Tracing Holocene temperatures and human impact in a Greenlandic Lake: Novel insights from hyperspectral imaging and lipid biomarkers

T. Schneider a,b,1,*, I.S. Castañeda a, B. Zhao a, S. Krüger c, J.M. Salacup a, R.S. Bradley a

a Department of Earth, Geographic and Climate Sciences, University of Massachusetts, Amherst, MA, USA
b Swiss Federal Institute of Aquatic Science and Technology (Eawag), CH-8600 Dübendorf, Switzerland
1 National Museum of Denmark, Environmental Archaeology and Materials Science, Copenhagen, Denmark

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ABSTRACT

Global warming particularly impacts terrestrial and aquatic ecosystems in the Arctic. To constrain the sensitivity of Arctic lakes and make meaningful predictions about future change under global warming, we need to examine their response to previous warm phases. Lake sediments from Greenland’s deglaciated area offer valuable archives to investigate past climate variability and associated lake changes.

Here, we applied hyperspectral imaging and lipid biomarker thermometry to a Holocene-length sediment record from Lake 578 in the Eastern Settlement of the Norse (61.08° N, 45.62° W; ~155 m a.s.l) to investigate local temperature, productivity, and anoxia histories. We calibrated branched glycerol dialkyl glycerol tetraethers (brGDGTs) with summer mean water temperatures (SMWT) using a previously published site-specific calibration and analyzed pigment fluxes based on hyperspectral imaging. Notably, the anoxia reconstructions were corroborated with two independent proxies (GDGT-0/Crenarchaeol and bacterio pheophytins). We investigated the lake’s environmental history and identified periods of significant change by employing generalized additive models (GAMs).

Our results reveal significant transitions in Lake 578 driven both by natural climate shifts and anthropogenic impacts. During the early Holocene, low SMWT and productivity coupled with high anoxia suggest strong seasonality and prolonged inverted thermal stratification, possibly enhanced by extended ice cover. The mid-Holocene showed higher SMWT and productivity along with low anoxia, indicating a dimictic lake system. The early Holocene temperature rise lagged that of the Northern Hemisphere, but closely followed the Atlantic-Fennoscandian stack. The Holocene Thermal Maximum (7.5–4.5 cal ka BP) aligns with other regional reconstructions. After 3 cal ka BP, we observed a Neoglacial cooling characterized by increased anoxia and reduced temperatures due to enhanced stratification. At around 1.0 cal ka BP, Lake 578 saw a surge in productivity and anoxia, which we attribute to land use and lake damming by the Norse. Despite a post-Norse decline in productivity and disappearance of anoxia, the lake never reverted to its pre-Norse state, with modern sheep farming further intensifying productivity in recent decades. While early Holocene anoxia resulted from natural cold temperature stratification, anoxia during the Norse period was anthropogenically induced.

This research underscores the value of integrating lipid biomarkers with hyperspectral imaging for detailed reconstructions of changes within Arctic lakes. It provides crucial insights for anticipating the ecologic and climatic resilience of Arctic lakes to ongoing global warming and anthropogenic influence.

1. Introduction

Arctic amplification of global warming (Jansen et al., 2020; Previdi et al., 2021) is resulting in alarming rates of temperature rise in the Arctic, impacting both terrestrial and aquatic ecosystems (Jane et al., 2018; Lehnherr et al., 2018; Leppi et al., 2016; Mueller et al., 2009). Recent research on arctic lakes highlights that global warming enhances...
nutrient input from permafrost thawing (Reyes and Lougheed, 2015), increases summer thermal stratification and microbial respiration (Jane et al., 2023; Klanten et al., 2023), and reduces oxygen solubility, which together exacerbate anoxic conditions threatening aquatic ecosystems (Jane et al., 2023; Jenny et al., 2016). Conversely, a reduction in ice-cover duration and frequency may improve lake aeration and diminish anoxia, presenting a complex interplay of warming effects (Kirlin et al., 2012; Lehnherr et al., 2018; Mueller et al., 2009).

Understanding the response and sensitivity of Arctic lakes to past climatic shifts is crucial for predicting the impact of ongoing global warming. The Holocene epoch (~11.7 cal ka BP – present), marked by significant warm periods such as the Early Holocene Temperature Rise (~11.7–8.5 cal ka BP) and the Holocene Thermal Maximum (~9.5–4.5 cal ka BP), is ideal to contextualize the current global warming (Axford et al., 2021; Briner et al., 2016; Fletcher et al., 2024; Kaufman et al., 2020; Kobashi et al., 2017). Greenland’s deglaciated area comprises a wealth of lakes (Klanten et al., 2023; Leppi et al., 2016) which provide Holocene-length sedimentary records of environmental and climate change, ideally suited to study the implications of previous warmer periods on lacustrine ecosystems (Axford et al., 2021; Briner et al., 2016; Gajewski, 2015; Sundqvist et al., 2014). Briner et al. (2016) found that orbitally driven solar insolation was a main driver of Greenland’s climate during the Holocene in addition to periodic freshwater releases (Alley and Ågüstdóttir, 2005; Fleitmann et al., 2008; Jennings et al., 2015; Nesje et al., 2004; Thomas et al., 2007). However, recent studies also show that the spatiotemporal pattern of past temperature variability in Greenland is complex, which (apart from geographical differences) may also be attributed to the use of different proxies, with their inherent uncertainties and seasonality biases. This underscores the need for further reconstructions in Greenland’s deglaciated area to clarify Holocene climate variations (Axford et al., 2021; Briner et al., 2016; Kaufman et al., 2020; Sejrup et al., 2016).

Despite the necessity to understand the impact of temperature variability on lacustrine ecosystems, comprehensive studies on Holocene-length productivity in Greenlandic lakes are scarce, and anoxia reconstructions remain particularly limited. Lakes from south Greenland provide an excellent opportunity to investigate both the natural and anthropogenic influences on lacustrine productivity and anoxia as well as to elucidate the lakes’ recovery, given the well documented periods of human settlement and abandonment by the Norse from ca. 985–1420CE, and a resettlement starting in ca. 1880CE (Gauthier et al., 2010; Massa et al., 2012b; Millet et al., 2014; Perren et al., 2012). Branched glycerol dialkyl glycerol tetraethers (brGDGTs), a suite of bacterial membrane lipids, extracted from sediments are useful summer mean water temperature (SMWT) proxies in Greenlandic lakes (Colcord et al., 2015; Zhao et al., 2021, 2022). Recent advancements in hyper-spectral imaging (HSI) of lake sediments enable the rapid and cost-efficient reconstruction of continuous high-resolution Holocene-length lacustrine productivity and anoxia histories (Butz et al., 2015, 2017; Schneider et al., 2018; Zander et al., 2022). This method was used to reconstruct the paleo-productivity in a lake in NE Greenland (García-Oteyza et al., 2024), so far it has never been employed to reconstruct anoxia histories in Greenland. Thus, exploring the potential of this method is crucial to contributing to the understanding of the spatiotemporal pattern of Holocene-length anoxia histories in Greenlandic lakes.

In this study we present Holocene-length temperature, eutrophication, and anoxia reconstructions from Lake 578 in the Eastern Settlement of the Norse in South Greenland to address the following research questions: (i) How can non-destructive hyperspectral imaging, complemented by lipid biomarker analyses, advance our understanding of productivity and anoxia histories in Lake 578’s sediments? (ii) How did water temperature, productivity, and anoxia histories of Lake 578 vary throughout the Holocene? (iii) Was there any difference in reconstructed natural variability compared to anthropogenically influenced periods? (iv) How does the Holocene-length temperature reconstruction of Lake 578 compare to other regional reconstructions?

We elucidate the interplay between lake water temperature, productivity, and anoxia during natural and anthropogenically influenced conditions and identify periods of significant change. Finally, we compare the local reconstructions with other records from the region. This comparison aids in contextualizing Lake 578’s climate history within the broader patterns of Arctic environmental change, offering insights into regional climate dynamics during the Holocene.

2. Material and methods

2.1. Study site

Today, Lake 578 (61.08° N, 45.62° W; ~155 m a.s.l.) is dimictic (winter and summer stratification; ice cover: November–May) and its summer water surface temperature may reach up to 18 °C (Zhao et al., 2021). The surface of Lake 578 is about 6.4 ha (excluding an 0.4 ha island), and it has a small seasonal outflow at the southwestern shore. A large fraction of the lake area (~4.2 ha) is shallow (<2 m) and characterized by high growing macrophytes (Myriophyllum alterniflorum). The lake’s maximum depth is ~16 m (Fig. 1). Its watershed is estimated at 89 ha and the water residence time was calculated at ~2 months (Zhao et al., 2021). A dry stream bed indicates an earlier connection to a small lake located at higher elevation, ca. 0.6 km eastward. The climate (1981–2010 CE) as recorded at the nearest weather station (Narsarsuaq airport) is characterized by 7 months of air temperature >0 °C, and one month (July) just above 10 °C, putting it close to the Tundra zone according to the Köppen climate classification (Beck et al., 2018). Monthly precipitation peaks in August and September with >100 mm month⁻¹ and sums up to ~940 mm yr⁻¹. The area’s vegetation is characterized by a patchy mix of shrub-woodland, heath, and tundra, typical of Low Arctic vegetation (Gajewski, 2015), with current patterns influenced by grazing from local sheep herds, evidenced by sheep skulls within the watershed.

Geologically, Lake 578 is embedded within the Eriksfjord formation, composed of Gardar sedimentary rocks and sandstone quartzite from the Mesoproterozoic era, atop the Julianehåb granite, a coarse-grained biotite granite from the Palaeoproterozoic era (Kokfelt et al., 2023). Norse ruins built of granitic blocks characterize the landscape close to steep granitic outcrops in the northeast of the watershed (Fig. 1). Archeological remains of snow collecting dams and evidence of anthropogenic manipulation of the inflow to the East date back to the Norse period (Guldager et al., 2002).

A dirt road to the south of Lake 578 connects the Sillisit sheep farm at the Eriksfjord (established in 1970CE, now home to over 600 mother sheep) and a sheep farm to the northwest (Zhao et al., 2021; Zhao et al., 2021), illustrating the ongoing human influence on the area’s landscape and potentially its aquatic ecosystems.

2.2. Sediment analyses

In August 2019, a 2 m long sediment core was obtained from the depocenter of Lake 578 (61.08° N; 45.61° W, Fig. 1) using a UWITEC gravity corer (9 cm diameter PP-liner) equipped with a percussion hammer. The sediment-water interface was preserved with wet floral foam. The sediment core was split into 1 m pieces in the field (578-1-19-I and 578-1-19-II) and shipped to the University of Massachusetts, where they were stored in dark and cold (4 °C) conditions.

The sediment cores were split lengthwise, sedimentologically described following Schnurrenberger et al. (2003), and subjected to non-destructive measurement techniques (micro-X-ray fluorescence, µXRF; and hyperspectral imaging). The semiquantitative elemental composition was measured on the fresh sediments with an ITRAX µXRF-scanner equipped with a Mo-tube (Croudace et al., 2006). The sampling resolution was set at 0.5 mm with an exposure time of 10 s (voltage: 30 kV, current: 55 mA). Diagnostic elements and elemental
ratios were calculated, and the ratio of incoherent to coherent current (Inc/Coh) served as a proxy for organic content (Davies et al., 2015).

Hyperspectral imaging was conducted at the University of Bern, Switzerland, following the methodologic workflow described in Butz et al. (2015) (Details in SI). The following semiquantitative spectral indices were calculated based on the acquired spectral data:

\[
RABD_{673} = \left( \frac{X \times R_{590} + Y \times R_{730}}{X + Y} \right) \bigg/ R_{\text{local trough minimum}}
\]

where RABD denotes the "relative absorption band depth", \( R_\lambda \) reflectance (R) at wavelength \( \lambda \), and \( X \) (Y) refers to the number of spectral bands between \( R_{590} \) (\( R_{730} \)) and the trough minimum (Schneider et al., 2018). This index is diagnostic for green chloro-pigments (chlorophylls a and b and derivatives), and it is interpreted as a lacustrine productivity proxy (Butz et al., 2015, 2017; Makri et al., 2020; Schneider et al., 2018; Zander et al., 2022).

Fig. 1. Study site and bathymetric map of Lake 578. a) Overview map of the study site, with the red square in the top right panel highlighting the location of the Eastern settlement. Lake 578 is situated northwest of Eriksfjord, with Lake Igaliku to the southeast. The small settlements of Igaliku and Sillisit are marked with black dots. b) Bathymetric map of Lake 578: Grey and dotted areas indicate rock outcrops; the green area to the south represents a marsh and the small houses to the north of the lake represent the Norse ruins. The varying shades of light blue represent different water depths (see legend for details). The coring location is marked with a white asterisk.
\[ R_{\text{BDB}} = \frac{(34 \times R_{\text{BDB}} + 35 \times R_{\text{BDB}})}{69} / R_{\text{BDB}} \]

This index is indicative of changes in bacterio-pheophytins (mainly BPhe a and b) and it is interpreted as an anoxia proxy (Butz et al., 2015; Makri et al., 2020; Zander et al., 2022).

We developed two quantitative linear regression calibrations to convert RABD$_{BDB}$ and RABD$_{SI}$ values into absolute green pigment (GP) and bacterio pheophytin (BPhe) concentrations respectively, based on previously published in-situ calibrations from the very same system (Makri et al., 2020; Schneider et al., 2018; Zander et al., 2022). Outliers and samples with large leverage (Cook’s distance >0.5) were identified based on residual analysis and excluded from the final calibrations. Both calibrations (with and without outliers) were plotted to visualize the differences. Root mean squared errors of prediction (RMSEP) were estimated by 10-fold, k-fold, and bootstrap approaches following Butz et al. (2015).

We divided the high-resolution datasets (hyperspectral imaging, and \( \mu \text{XRF} \)) congruently to each physical sub sample to enable direct comparison.

### 2.3. Lipid biomarker analysis

The sediment cores were sub sampled at ~1 cm resolution (209 samples), freeze-dried, and dry bulk density (DBD, g cm$^{-3}$) determined. To prevent contamination from the core liner and plastic film, the sample edges were trimmed.

Aliquots (0.7–2 g) of 115 lyophilized and homogenized sediment samples were used for organic geochemical analysis. The aliquots were mixed with combusted diatomaceous earth and total lipids (TLE) were extracted with a mixture of dichloromethane/methanol (DCM/MEOH, 9:1, \( \nu \nu \)) using an automated accelerated solvent extractor (ASE 200, Dionex). The solvents were evaporated with a TurboVap (Zymark, Natick) on a heating plate of 45 °C. The extracts were re-dissolved in hexane/isopropanol (99:1, \( \nu \nu \)) using an automated accelerated solvent extractor (ASE 200, Dionex). The solvents were evaporated with a TurboVap (Zymark, Natick) on a heating plate of 45 °C. The extracts were re-dissolved in hexane/isopropanol (99:1, \( \nu \nu \)) and filtered through PTFE filters (mesh size: 0.45 µm). A known amount of the C$_{60}$ GTG standard was added to each sample prior to analysis. The brGDGTs were measured using a High-Performance Liquid Chromatography system (HPLC, Agilent 1260) coupled to a Quadrupole Mass Selective Detector (MSD, Agilent 6120) following the UHPLC method established by Hopmans et al. (2016). Two coupled Waters UHPLC columns (150 mm × 2.1 mm x 1.7 µm) were used to separate the compounds. Chromatography was performed applying a three-phase isocratic solvent gradient using 100% hexane (solvent A) and hexane/isopropanol (9:1, \( \nu \nu \), solvent B): i) 18% B, 25 min; ii) ramp to 35% B, 25 min; iii) ramp to 100% B; 30 min and system rinse.

Different GDGT indices were calculated based on the resulting fractional abundances and we applied several global and local calibrations (with and without outliers) to convert the brGDGTs in temperature (Table 1).

### 2.4. Chronostratigraphy

The age depth model of core 578-01-19 (Fig. 2) was generated using a combination of short-lived radionuclide dating (\(^{210}\text{Pb}, ^{137}\text{Cs}\)) and radiocarbon dating (\(^{14}\text{C}\)). A \(^{210}\text{Pb}\) CRS age-depth model (verified with \(^{137}\text{Cs}\) as independent chrono-marker) was derived from a previous study working on cores from 2016 and used to constrain the top 6 cm of the 2019 core (Zhao et al. 2021, 2022). Alongside the \(^{210}\text{Pb}\) age-model, one \(^{14}\text{C}\)-date (UCLAMS-193285, Table 2) was included from the previous study based on a core-to-core correlation using visual inspection, \( \mu \text{XRF} \) data, and HSI data (Fig. S1). Macrofossils were either picked from the fresh sediment core, or the lyophilized samples, and measured at the Keck-Carbon Cycle AMS facility (University of California Irvine, CA).

![Table 1](image-url)

**Table 1** Overview of the applied GDGT-proxies. “Index” refers to the name as it is referred to in the main text.

<table>
<thead>
<tr>
<th>Index</th>
<th>Interpretation/Description</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>TLE</td>
<td>Total lipid extract (concentration and fluxes): productivity</td>
<td>Zhao et al. (2021)</td>
</tr>
<tr>
<td>SMW</td>
<td>Summer mean water temperature (JJJ), in-situ calibration of Lake 578</td>
<td>Zhao et al. (2021)</td>
</tr>
<tr>
<td>MAF.8B</td>
<td>Temperature for months above freezing based on the “full” fractional abundance set</td>
<td>Raberg et al. (2023)</td>
</tr>
<tr>
<td>MAF.Zh</td>
<td>Months above freezing based on mid- to high latitude sites, based on a new MBT$_{SIM}$ calibration</td>
<td>Zhao et al. (2023)</td>
</tr>
<tr>
<td>MLR.Zh</td>
<td>Months above freezing based on mid- to high latitude sites, based on a new multivariate linear regression (MLR) calibration</td>
<td>Zhao et al. (2023)</td>
</tr>
<tr>
<td>GDGT-0/ Cren</td>
<td>The ratio between Calderarchaeol (GDGT-0) and Cremanarchaeol (Cren); Dominance of methanogens, interpreted as anaerobic</td>
<td>Schouten et al. (2013), Blaga et al. (2009)</td>
</tr>
</tbody>
</table>

The samples were calibrated using the IntCal20 calibration curve (Reimer et al., 2020) and modeled using a Bayesian age-depth model (rbaco, Blauw and Christen, 2011).

Furthermore, we identified the palynological onset of introduced anthropochorous species such as *Rumex acetosa*-type pollen or *Ranunculus acris*-type pollen (Massa et al., 2012b; Schofield et al., 2013), and coprophilous fungal spores (Gauthier et al., 2010) to constrain the age-depth model with the historically documented *landnam* of the Norse (985CE; observed at 49.5–50.5 cm in core 578-01-19). At 52 cm depth, a comparison between a macrofossil and a plant-derived micro-charcoal inferred age revealed an offset of approximately 400 years (Table 2). If the macrofossil is of terrestrial origin, the offset likely arises from different transport and deposition pathways of the two samples. While the degraded macrofossil could originate from reworked terrestrial material (Gauthier et al., 2010; Strunk et al., 2020) - in the case of Lake 578 induced by the land use of the Norse - the micro-charcoal might have entered the lake directly via atmospheric deposition, eliminating any time lag from source to sink. Conversely, if of aquatic origin, a hard water effect could explain the discrepancy. To better understand impacts on the chronology and to make an informed age-depth model selection, we compared a range of different age models with different assumptions and corrections (Fig. S2). We implemented an old-age-correction of 400 years to the samples suspected to be from aquatic origin using the delta. R function in rbaco (Blauw and Christen, 2011) and selected the model as presented in Fig. 2 (corresponding to the black line in Fig. S2). The age-corrected version places the *landnam* (pollen layer) close to 985CE (the documented arrival of the Norse), in contrast to the uncorrected age-model (ca. 650CE; Fig. S2).

Lastly, we excluded a distinct clastic layer from 181 to 190 cm to check the robustness of the age-depth model. The layer-excised model only varies below the second lowest \(^{14}\text{C}\) date and stays within the 95% confidence interval of the original model, suggesting that the model-selection does not alter the interpretation of the sequence after the early Holocene.

Mass accumulation rates (MAR, g cm$^{-2}$ yr$^{-1}$) were calculated per sample by combining the DBD with the sediment accumulation rate (SAR, cm yr$^{-1}$) and used to calculate substance fluxes (Håkanson and Jansson, 2002).

### 2.5. Data analysis and identification of periods of significant change

Statistical analyses were conducted in R (R Core Team, 2019) using a variety of software packages (Table S2). The data was standardized prior to the application of cluster and ordination analyses. Significant clusters in the stratigraphy were identified based on a constrained incremental sum of squares cluster analysis (CONISS; Grimm, 1987). Principal Component Analysis (PCA) was performed on Si, K, Ca, Ti, Fe, Sr, and...
the ratio of incoherent to coherent current (Inc/Coh), with the first principal component (PC1) interpreted as the lithogenic fraction.

We applied a generalized additive model (GAM) workflow (Simpson, 2018) to identify periods of significant change (e.g., natural vs. anthropogenic) in the irregularly spaced productivity (GP-flux), anoxia (BPhe-flux), and MAR time series (parameters in Table S3). We conducted a restricted maximum likelihood approach for the model-smoothness selection (Simpson, 2018). Simultaneous confidence intervals (95%) were calculated based on a posterior process (30 simulations, Simpson, 2018). The first derivative and its 95% simultaneous confidence were calculated, and periods of significant change were defined as periods where the confidence interval excludes the value 0 (Simpson, 2018).

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Table 2
Summary of the $^{14}$C data showing the Lab ID, sample ID, sediment depth, sample type and mass, the $^{14}$C ages, the calibrated ages (IntCal20 0.95 confidence range), and the resulting modeled age (“median” year from the rbacon model at the sample depth).Italic fonts indicate a sample that was projected from a previous study (Zhao et al., 2022) (light green in Fig. 2). Asterisks (*) in the column “Lab-ID” mark samples that were included in the final chronologies. $^*$-symbols indicate samples that were reservoir age corrected. The different age models are represented in Fig. S2. The data and model output can be found at the NSF Arctic Data Center.

<table>
<thead>
<tr>
<th>Lab-ID</th>
<th>Sample ID</th>
<th>Depth [cm]</th>
<th>Material</th>
<th>C [mg]</th>
<th>$^{14}$C yr BP</th>
<th>Error [± 1σ]</th>
<th>Yr cal BP</th>
<th>Modeled Yr cal BP</th>
</tr>
</thead>
<tbody>
<tr>
<td>UCIAMS-282432</td>
<td>Gr-C14-233</td>
<td>32.35</td>
<td>Plant tissue</td>
<td>0.021</td>
<td>2990</td>
<td>80</td>
<td>Outlier</td>
<td>Outlier</td>
</tr>
<tr>
<td>UCIAMS-245138</td>
<td>Gr-C14-243</td>
<td>44.6</td>
<td>Plant tissue, pollen</td>
<td>0.021</td>
<td>1240</td>
<td>110</td>
<td>1344-931</td>
<td>832</td>
</tr>
<tr>
<td>UCIAMS-282434</td>
<td>Gr-C14-241</td>
<td>52</td>
<td>Charred plant tissue</td>
<td>0.11</td>
<td>1310</td>
<td>20</td>
<td>1068-959</td>
<td>999</td>
</tr>
<tr>
<td>UCIAMS-193285</td>
<td>Gr-C14-240</td>
<td>81</td>
<td>Woody plant tissue</td>
<td>0.11</td>
<td>2835</td>
<td>20</td>
<td>3001-2869</td>
<td>2913</td>
</tr>
<tr>
<td>UCIAMS-245139</td>
<td>TS-109</td>
<td>110.55</td>
<td>Plant tissue</td>
<td>N/A</td>
<td>4400</td>
<td>20</td>
<td>5043-4874</td>
<td>4929</td>
</tr>
<tr>
<td>UCIAMS-245140</td>
<td>TS-132</td>
<td>185.85</td>
<td>Small plant tissue</td>
<td>0.042</td>
<td>2270</td>
<td>60</td>
<td>Outlier</td>
<td>Outlier</td>
</tr>
<tr>
<td>UCIAMS-229939</td>
<td>Gr-C14-209</td>
<td>197.6</td>
<td>Pieces of plants</td>
<td>0.17</td>
<td>9580</td>
<td>30</td>
<td>11106-10747</td>
<td>10854</td>
</tr>
</tbody>
</table>

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Fig. 2. Sedimentary classification and chronology plotted versus depth for the sediment core from Lake 578. The green (GP) and purple (BPhe) photos illustrate the RABD-distribution maps, derived from Hyperspectral Imaging (HSI). The image in the middle is a HSI-derived true color image of the sediment core. The pictograph next to the BPhe photo depicts the sediment description (please refer to the legend below the age model for details on sediment characteristics), and the numbers with dashed horizontal lines indicate the four distinct facies. The large panel shows the selected age depth model of Lake 578 (See also Fig. S2). Terrigenous elements (Sr, Ti, K), the ratio of incoherent to coherent currents (INC/COH; indicative of organic content), water content (dashed line), dry bulk density (DBD; grey shaded plot), sediment accumulation rate (SAR, dashed line), and mass accumulation rate (MAR; grey shaded plot) are shown for the 2 m sediment sequence cored in August 2019.
3. Results

3.1. Sediment characteristics

The bottom of sediment core 578-01-19 is estimated to date to ~11 cal ka BP (+357, −249). This results in a SAR ranging from 0.105 to 0.67 mm a\(^{-1}\) and a MAR between 1.4 and 21.4 mg cm\(^{-2}\) yr\(^{-1}\) (Fig. 2). The sediment description revealed that the core is characterized by four distinct lithofacies (Fig. 3). Facies 1 (200–172 cm) exhibits alternating pinkish silty clay layers enriched in lithogenic elements (Sr, Ti, and K) and organic-rich darker layers with elevated RABD\(_{673}\) and Inc/Coh values and reduced DBD. A distinct clastic layer (190–181 cm) indicates a gradual increase in Sr, Ti, and K, suggesting a sorted grain size succession. It is capped with a thin layer concentrated in Sr, Ti, and K. Facies 1 exhibits the highest MAR values, while showing a relatively stable SAR (Fig. 2). Notably, this lithofacies is characterized by exhibiting the relatively highest RABD\(_{845}\) values. Facies 2 (172–158 cm) reveals a shift from clay- to organic-rich layers, as denoted by lower levels of Sr, Ti, and K, relatively lower DBD, and marginally elevated RABD\(_{673}\) values compared to lithofacies 1. The onset of facies 3 (158–50 cm) marks a continuing decrease in DBD, Sr, Ti, K, and RABD\(_{845}\) values alongside rising RABD\(_{673}\) values, highlighting a transition to predominant organic autochthonous sedimentation. Between ca. 130 cm and 100 cm depth, a slight increase in Sr and Ti counts, along with an

![Fig. 3. Multiproxy sediment stratigraphy and ontogeny.](image-url)

(a) PC-1 of PCA conducted on minerogenic µXRF elements. SAR (dashed line), MAR (brown shade), GP concentration (dashed line, bottom), GP flux (green shade, top), TLE concentration (dashed line, bottom), TLE flux (green shade, top), BPhe concentration (dashed line, bottom), BPhe flux (purple shade, top). "SMWT": Summer mean water temperature calibration by Zhao et al. (2021). "MAF.Rb": Months above freezing temperature calibration by Raberg et al. (2021). Roman numerals and dashed horizontal lines represent the five clusters identified by CONISS analysis, with horizontal brown boxes indicating periods of documented regional human land use (Norse and modern sheep farming).

(b) Ontogeny of Lake 578 per cluster: I) Inverse temperature density stratification, possibly due to strong seasonality and extended ice cover in the early Holocene leading to methanogen (purple) and bacterial (yellow) blooms, depositing pigments and lipids into sediments. II) Reduced ice cover promotes polymeric events that lead to decreased anoxic conditions. III + IV) Holomictic events, driven by wind shear, redistribute oxygen and nutrient-rich bottom waters, triggering algal blooms (green) and disappearance of BPhe (reduced anoxia). V) Lake level rise and human land use increase nutrient input, boosting productivity and enhance anoxic conditions, especially during the Norse period.
increase in DBD, suggest a period of enhanced clastic material deposition. Facies 4 represents the uppermost core section (50–0 cm). Fine laminations fade out towards homogenous and organic-rich sediments. The start of this section is marked by an abrupt increase in water content, and RABD$_{O73}$ and RABD$_{A45}$ values. The lowest DBD and the highest RABD$_{O73}$ values are contained in the top 2 cm. Notably, Facies 4 is characterized by the highest SAR, while the MAR is relatively low compared to those observed in Facies 1.

### 3.2. Hyperspectral imaging calibration and pigment fluxes

The RABD distribution maps (Fig. 2) represent the spectral absorption trough depths at 673 nm (RABD$_{O73}$ values between 1 and 1.633), and 845 nm (RABD$_{A45}$ values between 0.98 and 1.18) per pixel. The discrete sub samples show RABD$_{O73}$ values ranging from 1.019 to 1.55 and RABD$_{A45}$ values from 0.9835 to 1.046. The residual analyses of the RABD calibrations (Figs. S6 and S7) highlighted outliers (GP: 7; BPhe: 5) which were excluded for the calibration. This step reduced influence of extreme values (samples with large impact on the regression) and made the models more robust, despite reducing the adjusted coefficient of determination (R$_{adj}^2$, BPhe: 0.86, GP: 0.7, p < 0.001). The root mean squared errors of prediction (RMSEP) were ~12.1% for the GP and ~8% for the BPhe. The applied RABD calibration with GP concentrations spans from 2.08 to 511.21 μg g$^{-1}$ and RABD$_{O73}$ values from 1.022 to 1.475. The BPhe calibration covers RABD$_{A45}$ values from 0.9875 to 1.045 and concentrations from 0.2 to 48.35 μg g$^{-1}$. Values beyond these ranges are linearly extrapolated. These are the first calibrations based on multiple cores from different study sites and they were used primarily to calculate pigment fluxes and investigate the relative temporal changes of pigment fluxes.

The resulting HSI-derived GP concentrations mostly follow the GP fluxes (Fig. 3). They increase after ca. 8.0 cal ka BP along with the fade-out of the clastic layers (Facies 2) and stay at relatively high levels. They gradually decrease after ca. 4.0 cal ka BP towards a period characterized by the lowest values from ca. 2.5–1.1 cal ka BP. After ca. 980CE, the GP concentrations and fluxes rapidly increase and exhibit a 400-year period of unprecedented values, before they abruptly decrease. After ca. 1900CE, the GP values show a strong increasing trend that results in peak values after 1970CE. Bacteriophaephytin (BPhe) concentrations and fluxes both show highest values during the early Holocene period (Facies 1) with peak values at ca. 9.2 cal ka BP (Fig. 3). Notably, the peak values coincide with the organic-rich layers (Fig. 2). They start fading out thereafter, along the disappearing laminations observed in facies 2, and completely disappear between 5.8 cal ka BP and ca. 3.7 cal ka BP. Thereafter, both BPhe fluxes and concentrations slightly increase, before disappearing around 1.2 cal ka BP. With a small lag compared to the GP values, BPhe starts to strongly increase after ca. 980CE and disappears after ca. 1450CE.

### 3.3. GDGT proxy reconstructions

Branched GDGTs have previously been applied to Lake 578 to examine the temperature history of the past ~1800 years (Zhao et al., 2022). Here, we extend the record throughout the Holocene based on new sediment cores and additionally measured GDGT-0 (Caldarchaeol) and Crenarchaeol (Cren). We particularly focus on the in-situ summer mean water temperature of Lake 578 (SMWT; Zhao et al., 2021, Figs. 3 and S9, S10), as the brGDGT provenance is predominantly lacustrine and distinct from the watershed soils (Figs. S3–S5, and S1). The comparison of the SMWT reconstruction to three other recently published calibrations reveals that the SMWT reconstruction yields substantially higher absolute temperatures and covers a wider temperature range (Figs. S9 and S10). While the reconstruction based on the MAF calibration by Raberg et al. (2021) follows the general short term trends of the SMWT (Fig. 3, S9, S10), the two other reconstructions (MAF.Zh, and ML.Zh, Figs. S9 and S10; Zhao et al., 2023) show much less variation. Except for one calibration (ML.Zh, Figs. S9 and S10) the reconstructions indicate an initial warming trend between 11 cal ka BP and 8 cal ka BP, characterized by several significant cooling events. All reconstructions indicate a rapid temperature drop between 9.5 and 9.3 cal ka BP. Subsequently, the temperature strongly increases, until ~9.2 cal ka BP thereafter sharply drops again and increases after 8.9 cal ka BP. Consistently, the reconstructions tend to plateau between 8.8 cal ka BP and 8.5 cal ka BP, decline sharply after 8.2 cal ka BP and thereafter exhibit the highest values between 7.95 and 7.65 cal ka BP. A successive cooling trend toward relatively colder temperatures at 7.2 cal ka BP characterizes the subsequent 400 years. The remainder of the middle Holocene period (high resolution only until 4.8 cal ka BP) is at relatively higher temperatures with smaller amplitude variability. The sampling resolution between 4.8 cal ka BP and 1.7 cal ka BP is insufficient to support definitive statements about temperature variability during this period. Over the past 1.7 ka, each calibration indicates a decreasing temperature trend (Figs. S9 and S10).

We used the ratio between GDGT-0 and Crenarchaeol as a proxy to reconstruct the anoxia history of Lake 578 (Fig. 3; Blaga et al., 2009; Cao et al., 2020; Schouten et al., 2013). The trend in the GDGT-0/Cren ratio follows closely the trend observed in BPhe concentrations and fluxes: The early Holocene is characterized by the relatively highest values and subsequently both the GDGT-0/Cren ratio and the BPhe proxies drop to the lowest values after ca. 6.5 cal ka BP. The period between 980 and 1450CE is characterized by increased anoxia, which subsequently disappears throughout the rest of the sequence (for further details refer to the SI).

### 3.4. Multiproxy stratigraphy and periods of significant change

To objectively identify significant changes in the multiproxy dataset and the single time series, we applied a CONISS cluster analysis (Grimm, 1987) and a general additive model (GAMs) workflow (Simpson, 2018). The CONISS cluster analysis applied to the multiproxy dataset revealed five distinct clusters (Fig. 3).

Cluster I (Early Holocene, 11–9.6 cal ka BP) is characterized by low temperatures (SMWT and MAF), low productivity (GP), and relatively high anoxia as reconstructed independently by hyperspectral imaging (BPhe), GDGTs (GDGT-0/Cren), and µXRF (S). Notably, short-term drops in temperature are closely followed by elevated anoxia values.

Cluster II (9.6–7.9 cal ka BP) is marked by an overall temperature increase along with decreasing anoxia proxies, a reduction in mineralogic sediments and an increasing trend in productivity. The start of Cluster II exhibits relatively low temperature and high anoxia values. Towards the end of Cluster II, anoxia proxies decrease to minimum levels starting around 8.3 cal ka BP, along with increased water temperatures (SMWT). The transition to Cluster III features rising temperatures, reduced anoxia, and a shift towards more organic-rich sediments, indicated by lower lithogenic counts (PC1 scores) and higher productivity and Inc/Coh values. This shift is also visually evident in the sediments, where fine laminations transition into almost homogeneous organic sediments (Fig. 2).

Cluster III (7.9–5.9 cal ka BP) is characterized by the relatively highest temperatures, increased productivity, and fading anoxia. This goes along with a reduction in lithogenic content (lower PC1 scores, and DBD) and an increasing Inc/Coh ratio (Fig. 2).

Cluster IV (5.9–1.0 cal ka BP) exhibits relatively stable productivity and anoxia values, with relatively high temperatures. Interestingly, the minerogenic contents (PC1 scores) increases at the beginning of the cluster and decreases after ca. 4 cal ka BP. The boundary between cluster IV and V is characterized by a rapid increase in productivity and anoxia proxies, while the temperature slightly decreased. Note that the sampling resolution of the lipid biomarker samples is relatively low in this section.

Cluster V (1.0 cal ka BP – 2019CE) exhibits the highest productivity values and organic content (Inc/Coh) with relatively higher anoxia
proxies at the beginning compared to Cluster IV. Both, the SMWT and MAF temperatures show a weak decreasing trend throughout most of the cluster, except for the topmost rapid drop (Fig. 3).

The generalized additive models (GAMs) fitted to the MAR, GP (productivity), and BPhe (anoxia) time series and their first derivatives revealed several periods of significant change (Fig. 4, S8). While the MAR exhibited two periods of significant change in the Early Holocene, the high resolution BPhe flux time series exhibits eight periods of significant change in the Early Holocene and one coinciding with the Norse period (Fig. 4). The GAM fitted to the GP flux time series revealed two periods of significant change, one coinciding with the Norse period and the other one coinciding with the modern sheep farming period, concurrent to findings from Lake Igaliku (Bichet et al., 2013).

4. Discussion

Lakes in southern Greenland, including Lake 578, have been the focus of previous paleoclimate research, using a range of proxies to study past climate conditions, deglaciation processes, environmental changes, and human impacts, such as Norse settlements (e.g., Axford et al., 2021; Briner et al., 2016; Gajewski, 2015; Larocca and Axford, 2022; Sparrenbom et al., 2013, 2006a, 2006b; Sundqvist et al., 2014; Zhao et al., 2022). Despite this breadth of research, the variability of anoxia, productivity, and temperature – key factors in understanding lake responses to climatic and anthropogenic influences – remain understudied for this region. The following discussion centers on this gap with comprehensive Holocene-length reconstructions from Lake 578.

4.1. Lake formation

During the early Holocene period, prior to ca. 9.6 cal ka BP, sedimentation in Lake 578 was predominantly minerogenic with interspersed small organic-rich layers (Fig. 2). This may suggest that akin to Lake Igaliku (Massa et al., 2012b), other lakes in the Qaqortoq area (Sparrenbom et al., 2006a, 2013) and other isolation lakes further NW (Larocca et al., 2020b), Lake 578 might have been isolated from the sea. This could have been due to the isostatic rebound of Greenland, a process that would result in the deposition of unsorted minerogenic sediments. Alternatively, the lake could have been influenced by minerogenic glacial meltwater inputs and prolonged ice cover, leading

Fig. 4. Generalized additive models (GAM) and identification of periods of significant change. The figure represents the fitted time series of green pigment flux (GP-Flux), bacterio pheophytin flux (BPhe-Flux), and mass accumulation rate (MAR) using different model parameters (details in Table S3 and Fig. S8). The colored ribbons represent the 97.5% confidence interval. The first mathematical derivatives are shown in Fig. S8. The timeline at the bottom depicts the periods of significant change per time series, where the simultaneous confidence interval does not include the value 0: top (green): GP, middle (purple): BPhe, and bottom (red): MAR. The grey vertical shade indicates the Norse period.

Fig. 5. Temperature reconstruction of Lake 578 in regional context. The top panel represents the reconstruction of Lake 578 (this study), numbers 1 to 4 indicate different rapid freshwater discharges on the Northern Hemisphere: 1. Nesje et al. (2004); 2. Jennings et al. (2015); 3. Alley and Agústsdóttir (2005); 4. Fleitmann et al. (2008). Kobashi et al. represents the ice core reconstruction presented in Kobashi et al. (2017); Marcott et al. displays the multiproxy reconstruction (30–90° N) presented in Marcott et al. (2013), Kaufmann et al. shows the multiproxy reconstruction from Kaufman et al. (2020) (60–90° N), Hancock et al. illustrates the multiproxy reconstruction (GIC) as presented in Hancock et al. (2023), Gajewski depicts the multiproxy reconstruction for S Greenland presented in Gajewski (2015).
to low productivity, comparable to the conditions observed in lakes slightly further south, such as Lakes Quvnerit, Alakariqssoq, or Uunartoq (Larocca et al., 2020a). A recent regional deglaciation survey conducted by Levy et al. (2020) suggests that the area around Lake 578 was deglaciated between 12.3 and 11.1 cal ka BP, corresponding to the age of the oldest sediments in the present record. Concurrently, the relative sea level during this period is estimated to have been approximately 120m above present sea level (Sparrenbom et al., 2013). Considering Lake 578’s altitude at 155 m a.s.l. and the presence of sorted clay facies in its sediments, as evidenced by the trends in high-resolution Sr, Ti, and K counts (Fig. 2), the hypothesis of (de)glacial meltwater influence appears more substantiated. Therefore, we interpret Lake 578 as a post-glacial lake with sediments that provide a continuous terrestrial archive of climate and environmental changes throughout the Holocene.

4.2. Early Holocene: a phase characterised by low temperatures and anoxia

The reconstruction for Lake 578 indicates an Early Holocene temperature rise prior to 8 cal ka BP (Fig. 3). Notably, this Early Holocene warming is delayed compared to hemispheric and regional models and multiple post-glacial temperature reconstructions, which are mostly dominated by ice cores (Fig. 5). This is not surprising, as it is in accordance with findings of other lake sedimentary reconstructions in the region, regardless of the investigated temperature proxies (Lake Igaliiku: Massa et al., 2012b; Lake Qipisarqo: Kaplan et al., 2002; Lake N14: Andresen et al., 2004; Lakes Quvnerit and Alakariqssoq: Larocca et al., 2020a). Similarly, in their comprehensive analysis of a North Atlantic Fennoscandian stack of summer temperatures (composed of 63 reconstructions from both marine and terrestrial archives) Sejrup et al. (2016) find negative anomalies between 10 and 8.5 cal ka BP, followed by a warming trend until around 6.5 cal ka BP. They suggest that the large-scale North Atlantic regional climate was decoupled from global and northern hemispheric temperature trends during the early Holocene, attributing this to the extended cooling effect of the melting Laurentide Ice Sheet. Their spatial analysis further reveals a strong negative temperature anomaly over the Labrador Sea, extending well over the southwest coast of Greenland around 9 cal ka BP. Thus, the cold anomalies observed in Lake 578 and other lakes in South Greenland may be explained, as presented in Sejrup et al. (2016), by the persistent impact of the collapsing Laurentide Ice Sheet.

The pronounced rapid cooling events observed in the SMWT reconstruction prior to 8.5 cal ka BP largely coincide with significant freshwater discharges into the North Atlantic, leading to regional cooling events as registered in several Arctic marine and ice core reconstructions (Fig. 5, Alley and Agüestdottir, 2005; Fleitmann et al., 2008; Jennings et al., 2015; Nesje et al., 2004). These cooling events, while prominently recorded in the SMWT, are reflected at a much smaller amplitude in the MAF reconstructions (Fig. 3, S9, S10), indicating a more substantial impact on the summer temperature (JJA) compared to the broader MAF season (AMJJASO).

Concurrently with some of these cooling events, rapid increases in anoxia values suggest enhanced oxygen depletion in the bottom waters (Fig. 3), which could be explained by several hypotheses. One hypothesis is that the anoxia in the bottom waters is potentially indicative of prolonged invertebrate stratification caused by extensive ice cover, as observed in contemporary Arctic lakes (Kirillin et al., 2012; Klanten et al., 2023; Klaus et al., 2021; Schwefel et al., 2023; Yang et al., 2021). A second hypothesis concerns extreme seasonality in the early Holocene due to high summer and low winter insolation: this could have led to strong summer productivity which produced organic matter, fueling respiration that then caused anoxia during colder-than-modern winters. Thirdly, changes in the wind system could have reduced the wind fetch and thus suppressed wind-induced turnover events and consequently oxygen redistribution. Finally, it cannot be excluded that the brGDGT thermometer may be affected by anoxic conditions, leading to a cold bias, as observed in the Eifel maar lakes stack record (Zander et al., 2024), or in other Greenland lakes (McFarlin et al., 2023). However, as the temperature trends observed in Lake 578 follow those of other regional reconstructions relying on different proxies (Andresen et al., 2004; Kaplan et al., 2002; Larocca et al., 2020a; Massa et al., 2012b; Sejrup et al., 2016), we find this interpretation to be improbable.

While all these hypotheses are speculative, we perceive the interpretation of extended ice cover combined with strong seasonality to be the most substantiated. We observe anoxia spikes shortly after the onset of organic rich layers (Fig. 3), implying that respiratory biologic activity may have depleted the oxygen in the possibly prolonged ice-covered lake (Leppi et al., 2016; Yang et al., 2021). Once the oxygen in the water column was consumed, bacteria and methanogens (e.g., purple sulfur bacteria) which thrive in anoxic conditions may have proliferated, resulting in the production of bacterio pheophytins and an increased GDGT-0/Gren ratio (Blaga et al., 2009).

A subsequent cold event at ca. 7.5 cal ka BP, aligns with regional cooling observed in annual temperature reconstructions, indicating a broader regional cooling effect extending from the ice sheet to the sea (Gajewski, 2015; Hancock et al., 2023; Kobashi et al., 2017). However, the mechanism cannot be attributed to larger freshwater pulses to the Labrador Sea or the North Atlantic (Fig. 5).

4.3. Middle to late Holocene: a stable postglacial lake system

After ~8.7 cal ka BP, the fine sediment laminations fade out while anoxia values drop and SMWT and productivity increase, suggesting seasonally ice-free periods that caused holomictic events within the lake, enhancing the oxygenation of the hypolimnetic waters and nutrient redistribution to the productive layer (Kirillin et al., 2012; Klanten et al., 2023; Schwefel et al., 2023). We interpret the highest observed SMWT as the expression of the Holocene Thermal Maximum caused by a regional atmospheric warming. Similar transformations at around the same time were also observed in three lakes further south from Lake 578 (Larocca et al., 2020a, 2020b) and in Lake Igaliiku (Massa et al., 2012b), as well as in the North Atlantic-Fennoscandian stack including Iceland (Sejrup et al., 2016). The period of persistently low anoxia, relatively high temperature and continuous but relatively low productivity may indicate a dimictic and oligotrophic lake system, which aligns with the findings in Lake Igaliiku (Massa et al., 2012b). Interestingly, the partially model-derived regional summer (JJA) reconstruction by Buizert et al. (2018) suggests a relatively brief Holocene Thermal Maximum and an earlier onset of the Neoglacial period, marked by lower temperature anomalies compared to those reconstructed for Lake 578. We attribute these differences to several factors: the altitude of the ice cores used in their study (400 m a.s.l) versus Lake 578 (155 m a.s.l), the broader area integrated in their model (spanning from the southern tip to Narssarsup), and the potential decoupling between atmospheric temperatures (reflected in ice cores) and lake water temperatures (Lake 578). However, our findings align with other regional lake reconstructions at lower elevations and with the North Atlantic Fennoscandian stack, suggesting that variations in climate archives -as highlighted in Axford et al. (2021)- or oceanic processes may explain these discrepancies.

The slow increase in anoxia and the drop in productivity after ca. 4.1 cal ka BP initiate a transition towards prolonged winter density stratification, possibly caused by extended ice-cover. Concurrently, Lake Igaliiku faced a period of reduced summer stratification after ~4.8 cal ka BP, reflected in a shift from benthic to planktonic diatoms, which is interpreted as summer cooling (Massa et al., 2012b). The temperature reconstruction of Lake 578, however, does not exhibit strong cooling, but this may also be an artifact of the very low temporal resolution of the record at around this time. These local observations of cooler summers and prolonged ice cover durations may be the response to Neoglacial cooling (Briner et al., 2016).
4.4. Late Holocene: anthropogenic impacts on the lake

A rapid increase in productivity, alongside increased anoxia between 970 and 1000CE, coincides with the arrival of the Norse in the watershed, highlighting the onset of their impact on the lake system. Local land use and sheep herding possibly increased nutrient inputs, enhancing eutrophication and associated anoxia, similar to observations in mid-latitude alpine lakes (Jenny et al., 2014). Zhao et al. (2022) previously hypothesized that a significant productivity increase in Lake 578 around 650CE might have resulted from a water level rise, submerging nutrient-rich watershed soil, thereby explaining the extensive shallow area (<2 m, Fig. 1). Indeed, their relative humidity reconstruction revealed high values during this period, indicative of intensified precipitation (Zhao et al., 2022). Nonetheless, our comparison between charcoal and microfossil samples suggests that some of Zhao et al.’s dated macrofossils might either originate from reworked soil material or be of aquatic origin and affected by a hard water effect (Table 2). Recalculating the previously published age model with the determined 400-year correction places this lake level increase post the Norse arrival. This suggests that the observed increase in productivity and anoxia in our study likely results from a combination of a lake level increase that submerged nutrient-rich soils, and anthropogenically caused nutrient inputs. This mirrors the Norse impact on Lake Igaliku, (Gauthier et al., 2010; Massa et al., 2012a; Millet et al., 2014), albeit perhaps less intensely than previously reported (Perren et al., 2012). Given Norse practices of constructing dams and drainage systems, as evidenced in Garðar/Igaliku (Edwards and Schofield, 2013), and many other places in Greenland (Arneborg, 2005), it is plausible that lake levels were artificially elevated by dams in strategic locations in the southwest of Lake 578. However, this remains hypothetical, necessitating further field and archeologic investigation to validate.

After the abandonment of the Eastern Settlement in the early 15th century (Zhao et al., 2022) both the productivity and anoxia rapidly decreased, similar to the organic carbon accumulation decline in Lake Igaliku (Millet et al., 2014). The productivity, however, stayed above pre-Norse values, highlighting the sensitivity of the aquatic ecosystem beyond direct anthropogenic activities. Similar processes have been observed in lakes on Baffin Island, where Arctic paleo-people have altered some lake systems ~4.5 cal ka BP and the effect of this anthropogenic legacy is still visible today (Michelutti et al., 2013).

Interestingly, during the Little Ice Age (LIA, in the Northern Hemisphere from about 1250 to 1860CE, Wanner et al., 2022) anoxia was at low levels, suggesting either that the anthropogenic perturbation altered the system causing the disappearance of BPhe producers and anoxia-sensitive bacteria, or that the lake was not perennially ice-covered. Given the relatively stable SMWT in Lake 578 during that time, the latter seems plausible. This is not surprising as findings from the European mid-latitudes show that the summer temperature responded unproportionally to the LIA cooling compared to annual temperature (Wanner et al., 2022). Concurrently, the regional vegetation composition and spatial expand stayed rather stable (Millet et al., 2014), indicating that large-scale atmospheric patterns such as changes in the North Atlantic Oscillation (NAO) may have reduced the frequency of low temperature summers during the LIA (Wanner et al., 2022; Zhao et al., 2022).

The area around Igaliku was resettled in the middle of the 18th century (Arneborg, 2023; Massa et al., 2012a) and modern farming began after 1920CE (Gauthier et al., 2010; Massa et al., 2012a). In Lake 578, productivity started to increase after ca. 1880CE, which may be again related to regional anthropogenic land use. A sharp increase in productivity after ca. 1970CE may be related to the modern sheep farming that took place in Lake 578’s region (e.g., Sillisit). The absence of anoxia proxies in the modern period of land use suggests that mixing events in Lake 578 may compensate for the higher trophic conditions, which is confirmed with in-situ measurements (Zhao et al., 2021).

Observed trends in temperate and alpine lakes globally reveal that warm temperature stratification patterns become more prevalent, promoting anoxia in some cases (Jane et al., 2023; Jenny et al., 2014, 2016). Recent studies further indicate that this development is starting to occur in large Arctic lakes as well (Klanten et al., 2023). Although the multiproxy reconstruction of Lake 578 did not indicate anoxic conditions during the Holocene Thermal Maximum, it cannot be ruled out that the lake level increase during the Norse period altered the system. Combined with current climate trajectories, it is anticipated that summer season stratification in Lake 578 could intensify, possibly causing new phases of anoxia. For Lake 578, this would represent the third ‘type’ of anoxia, following the naturally caused post-deglaciation anoxia in the Early Holocene and the anthropogenically induced phase during the Norse period. These evolving dynamics underscore the complex interplay between climate change and lake ecosystems, highlighting the need for continued research and monitoring to understand and mitigate these impacts. Just as we need to understand climate variability of the past to better understand processes in the future, we should further explore the oxygenation and eutrophication states of lakes throughout the Holocene.

5. Conclusions

This study demonstrates the potential of integrating lipid biomarkers with hyperspectral imaging to elucidate Holocene-length lacustrine climate and human induced processes, crucial for anticipating future changes under global warming and increasing human influence. We find that Lake 578 experienced significant transitions, manifested by fluctuations in temperature, productivity, and anoxia, influenced by both natural climatic shifts and anthropogenic activities. Notably, the early Holocene was characterized by enhanced anoxic conditions, likely caused by a combination of strong seasonality and extended ice cover leading to pronounced stratification. The early Holocene temperature rise occurred later than suggested by ice core and physical model climate reconstructions, possibly highlighting the prolonged effects of the Laurentide Ice Sheet disintegration, as observed in the Atlantic-Fennoscandian stack, and other lacustrine reconstructions in the region.

Our findings align with other regional reconstructions of the Holocene Thermal Maximum (7.5–4.5 cal ka BP). Neoglacial cooling starts in the period between 4.5 and 3 cal ka BP and is associated with lower productivity and slightly increased anoxia. We attribute the surge in productivity and anoxia observed during the Norse period to a potentially human-induced lake-level increase that submerged nutrient-rich soils, combined with local land use. Despite a post-Norse decrease in productivity and anoxia, the lake never reverted to pre-Norse ecological baselines, emphasizing the long-lasting impact of human activities on lake systems. Modern sheep farming activities have increased productivity to high levels in recent decades. Conclusively, while early Holocene anoxia was probably due to natural cold temperature stratification, anoxia during the Norse period was primarily anthropogenically induced. The anticipated future warming may introduce a new regime of anoxia due to warm season stratification, indicating the need for continued monitoring and research on Arctic lakes.

Our research further underscores the importance of using a varied set of stratigraphic chronomarkers and exploring novel dating techniques to enhance the accuracy of age-depth models in this region. Furthermore, the selection of appropriate GDGT calibrations is crucial, suggesting that a Greenland-wide calibration set could significantly improve paleotemperature reconstructions.

These findings contribute to our general understanding of the nexus between climate, human activity, and Arctic lake ecosystems and emphasize the potential of the employed method combination in Arctic lakes. Further high-resolution multiproxy studies are essential to refine our understanding of these critical processes and for anticipating Arctic lakes’ responses to future global warming and anthropogenic influence.
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Credit author statement

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

Data availability
The dataset acquired for this project can be found at the NSF Arctic Data Center (https://arcticdata.io/), doi: https://doi.org/10.18739/A2M91C5S.

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Appendix A. Supplementary data
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