquatic system, and explain the low bSi concentrations and low diatom abundance in this part of the record.

The slight increase in C/N indicates a gradually increasing export of terrestrial OM into the lake, which would result from the gradual development of vegetation cover and soil formation in the catchment. This initial landscape development would similarly explain the observed trends for TOC, TN, δ15N and δ13C. Conclusive interpretations of the δ13C and δ15N trends are complicated by the complex nature of bulk OM isotope signals (Meyers and Ishiwatari, 1993; Meyers, 2003) and the low magnitude of change of ~2‰ throughout the early and mid-Holocene; however Hammarlund (1993) and Hammarlund et al. (1997) suggested that early Holocene declines in δ13C were the result of the transition from a deglaciated, vegetation-free landscape to tundra and finally forest vegetation during the initial landscape development. The authors proposed that the development of vegetation and soils enhanced soil respiration, which led to an increased export of 13C-depleted carbon dioxide to the aquatic system and subsequently to a 13C-depletion in phytoplankton.

Altered soil biogeochemical processes in response to a gradual development of vegetation cover would likewise account for the increase of δ15N in core Co1282. In general, bulk OM δ15N increases in soils with depth and OM age, which has been shown for soil profiles in the Torneträsk area (Makarov et al., 2008). During decomposition, the heavier 15N is gradually enriched in the residual OM by isotopic fractionation (Högberg, 1997). Correspondingly, the increasing δ15N trend in core Co1282 would result from enhanced soil OM decomposition and
increased leaching of $^{15}$N-enriched nitrate in the course of progressing soil and vegetation development in the Torneträsk catchment.

Figure 6. a) Chemical Index (CI), b) biogenic silica (bSi) and C/N ratios, c) total organic carbon (TOC) and nitrogen (TN), d) $\delta^{13}$C, and $\delta^{15}$N, and e) K/Rb and Zr/Ti ratios versus inferred sediment ages for core Co1282 from Torneträsk over the Holocene.
Figure 7. Relative abundance of selected diatom taxa for core Co1282 from Torneträsk over the Holocene. The scale at the base of each panel is given in percentages.

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

Compared to the relatively uniform composition of the early Holocene sediments, the sediment composition gradually changed between ~6600 and ~3400 cal yr BP. Approximately 3000 years after deglaciation, concentrations of K and Al started to progressively decline from ~2.9 to ~2.2% and from ~8.0 to 6.8%, respectively. Simultaneously, K/Al ratios and CI decreased from ~0.35 to ~0.33 and from ~12.2 to 11.2, while K/Rb and Zr/Ti ratios remained stable with values fluctuating around 219 and 0.043, respectively (Figs. 3 and 6). bSi levels slowly increased from <0.1 to ~2%, whereas C/N ratios decreased from ~18 to 14.4. Other proxies associated with OM follow their early Holocene trends and continuously increased from 0.7 to 1.0% for TOC, from
0.05 to 0.08% for TN, and from +3.8 to +4.5‰ for δ¹⁵N, while δ¹³C continued to decrease from –23.2 to –24.0‰ (Fig. 6). The CI in the sediment record from Abiskojaure likewise declined during this period from ~12 to ~11.4 (Fig. 4).

The distinct declines in K/Al ratios and CI indicate a qualitative change in the sediment composition that is likely driven by the depletion of mobile alkali and alkaline earth elements through chemical weathering of mineral matter in catchment soils because elemental ratios indicative for changes in grain-size (K/Rb) and source area (Zr/Ti) remained unchanged during this period. Studies from Lake Kilpisjärvi, northern Finland (Kauppila and Salonen, 1997), and from NW America (Whitehead et al., 1989) demonstrated that increased chemical weathering rates (and a loss of more mobile elements) followed the establishment of denser forests in previously glaciated areas. We therefore assume that the qualitative geochemical changes in core Co1282 are likewise driven by the widespread establishment of denser pine-birch forests, which commenced at ~7600 cal yr BP and reached maximum density during the HTM at ~6300 cal yr BP in the western Torneträsk catchment (Barnekow, 2000; Barnekow and Sandgren, 2001). The climate optimum during the HTM (~8000–4800 cal yr BP), with temperatures 1.5-2.0 °C warmer in northern Fennoscandia than at present (Barnekow, 2000; Barnekow and Sandgren, 2001, Seppä et al., 2009) (Fig. 8), also had an impact on in-lake processes and the diatom community composition in Torneträsk. The diatom assemblage shows a gradual shift with an increased abundance of Cyclotella comensis and decrease in Aulacoseira ambiguа and subarctica (Fig. 7) likely in response to elevated temperatures during this period. Warming and earlier ice break-up may enhance the strength and duration of thermal stratification, which favors more buoyant Cyclotella taxa over heavier Aulacoseira taxa (Rühland et al., 2015). Bigler et al. (2006) found a similar shift in the diatom community in the small lake Vuolep Njakajaure (Fig. 1 F); however,
this shift occurred ~1000 years earlier, which indicates the slower (delayed) response to climate
change in the large and deep Torneträsk as also shown for other (sub)arctic lakes (Rühland et al.,
2015 and references therein).

The proceeding catchment transformation from bare bedrock and glacigenic sediment deposits
after deglaciation to a forested catchment supporting soil formation during the mid-Holocene is
corroborated by the consistent trends of δ¹⁵N and δ¹³C, assuming that biogeochemical processes
in soils primarily control trends in stable C and N isotopes in the Torneträsk sediment during the
early lake development (cf., Hammarlund, 1993; Hammarlund et al., 1997; Wolfe et al., 1999).

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

Changes in the availability of DSi can be traced using Si isotope analyses of diatom silica. In core
Co1282, we could extract sufficient material for diatom isotope analyses from 76 cm blf (~6000
cal yr BP) upwards when bSi concentrations were ≥0.2%. δ³⁰Si_diatom decrease from +0.37 to 0‰
between c. 6000 and 3000 cal yr BP, followed by lower values fluctuating around 0‰ (min: –
0.10, max: +0.15‰) during the remainder of the late Holocene until present (Fig. 8). Diatoms
preferentially incorporate ²⁸Si compared to ³⁰Si (and ²⁹Si), and consequently with enhanced
productivity and utilization of DSi, $\delta^{30}\text{Si}$ ($^{30}\text{Si}/^{28}\text{Si}$) progressively increases in the DSi pool. This productivity signal is recorded in the $\delta^{30}\text{Si}$ of diatoms because diatoms then incorporate DSi from this progressively $^{30}\text{Si}$-enriched pool. Variations in the DSi supply can also alter the $\delta^{30}\text{Si}$ signal of diatoms by affecting the biological DSi demand (Leng et al., 2009; Swann et al., 2010). In the sediments of Torneträsk, the initial $\delta^{30}\text{Si}_{\text{diatom}}$ decline indicates increased DSi utilization during the earlier lake development. Because diatom productivity was very low during this period, we interpret the increased utilization as an indication for a higher DSi demand that resulted from the initially low weathering rates and thus low DSi supply. This suggests that diatom productivity in Torneträsk was Si limited during the early lake ontogeny, which delayed the onset of notable diatom production by $\sim$3000 years after deglaciation until vegetation development facilitated Si export. Alternatively, an increase in weathering rates could also contribute to the initial $\delta^{30}\text{Si}$ decline; for example, Opfergelt and Delmelle (2012) found that $\delta^{30}\text{Si}$ decreased in soil profiles with increasing weathering degree.
Figure 8. a) July Insolation at 65°N (Berger and Loutre, 1991), b) LOESS smoothed (span: 0.05) stacked summary curve of the pollen-based temperature variability for Northern Europe (Seppä et al., 2009), c) $\delta^{18}$O_diatom for Lake 850 (Fig. 1 I, Shemesh et al., 2001), percentage terrestrial pollen from d) *Pinus sylvestris* and shrub vegetation in the sediment record of Vuolep Njakajaure (Fig. 1 F; Barnekow, 2000) compared to e) the Chemical Index (CI), f) biogenic silica (bSi) concentrations and accumulation rates ($AR_{bSi}$), g) diatom $\delta^{18}$O and $\delta^{30}$Si in the sediment core Co1282 from Torneträsk over the past 6000 and 10000 years, respectively.

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

Following the long-term vegetation expansion and catchment stabilization during the mid-Holocene, most organic and inorganic geochemical properties in core Co1282 show no major
variations between ~3400 and 750 cal yr BP. Major element concentrations, K/Al and Zr/Ti 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, and 10.9 for CI (Fig. 3). C/N ratios, δ¹³C, and δ¹⁵N are relatively stable during this period with values around 14.4, –24.1‰, and +4.5‰, respectively, while TOC and TN concentrations continued to increase slightly from 1.0 to 1.6% and from 0.08 to 0.12%, respectively (Fig. 6). In contrast, bSi increased from 2 to above 7% over the whole period, whereas K/Rb ratios started to decline around 2200 cal yr BP from values around 220 during the early and late Holocene to ~200 by ~750 cal yr BP. Fe and Mn co-vary during this part of the sediment record and layers enriched in Fe and Mn are most likely buried relicts of the active redox horizon that continually moves upward with progressing sediment accumulation (Och et al., 2012) (Fig. 3). The CI in the corresponding sediment section from Abiskojaure reached its lowest values during the Holocene and stabilized around 10.7 (Fig. 4).

The relatively constant sediment geochemistry between ~3400 and 750 cal yr BP indicates no larger changes in the composition of the catchment-derived sediment delivered to the coring site in the western lake basin. In comparison, studies of smaller lakes at higher altitudes in the catchment suggest increased soil erosion during the late Holocene from 3200–2600 cal yr BP onwards, respectively (Berglund et al., 1996; Snowball and Sandgren, 1996; Rubensdotter and Rosqvist, 2003). For the late Holocene, several paleolimnological studies in the Torneträsk area indicate a climate cooling that led to a tree-line retreat, which would expose previously forested soils and thus facilitate soil erosion. Pollen and macrofossil records indicate a lowering of the pine tree line by ~175 m and also of mountain birch already from ~4500 cal yr BP, which is assumed to represent a decline of the growing season temperature of ~1.5-2°C compared to today. This tree-line retreat ultimately led to the establishment of today’s more sparse subalpine
Birch forests and an increased presence of dwarf shrubs, grasses and sedges across the landscape also at lower altitudes (Barnekow, 1999, 2000) (Fig. 8). Temperature reconstructions based on diatoms, chironomids and pollen likewise indicate a pronounced cooling during the late Holocene by ~1-2°C, but depending on the site and the chosen biological indicator the timing of the cooling onset varies between ~3500 and 1900 cal yr BP (Bigler et al., 2002; 2006). These landscape dynamics agree well with the proposed neoglacial expansions of the Kårsa glacier (Fig. 1 B) around 3300 cal yr BP (Berglund et al., 1996; Snowball and Sandgren, 1996) and of the Kalanvare glacier, located just outside the Torneträsk catchment (Fig. 1 A), at c. 4400, 3000, 2000, and after 1200 cal yr BP (Rosqvist et al., 2004). Increased glacial activity would increase production of finer-grained material (rock flour), which would ultimately be accumulated in Torneträsk. Declining K/Rb ratios in core Co1282 indicate such an increased contribution of finer-grained material from ~2200 cal yr BP onwards. This trend is corroborated by grain-size analyses from a short core that was recovered close to the coring site of core Co1282, which shows an increasing clay content over the corresponding depth interval (Vonk et al. 2012).

The continuous increase of bSi concentrations during this period would suggest a progressive improvement of environmental conditions for aquatic production but accumulation rates of bSi reveal instead that diatom production reached a somewhat stable level with AR_{bSi} around 4.5±1 g·m⁻²·yr⁻¹ after the initial increase (Fig. 8). The overall low AR_{bSi} in Torneträsk is of the same order of magnitude as those reconstructed for other large lakes such as Lake Baikal (Qiu et al., 1993) or Lake El’gygytgyn (Meyer-Jacob et al., 2014b) in Siberia. Regardless of changes in bSi content and flux there was no significant change in the diatom assemblage (Fig. 7).

Soil erosion during the Little Ice Age (13–1 cm blf; lithofacies D; ~750-50 cal yr BP)
The relatively stable sediment geochemistry during the late Holocene was interrupted by a pronounced shift in the sediment composition commencing at ~750±275 cal yr BP contemporaneous with the beginning of the LIA (Matskovsky and Helama, 2014; Melvin et al., 2012; Seppä et al., 2009). Major element concentrations, K/Al ratios, and the CI shifted to higher values of ~2.6% for K, ~7.2% for Al, ~0.36 for K/Al, and ~11.7 for the CI, while Zr/Ti and K/Rb show no major changes and remain around 0.043 and 200, respectively (Fig. 3). bSi concentrations dropped to ~5%, whereas C/N ratios and δ^{15}N increased to maximum values of 19.4 and +5.8‰. δ^{13}C slightly increases only initially to –23.6‰ and drops to Holocene minimum values of –25.5‰ at the sediment surface, while TOC and TN follow their Holocene trend and reach maximum values of 2.6 and 0.22% in the most recent sediment (Fig. 6). In the corresponding sediment from Abiskojaure, the CI also shifted to significantly higher values (~11.1) (Fig. 4).

Elevated K/Al ratios and CI, together with constant Zr/Ti and K/Rb ratios, indicate a stable grain-size distribution and sediment source but that the minerogenic material deposited in Torneträsk from ~750 cal yr BP onwards is less-intensively weathered compared to sediment that accumulated since the beginning of the late Holocene cooling. Less-weathered material would come from deeper mineral soil layers, and together with an increase in C/N ratios indicating more terrestrial OM, this signifies enhanced soil erosion in the Torneträsk area. An enhanced erosion of catchment soils is also supported by the simultaneous increase of δ^{15}N in the sediment because OM from deeper soils is enriched in ^{15}N (Högberg, 1997); for example, Makarov et al. (2008) showed that deeper soil horizons (>10 cm soil depth) in the Abisko area exhibit δ^{15}N of >+5‰.

During the LIA, temperatures were ~1 °C cooler than at present in Fennoscandia with coldest tree-ring inferred summer temperatures during the 13th through 19th centuries for the Torneträsk
Paleoecological temperature reconstructions from smaller lakes in the Torneälven watershed likewise suggest a ~1 °C cooling over the last millennia (e.g., Bigler et al., 2002, 2003; Hammarlund et al., 2002). The pronounced cooling during this period is emphasized by the formation of permafrost in the area after 700 cal yr BP (Kokfelt et al., 2010; Rydberg et al., 2010). Neo-glacial activity also increased over the last millennium until the end of the LIA (Rosqvist et al., 2004; Snowball and Sandgren, 1996), which is reflected in the Torneälven record by a higher input of finer-grained material indicated by the lowest K/Rb ratios recorded for the Holocene during this period (Fig. 6). However, increased glacial activity alone cannot explain the change in sediment composition during the LIA because, while increased glacial erosion would have exposed less-weathered material, it would not have altered the quality of the catchment-derived OM.

Carbonate and diatom silica δ¹⁸O from several small-lake records in the study region indicate changes in atmospheric circulation patterns over the Holocene, and particularly during the late Holocene (e.g., Rosqvist et al., 2004, 2007; Shemesh et al., 2001). In the sediment sequence from Torneälven, δ¹⁸O in diatom silica declined by 4.5‰ during the past ~6000 years. The δ¹⁸O decline was relatively slow between ~6000 and 750 cal yr BP with a total decrease of ~1.5‰, but accelerated substantially during the past ~750 years when δ¹⁸O declined by an additional ~3‰ (Fig. 8). The successive long-term decline over the Holocene has been ascribed to changes in the relative contribution of different air masses to the area, changing from a maritime climate with a strong zonal airflow from W/SW to a more continental climate with a dominant meridional airflow from N/NE (Hammarlund et al., 2002; Shemesh et al., 2001).
Compared to the slight decline (1.5‰) over most of the Holocene, the strongly amplified decline (3‰) in diatom δ¹⁸O during the past ~750 years requires a more significant change in climate. The main factors controlling diatom δ¹⁸O are lake-water temperature and the δ¹⁸O composition of the ambient lake-water, which in turn depends on the δ¹⁸O composition of the local precipitation and evaporation. Several studies from smaller lakes in the Torneträsk area have suggested that temperature could not be the sole driver of the observed δ¹⁸O changes and that a 1°C temperature decline, as indicated for the LIA, alone would have caused a δ¹⁸O decline of <1‰ (e.g., Hammarlund et al., 2002; Rosqvist et al., 2013; Shemesh et al., 2001). Reduced evaporation could amplify the δ¹⁸O decline, but lake-water surveys show that lakes in the area are generally little affected by evaporation (Jonsson et al., 2009); furthermore, lake-water δ¹⁸O and δ²H values from Torneträsk itself (Shemesh et al., 2001) fall on the local meteoric water line established by Jonsson et al. (2009), which indicates that the lake is – at least at present – unaffected by evaporation.

Changes in the δ¹⁸O of precipitation is likely the most critical factor in the Torneträsk watershed and could have contributed to the pronounced δ¹⁸O decline during the past ~750 years in two ways; one is an even stronger dominance of ¹⁸O-depleted air masses, while the other is a relative increase in winter precipitation, which is more ¹⁸O-depleted compared to summer precipitation due to the temperature difference during condensation (Dansgaard 1964). Each of these has been proposed by different paleoecological studies in the area. Rosqvist et al. (2004; 2007) suggested that lake-water δ¹⁸O declined during periods dominated by dry and cool arctic air masses from N, leading to cooler summers. For the LIA, Loader et al. (2013) inferred an increased contribution of arctic air leading to cool and sunny summers from tree-ring based temperature and sunshine reconstructions for the Torneträsk area (Fig. 9).
Alternatively, Jonsson et al. (2010) suggested that increased winter precipitation and elevated winter/summer precipitation ratios led to a $\delta^{18}$O decline in a high-altitude alpine lake record in the Torneträsk catchment. However, precipitation reconstructions for the study region exhibit mixed results (Hammarlund et al., 2002; Seppä and Birks, 2001). Lower summer temperatures as well as increased winter precipitation could both contribute to neo-glaciation, which has been observed during the LIA in Fennoscandia as well as specifically in the Torneträsk area (Karlén 1988; Nesje 2009; Rosqvist et al., 2004; Snowball and Sandgren, 1996).

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

Remarkably, the pronounced shift in sediment composition in Torneträsk did not occur until the start of the LIA despite an already ongoing climate cooling for more than 2000 years. In addition,
the abrupt nature of these changes in the sediment record indicates either a rather sudden change of environmental conditions with the onset of the LIA or alternatively that environmental conditions crossed a critical threshold in the Torneträsk watershed when temperatures further declined during the LIA. Climate reconstruction for the Torneträsk region indicate that the scale of climate change during the LIA was not unprecedented during the late Holocene. For example, no qualitative changes in the deposited sediments occurred in response to the gradual 1-2°C decline since the HTM or in response to a temperature decline of equal magnitude to the LIA during a cold period ~1400 cal yr BP (Matskovsky and Helama, 2014; Seppä et al., 2009) (Figs. 8 and 9).

Sediment records from smaller lakes in the Torneträsk catchment indicate increased mobilization of minerogenic matter already from 3400-2600 cal yr BP onwards, following tree-line retreat as well as increasing glacial activity during the late Holocene (Berglund et al., 1996; Rubensdotter and Rosqvist, 2003; Snowball and Sandgren, 1996). Similarly, increased soil erosion has been reported for higher elevation lakes in other parts of the Scandinavian Mountains following the late Holocene tree-line retreat (Barnett et al, 2001; Hammarlund et al. 2004). In contrast, increased erosion of catchment soils is not recorded in sediments of Torneträsk’s western basin before the onset of the LIA when pollen records from the area show an increasingly open landscape, indicated by the further increased presence of dwarf shrubs, grasses and sedges (Barnekow, 1999, 2000; Bigler et al., 2002) (Fig. 8). The increasingly sparse vegetation cover would have further increased the susceptibility of soils to erosion, which are generally thin (~1 m), poorly developed, and have relatively shallow organic horizons (<15 cm) above the modern tree-line in the Torneträsk area (Darmody et al. 2004; Makarov et al. 2008). This is supported by radiocarbon-dated buried soils from the catchment that suggest highest deposition rates of
upslope eroded material during the LIA (Darmody et al. 2004). We propose that, following the
tree-line retreat, catchment destabilization and subsequent soil mobilization occurred first mainly
locally at higher elevations. With diminishing vegetation cover, the Torneträsk catchment
gradually became more sensitive to soil erosion through, e.g., rainfall events and snowmelt run-
off, and crossed a critical threshold around ~750 cal yr BP. From that time onwards, previously
forested soils in large parts of the catchment were increasingly exposed to erosion and less-
weathered material was also transported into Torneträsk’s western basin.

Figure 9. a) Summer (May to August) temperature reconstructions for the Torneträsk area based
tree ring maximum late-wood-density (Melvin et al., 2012) and b) dominant atmospheric
circulation based on departures between tree ring inferred temperature (X-ray density) and
sunshine (stable carbon isotope ratio), where negative values indicate a dominance of cold-sunny
weather (Arctic air dominant) and positive values indicate a dominance of warm-cloudy
conditions (maritime air dominant) (Loader et al., 2013), compared to c) biogenic silica
accumulation rates ($AR_{\text{bSi}}$) and diatom-$\delta^{18}O$ of core Co1282 from Torneträsk for the period 1600 to -58 cal yr BP.

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

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

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

Although the dating of the sediment record from Torneträsk relies on a combination of directly determined and inferred ages that we could transfer from a dated sediment record from a lake located in one of Torneträsk’s main subcatchments, the sediment record of this 330 km² lake provides valuable insights into regional landscape-scale changes over the past c. 9500 yr. The initial landscape development and accompanying lake ontogeny were the dominating influences on sediment composition in Torneträsk after deglaciation. During the early to mid-Holocene, gradual processes associated with the development of soils and vegetation drove the element cycling in the aquatic and terrestrial environment and led to an initial nutrient (Si) limitation in Torneträsk preventing notable diatom production until ~6600 cal yr BP. With progressing ecosystem stabilization, the lake became more climate-sensitive during the late Holocene when diatom production responded to climate variations (tree-ring inferred temperature) during the last millennia, and a drop in diatom δ¹⁸O suggests a larger shift in atmospheric circulation and/or precipitation patterns.

In contrast to the successive change in the sediment composition following landscape development, climate cooling during the late Holocene led to a pronounced shift in Torneträsk’s sediment composition that occurred rather abruptly (~100 years) but not before the onset of the LIA around ~750 cal yr BP. This shift marking increased soil erosion was delayed by >2000 years compared to the first indications (tree-line retreat, increased erosion) for climate cooling that are recorded in small lakes located in the Torneträsk catchment. The rather abrupt and delayed compositional change in Torneträsk suggests a non-linear threshold response to climate forcing. Following further climate cooling during the LIA, larger parts of the Torneträsk
watershed became susceptible to soil erosion, and from ~750 cal yr BP onwards environmental changes initially only registered in higher located sub-catchments more sensitive to tree-line dynamics were also registered in Torneträsk.

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

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