

<https://doi.org/10.1038/s43247-025-02064-z>

Hydroclimate intensification likely aided glacier survival on Svalbard in the Early Holocene



Andreea Gabriela Auer¹✉, Willem G. M. van der Bilt¹, Anders Schomacker², Jostein Bakke¹, Eivind W. N. Støren³, Joseph M. Buckby⁴, Jan Magne Cederstrøm¹ & Sander van der Plas⁵

Accelerated Arctic warming and wetting has global impacts, as the region's glaciers and ice caps respond to variations in temperature and precipitation, impacting global sea-level change. But as the observations needed to calibrate models are scarce, predictions cannot confirm if increases in snowfall can help offset melt. Here, we analyze two 14,000-year-long glacier-fed lake sediment records from the Svalbard archipelago to examine the response of a resilient ice cap (Åsgardfonna) to warmer-than-present Holocene Thermal Maximum conditions. End-Member Modelling allowed us to unmix the diluted grain size signal of rock flour – a widely used proxy for past glacier change, and surface runoff – an indicator of hydrological intensification. Our findings reveal that Åsgardfonna survived and may have advanced despite warmer conditions, possibly due to enhanced snowfall driven by sea-ice loss. This suggests that future increases in precipitation could moderate glacier retreat in similar settings.

Arctic climate change rates outpace the global average¹, as observations reveal that the region warms and wets more than twice as fast as the global average^{2–4}. The ongoing loss of Arctic sea ice plays an important role in this amplified response, as far more heat and moisture are transferred to the atmosphere from a seasonally ice-free surface ocean^{2,5–7}.

The regions' numerous glaciers and ice caps (GICs) are already dominant drivers of on-going sea-level rise^{8,9}, and respond to changes in both temperature and precipitation¹⁰. Therefore, the counteracting impacts of warmer (melt) as well as wetter (accumulation) conditions on the future evolution of Arctic glaciers can have global societal consequences¹¹.

Despite these ramifications, predictions of Arctic glacier mass balance and the associated sea-level contributions remain uncertain, as projections of the amount and phase of future precipitation change remain poorly constrained, in part because of a lack of robust observations. Arctic meteorological station data are scarce, and snowfall is often mismeasured because of wind-driven undercatch¹². As a result, simulations underestimate observed precipitation increases, while disagreeing on the seasonal distribution of future change^{4,6,13,14}.

Reconstructions of glacier-climate change from past warm periods can help answer how the uncertain interplay between warming and hydrological intensification may affect the future evolution of Arctic GICs. In this context, the Early Holocene (11.7–8.2 ka BP) is highly relevant as there is ample

evidence that conditions were warmer and wetter than today^{15–18}. Also, unlike geological archives from prior interglacials eroded by subsequent ice advances¹⁹, Early Holocene deposits have a high preservation potential. Critically, recent evidence suggests that some Arctic glaciers (temporarily) survived these conditions²⁰, and references therein.

By continuously tracking variations through time, sediment archives from proglacial lakes are particularly well-suited to reconstruct Early Holocene glacier-climate change²¹. Fundamentally, this approach builds on the proportional relation between the dimensions of warm-based or polythermal glaciers, erosion rates, and the production of glacial flour – the minerogenic product of glacial abrasion^{22,23}. Following evacuation in melt-water streams, this material is effectively trapped in distal glacier-fed lakes, continuously recording glacier changes through time^{24,25}. Recent method developments now provide us with multiple independent fingerprints of this glacier-climate signal on human-relevant (decades to centuries) timescales²⁶.

Here, we present a lake sediment-based reconstruction that resolves glacier variations throughout the culmination of Early Holocene warming on Svalbard. This archipelago is exceptionally sensitive to drivers of amplified Arctic glacier-climate change as reflected by record-breaking warming²⁷, the largest regional sea-ice loss²⁸, and high projected precipitation changes⁶. To warrant a continuous and representative Early Holocene

¹Department of Earth Science and Bjerknes Centre for Climate Research, University of Bergen, Bergen, Norway. ²Department of Geosciences, UiT The Arctic University of Norway, Tromsø, Norway. ³COWI AS, Bergen, Norway. ⁴Department of Arctic Geology, UNIS The University Centre in Svalbard, Longyearbyen, Norway. ⁵Huisartsenpraktijk Buis en van der Plas, Harderwijk, The Netherlands. ✉e-mail: andreea.auer@uib.no

glacier signal, we target **1**) two lakes – validating and upscaling site-specific findings using two ~14 ka long sediment records that are **2**) both fed by Åsgardfonna – an ice cap that can survive warmer-than-present conditions based on observational, modelling, and isostatic evidence^{29–31}. To rigorously resolve past changes on human-relevant (decades to centuries) timescales, we employ a multi-proxy toolbox that **1**) visualizes disturbances like event horizons in 3-D using X-ray Computed Tomography (CT), **2**) unmixes the grain size signal of glacier flour from other (overprinting) sediment sources using End-Member Modelling Analysis (EMMA), and **3**) utilizes sub-glacial elevation models and model output to link sediment transitions to past ice-marginal positions.

Results and Discussion

Setting

Our study sites, lakes *Berglivatnet* and *Lakssjøen* (79°42'N, 15°53'E), are both located on the eastern shore of Wijdefjorden in northeast Spitsbergen, Svalbard (Fig. 1). Lake Berglivatnet, the main focus of this study, measures 1.2 km², while our secondary site, Lake Lakssjøen, covers 3.6 km². The latter is separated from lake Røyesjøen by a shallow (~0.5 m deep) sill: Historical photographs reveal that this threshold has been subaerially exposed at times³². Situated at 46 and 74 meters above sea level (m a.s.l.), respectively, both basins lie above the postglacial marine limit in contrast to previously investigated lakes in the area^{33,34}. An index point at Gråhøken places the local marine limit at 42 m a.s.l.³⁵, while more recent work identifies a maximum of 65 m a.s.l. near Vassfarbukta³⁴ and 39 m a.s.l. in Flatøyrdalen³⁶. However, these studies

also highlight inordinate uncertainties and a need for further investigations of the local sea-level history.

Meltwater from one of the largest ice caps on Spitsbergen, 1230 km² Åsgardfonna³⁷, drains into both lakes. Lake Berglivatnet is fed by the Berglibreen valley glacier, while Lake Lakssjøen receives input from an unnamed lobe of the NW sector of Åsgardfonna (Fig. 2a). In contrast to glacial lakes previously investigated to constrain the past behavior of the ice cap³⁴, the glaciers that feed both lakes are classified as non-surge-type³⁸, and thus more likely to respond primarily to climate fluctuations²⁶. Satellite imagery reveals that Berglibreen has retreated ~0.23 km since 1936³⁹ – mere decades after the local culmination of the Little Ice Age (LIA)⁴⁰. In contrast, the unnamed glacier that feeds Lake Lakssjøen has remained stationary since³² (Fig. 2).

Beyond instrumental observations, marine sediment core data suggest that the adjacent Wijdefjorden deglaciated prior to the Younger Dryas period⁴¹ (YD: 12.9 – 11.7 cal. ka BP – calibrated kiloyear before present), while high sedimentation rates suggest particularly rapid glacier loss between 12.1 and 9.9 cal. ka BP⁴². Complementing this evidence, lake sediment data from nearby Lake Femmilsjøen³⁴ indicate the Åsgardfonna ice cap had retreated behind its present-day extent by 10.1 cal. ka BP. Iron fluxes into Wijdefjorden reveal limited glacial activity during the subsequent Middle Holocene⁴³, and basal ice-core ages suggest Åsgardfonna melted completely prior to 6 ka BP⁴⁴, although it should be noted that this chronology has been challenged⁴⁵. Therefore, evidence of the Early Holocene survival of Åsgardfonna remains ambiguous. However, glacier models and space-for-time substitution do suggest that the ice cap can survive

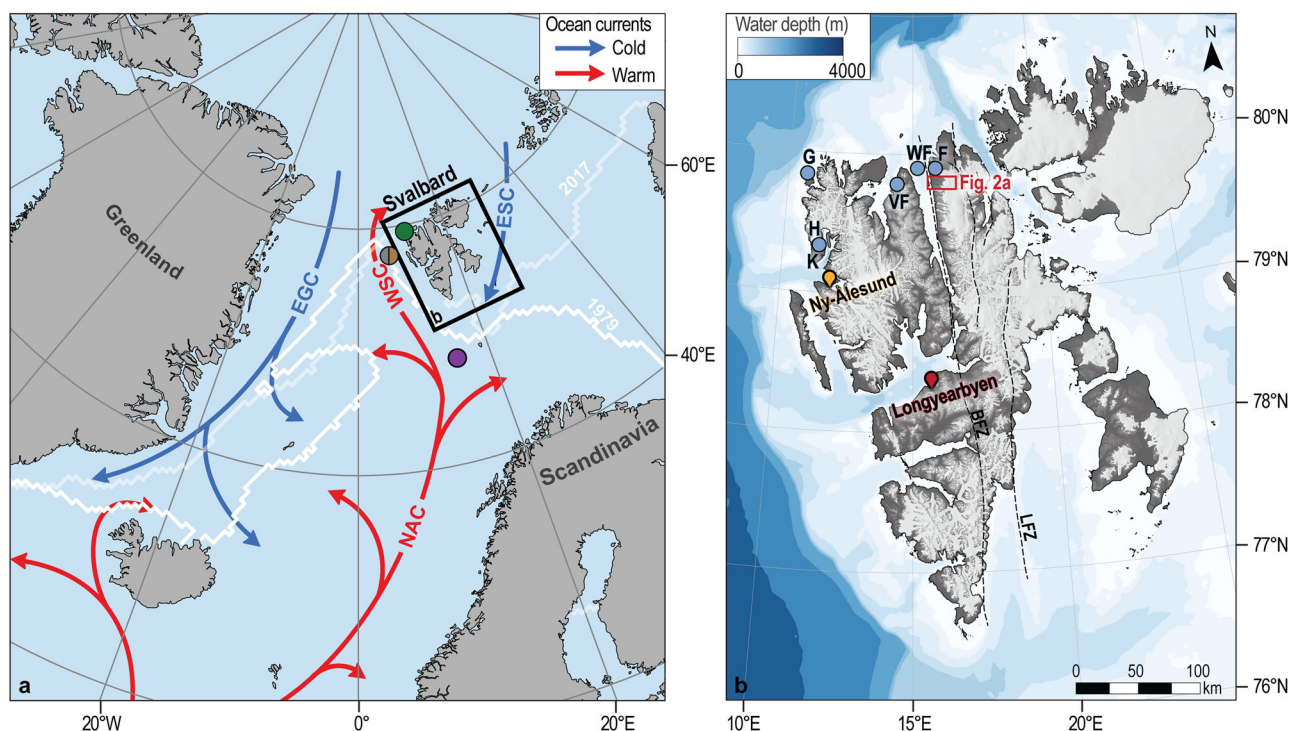


Fig. 1 | Regional maps of the study area. a Svalbard in a North Atlantic Arctic climate context. Warm Atlantic Ocean sourced currents are shown in red (NAC – North Atlantic Current, WSC – West Spitsbergen Current), and cold Arctic Ocean sourced currents are shown in blue (ESC – East Spitsbergen Current, EGC – East Greenland Current). Observed minimum (2017) and maximum (1979) March sea ice extents are marked with white lines¹³⁰. Localities of key proxy records used to contextualize our data are marked with colored circles corresponding to Fig. 7 – green: U_{37}^K -based preindustrial temperature anomaly composite record from lake Hakluytvatnet (core AMP112 between 0–5 ka BP and 7.5–9.5 cal. ka BP) and lake Gjøavatnet (core GJP0114 between 5–7.5 cal. ka BP)¹⁸, grey/brown: Ice-rafted debris (IRD) record (core MSM05/5-723-2)⁸⁴ and P_{BIP25} -derived sea ice cover

record (spliced data from core MSM5/5-712-2²³ between 0–8.7 cal. ka BP and core MSM05/5-723-2⁸⁴ between 8.7–9.5 cal. ka BP) from the Fram Strait, purple: U_{37}^K -based sea surface temperature record from the Barents Sea¹³¹. **b** Close-up of the archipelago, showing modern ice cover (white) and water depth after Jakobsson (2016)¹³². The location of our study area and Fig. 2 is highlighted with a red rectangle and the closest meteorological station, Ny-Ålesund, is highlighted with a yellow location marker. The locations of key glacier reconstructions are marked with blue dots and follow abbreviations as in Fig. 7 (F – Femmilsjøen, WF – Wijdefjorden, VF – Vårflusjøen, G – Gjøavatnet, K – Kløsa, H – Hajeren). The main fault zones adjacent to our study area are marked with stippled black lines (BFZ – Billefjorden Fault Zone, LFZ – Lomfjorden Fault Zone).

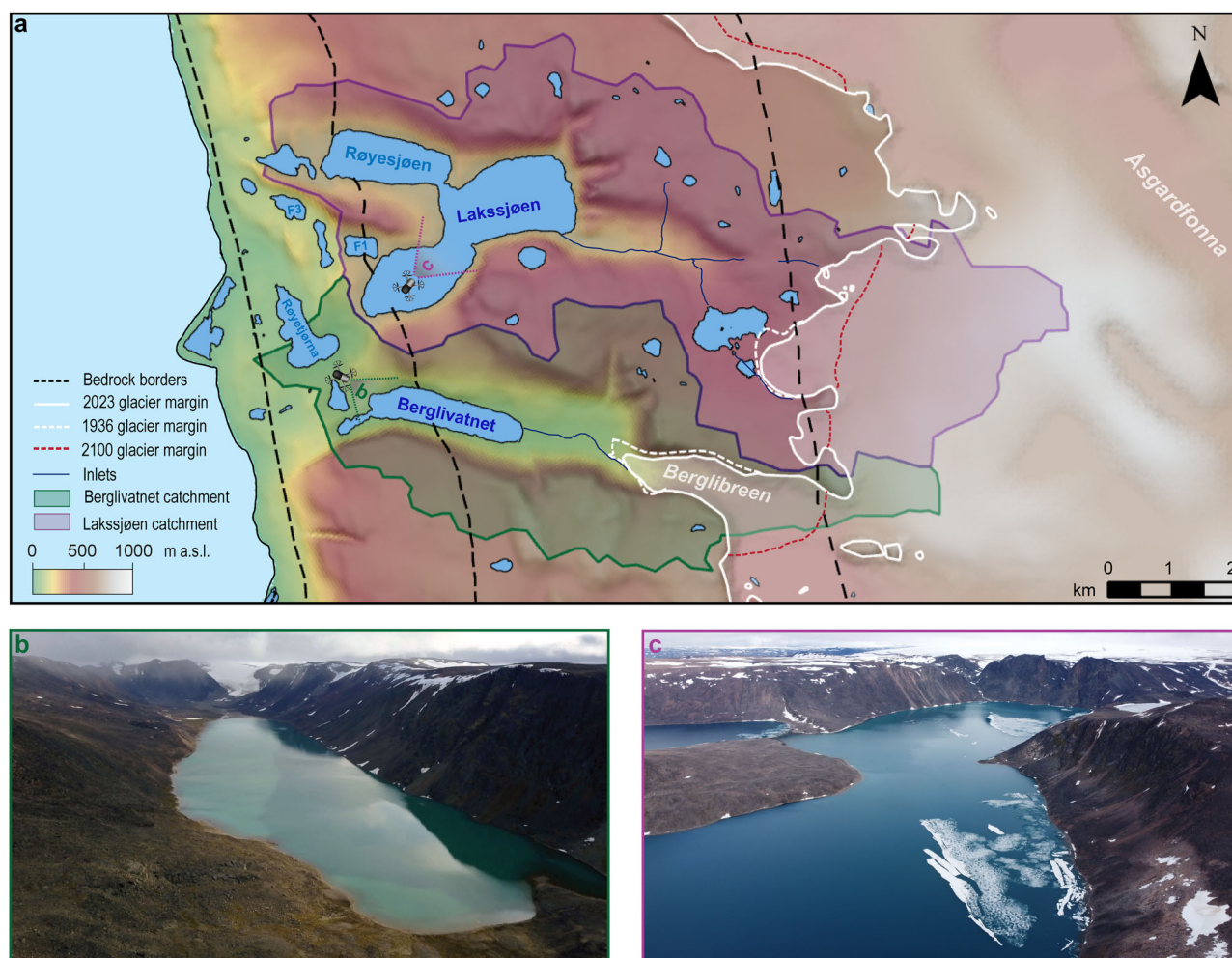


Fig. 2 | Study area overview and glacier catchments. **a** Surface and sub-glacial topography map of the study area generated using data from Farinotti et al.⁹⁰, Fürst et al.¹⁰⁵ and Porter et al.¹⁰⁶. Sub-glacial catchments of lakes Berglivatnet and Lakssjøen are marked with green and purple outlines, respectively. The present-day ice margin is marked with a white line after Pfeffer et al.¹²⁹, while the extended ice margin from 1936 CE is marked with a white dashed line after Luncke³² and the predicted ice margin for 2100 CE for Representative Concentration Pathway (RCP) 8.5 after Rounce et al.³¹ is marked with a red dashed line. Major bedrock boundaries

are marked with black and dashed lines after Dallman⁴⁷. Other lakes present in the wider study area are labelled with light blue (F1 – Fiskedammane 1, F3 – Fiskedammane 3). Lake outlines from © Norwegian Polar Institute **b** Unmanned Aerial Vehicle (UAV) photo of Lake Berglivatnet, taken during an August 2021 field campaign with a DJI Mavic 3, facing Berglibreen (photo by W.G.M. van der Bilt). **c** UAV photo, taken during the same field campaign, of the northern basin of Lake Lakssjøen and the shallow sill that separates it from Lake Røyesjøen towards the upper left (photo by W.G.M. van der Bilt).

Representative Concentration Pathway (RCP) 8.5 21st century warming of $\sim 8^{\circ}\text{C}$ ^{30,31}. This is comparable to reconstructed peak Early Holocene temperatures around 9.5 ka BP^{18,46}. This helped guide site selection and provides context for understanding the resilience of Åsgardfonna.

In both investigated catchments, the landscape is characterized by U-shaped valleys with a steep southern aspect ($40\text{--}60^{\circ}$), and gentler slopes towards the north ($10\text{--}30^{\circ}$). While the mountains are bare, high-Arctic tundra vegetation appears at lower elevations⁴⁷: field observations indicate the presence of moss mats around both lakes, with higher coverage ($\sim 30\%$) on the south-facing slopes. The investigated catchments are underlain by the metamorphic rocks of the Atomfjella antiform⁴⁷, which encompasses a stack of mineralogically distinct thrust sheets that are oriented perpendicular to the main direction of ice flow⁴⁸ (Fig. 2a). Changes in lake sediment geochemistry may thus be linked to past fluctuations of the ice margin position⁴⁹.

Regional climate is influenced by the dynamic confluence between warm Atlantic waters, transported North by the West Spitsbergen current (WSC), and polar waters exported from the Arctic by the East Spitsbergen current (ESC)⁵⁰ (Fig. 1a). The interplay between these water masses plays a critical role in local sea ice extent⁵¹, and therefore the

moisture (accumulation) and heat (melt) fluxes that drive regional glacier change⁵² – also see our introduction. Modern climate is characterized by maritime polar conditions, as the nearest weather station in Ny-Ålesund indicates a mean annual air temperature of -4.1°C and mean annual precipitation of $\sim 470\text{ mm}$ between 1991–2020^{53,54}. Notwithstanding the $\sim 120\text{ km}$ distance from our study area (Fig. 1b), a comparison of these values with down-scaled re-analysis data shows similar local climate conditions⁵⁵.

The two lake sediment records investigated reveal a near-identical lithostratigraphy (Fig. 3d, Suppl. Figure 1a). Within the covered timespan of the last $\sim 14\text{ ka}$, we observe similar transitions (Fig. 4, Suppl. Figure 2), which suggests that the records are representative of paleo-environmental changes in this sector of the Åsgardfonna ice cap. Combining evidence from visual inspection, cluster analysis and Principal Component Analysis (PCA) performed on selected sediment proxies (see Methods), we identify four main units (numbered 1 to 4 from top to bottom: see Fig. 5 and Suppl. Figure 3). Based on smaller-scale lithological changes, units 1 and 3 were further subdivided into subunits 1a to c and 3a and b, respectively. Each (sub)unit is interpreted to represent a different phase in the postglacial evolution of Åsgardfonna: deglaciation (unit 4), a seismic event (unit 3),

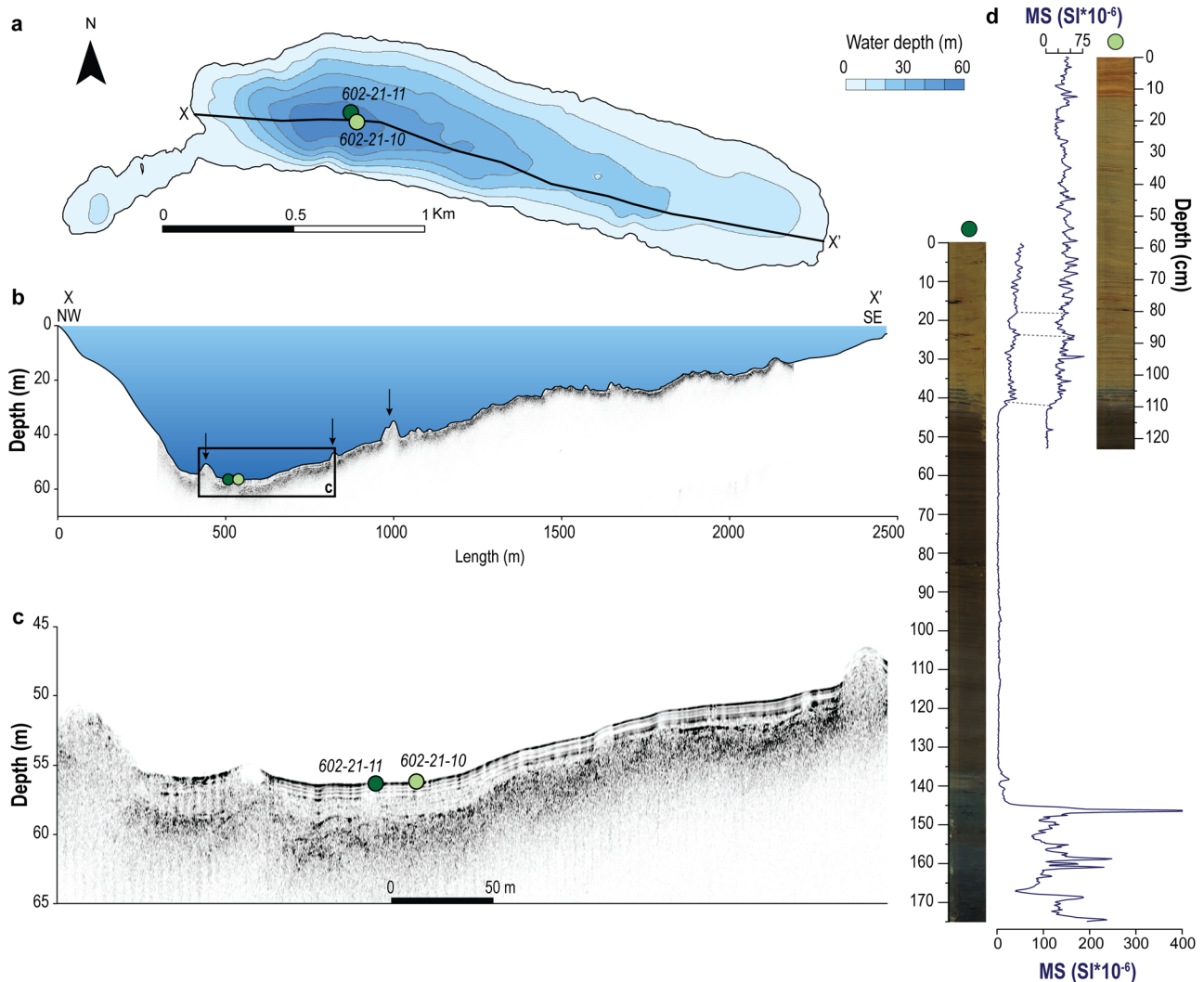


Fig. 3 | Bathymetric and seismic surveys of Lake Berglivatnet. **a** bathymetric map of Lake Berglivatnet, showing our coring locations (green dots) and seismic (CHIRP) profile (black line). **b** composite NW-SE-trending bathymetric-seismic profile that crosses both coring locations (green dots). Black arrows indicate the location of

multiple mounds observed on top of the acoustic basement and identified as moraines. **c** close-up, delineated by the black rectangle in panel **b**, highlighting the seismic stratigraphy of our coring locations (green dots). **d** Magnetic susceptibility (MS)-based alignment (see Methods) of the Berglivatnet composite record.

followed by glacier retreat when ice moves behind modern margins (unit 2), and finally glacier re-growth (unit 1).

Core chronology

All extracted radiocarbon (^{14}C) samples taken from the analyzed cores from lakes Berglivatnet and Lakssjøen are terrestrial macrofossil-derived (see Methods) and were calibrated using the IntCal20 northern hemisphere calibration curve⁵⁶. Age-depth models were subsequently produced using default settings in version 3 of the rbacon package in Rstudio⁵⁷.

All dates were included in both age models, with the exception of one deriving from the event deposit of unit 3 (see our Late glacial- Holocene lake evolution paragraph in Results & Discussion and Suppl. Table 1: sample Poz-157339). Here, we followed the recommendations of Sabatier et al. (2022)⁵⁸, and 1) assumed instantaneous deposition by excising the 216–224.5 cm and 56–75.5 cm unit 3 depth intervals from the Berglivatnet and Lakssjøen chronologies, respectively, before 2) determining upper and lower age limits by running separate age models for the deposits that adjoin unit 3 (see Fig. 4 and Suppl. Figure 2). Based on CT evidence of basal disturbance and erosion (see Late glacial- Holocene lake evolution paragraph in Results & Discussion and Suppl. Fig. 5f and 8), we use the coevally dated samples near the top of unit 3 in both deposits to constrain its age to ~9.5 cal. ka BP. In Berglivatnet, the excluded 10.6 cal. ka BP age was sampled

from a section with re-worked unit 2 sediments (see Late glacial- Holocene lake evolution paragraph in Results & Discussion) and marks the possible onset of organic sedimentation. The two chronologies date the onset of lake sedimentation to ~14 cal. ka BP in each basin (see Fig. 4 and Suppl. Fig. 2).

While both age models yield consistent ages for unit transitions, we observe marked differences in accumulation rates. As can be seen in Fig. 4 and Suppl. Fig. 2, Holocene sedimentation rates are more than an order of magnitude higher in Berglivatnet (0.33 cm/yr) compared to Lakssjøen (0.01 cm/yr). We attribute this difference to 1) accommodation space – Lakssjøen is three times larger (Suppl. Figs. 1b), 2) the presence of a small ice-contact lake along the path of one of the main inlets into Lakssjøen (Figs. 2a) and 3) glacial regime – minimal changes since 1936 suggest a cold-based regime for the unnamed outlet feeding Lakssjøen³⁰, with its front remaining stationary⁵⁹. As this study primarily seeks to resolve Early Holocene glacier-climate change on human-relevant (decades to centuries) resolution (see the introduction), we focus our discussion mainly on the Berglivatnet record.

Late glacial- Holocene Lake evolution

Basal unit 4, covering 14.5 cal. ka BP to 10.6 cal. ka BP in Berglivatnet (223.5–261 cm core depth) and ~13.2–10.2 cal. ka BP in Lakssjøen (75–225 cm core depth), consists of pebbles and small cobbles embedded in

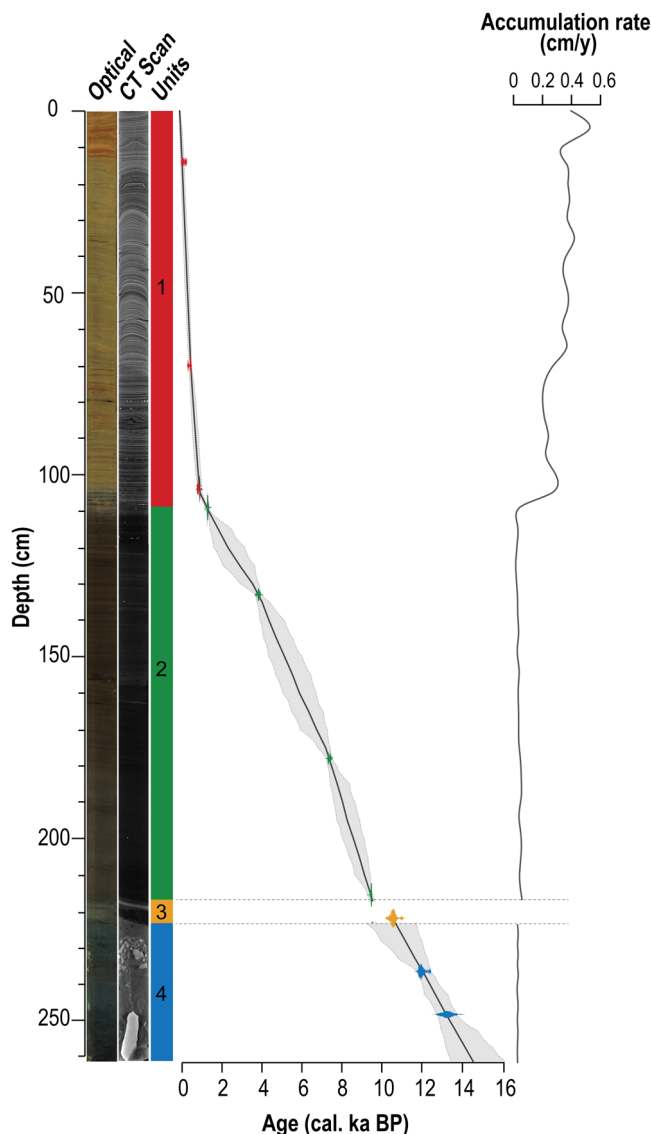


Fig. 4 | Berglivatnet chronology. The black line shows the weighted mean of our model, while grey shading delimits its 95% confidence limits. On the far left, we show optical and Computed Tomography (CT) imagery, as well as a log with unit divisions. ^{14}C sample age distributions are color-coded according to the units. Dashed lines mark the extent of the unit 3 event deposit (see the Postglacial neotectonic seismicity paragraph in Results & Discussion). Modelled accumulation rates are plotted on the right.

a silty matrix (Fig. 5 and Suppl. Figure 3). Detailed investigation of our CT imagery reveals that these clasts are sub-angular and often underlain by deformed sediments. Based on these characteristics, we identify them as dropstones after e.g., Carrivick & Tweed⁶⁰. The CT imagery for Berglivatnet also shows that the largest of these ($\varnothing$ 10 cm) appears to have been dragged down, presumably during coring. Because of this disturbance, we primarily rely on the unit 4 deposit in Lakssjøen for stratigraphic interpretation (see Suppl. Fig. 3). High Total Scatter Normalized (TSN) Titanium (Ti) ratios—a widely used indicator of clastic input⁶¹, as well as low Loss on Ignition (LOI) values (2–3 %) and Inc./Coh. scattering ratios—both measures of organic content⁶², reflect a minerogenic character. Compositionally, unit 4 sediments are furthermore characterized by a high Calcium (Ca) content (Fig. 5). While Ca is often used as a marine indicator⁶¹, we favor a clastic minerogenic source because 1) our sites are sitting above the local postglacial marine limit (see Setting section), and 2) Ca co-varies with Ti (Suppl. Table 2). Unit 4 sediments are furthermore dense, as shown by high

average dry bulk density (DBD) values of 1.5 g/cm^3 and CT grayscale values of ~ 5500 (Fig. 5b). Based on the observed diagnostic association of minerogenic sediments with dropstones⁶³, we argue that unit 4 was deposited in an ice-contact lake⁶³. In keeping with this interpretation, we identify the mounds observed on top of the acoustic basement in Berglivatnet (see Fig. 3b), perpendicular to the direction of ice flow, as moraines after Trotter et al.⁶⁴.

Unit 3, identified as an event deposit focused around 9.5 cal. ka BP (Berglivatnet: 217–223.5 cm core depth, Lakssjøen: 57–75 cm core depth), is characterized by two abrupt shifts across all plotted proxies (Berglivatnet—Fig. 5: 220 cm & 223.5 cm, Lakssjøen—Suppl. Figure 3: 60 cm & 75 cm). Based on our cluster analysis output (Suppl. Figure 4), these changes warrant further division into subunits 3a and b.

Lowermost **subunit 3b** is ~ 3.5 cm thick (15 cm in Lakssjøen) and characterized by a higher organic content than underlying unit 4, as marked by steep decreases in Ti and Ca counts, and an increase in Inc./Coh. ratios and LOI from 2 to 6% as can be seen in Fig. 5b. In contrast, super-imposed **subunit 3a** (3 cm thick in both lakes) has a denser, more minerogenic composition as shown by higher Ti counts ($\sim 3\%$ TSN) and increased DBD values (Fig. 5b), as well as a Ca content that is similar to that of unit 4 (max 2.5 % of TSN). These different characteristics are also reflected by our PCA output as subunit 3b shows a strong association with unit 2, while the composition of subunit 3a is similar to that of unit 4 (Fig. 6, Suppl. table 2).

Closer examination of our CT imagery allows for more detailed stratigraphic investigation (Fig. 5 and Suppl. Figure 5f and 8). The base of subunit 3b bears evidence of soft sediment deformation structures (SSDS), notably intraclast breccias that consist of a mix between lower-density mud clasts that appear darker on CT imagery, embedded in a denser matrix of unit 4 sediments, which appear lighter on the CT imagery. An erosional base separates this boundary interface from the remainder of subunit 3b, which consists of a sequence of low density (dark on the CT imagery) sediments that appears to have slumped based on the presence of tilted (vertical) laminae (Suppl. Fig. 8). Based on this evidence of slope failure, we interpret this subunit as a mass-transport deposit (MTD). Moreover, based on the aforementioned sedimentological characteristics and our PCA output (Fig. 6), we argue that this material consists of reworked unit 2 sediments.

Finally, **subunit 3a** sediments are normally graded (fining upwards) as seen in Fig. 5 and therefore interpreted as a turbidite. PCA output suggests that subunit 3a has a similar composition to unit 4 (Fig. 6). Based on this distinct association between soft-sediment deformation structures (SSDS), a mass-transport deposit (MTD) and a capping turbidite - that all suggest an in-lake composition based on the presented geochemical proxy evidence (see Fig. 6)—we interpret unit 3 as an earthquake-triggered deposit following the classification of Sabatier et al. (2022)⁵⁸. Earthquakes could have been caused by rapid glacio-isostatic rebound during and after deglaciation⁶⁵. In contrast, in Lakssjøen, we note the absence of an erosional surface at the base of unit 3, the absence of slumped deposits and the presence of a thicker mud clast conglomerate (subunit 3b 58.5–75 cm) (Suppl. Figure 3 and 8). We attribute these variations to a difference in slope steepness: the coring location in Berglivatnet is closer to steep ($\sim 50^\circ$) basin slopes, while Lakssjøen is more gentle ($\sim 25^\circ$).

Unit 2 extends from 9.5 cal. ka BP to 1.3 cal. ka BP in Berglivatnet (108.5–217 cm core depth) and ~ 9.3 to 1.2 cal. ka BP in Lakssjøen (4.1–57 cm core depth). Its comparatively organic character is reflected by steadily increasing LOI (6 to 15%) and Inc./Coh. values (Fig. 5b). Simultaneously, consistently low Ti, Ca and DBD values suggest little minerogenic input (Fig. 5b). However, grain size analysis reveals that the particle size distribution of this reduced clastic component is highly variable. As can be seen in Fig. 5d, we observe multi-modal distributions with peaks in the silt ($12\text{--}63 \mu\text{m}$) fractions. EMMA output reveals three significant End Members (EMs) that together explain 99% of all data variance (Fig. 5e). Differentiated by mean grain size (values in μm), EM1 consists of fine silt ($6 \mu\text{m}$), EM 2 is dominated by medium silt ($14 \mu\text{m}$) and EM 3 by coarse silt ($55 \mu\text{m}$). We note that while EM 1 remains present throughout unit 2, EM 2 and EM 3 dominate the unit. Catchment samples collected near the spring flood limit

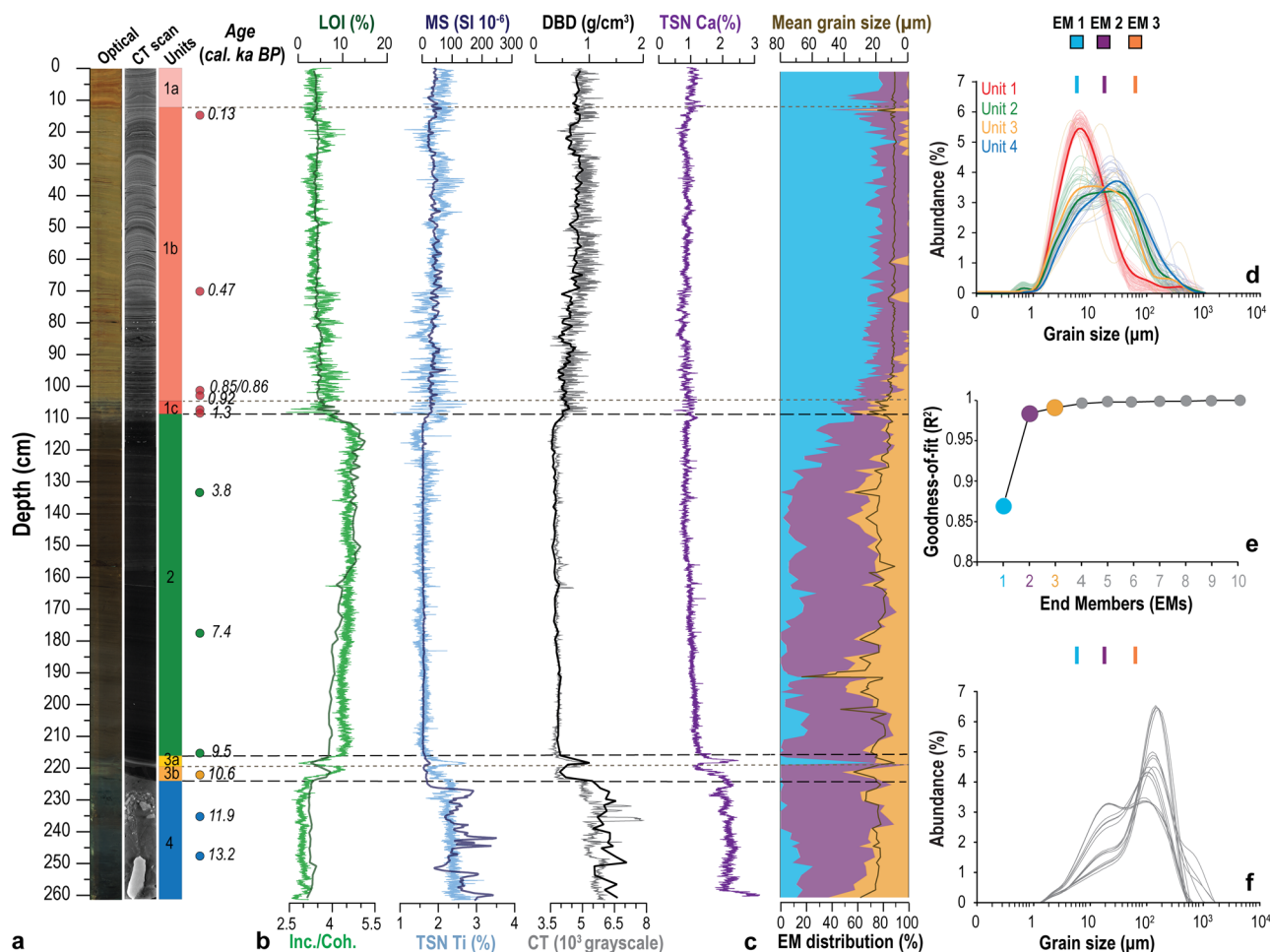


Fig. 5 | Berglivatnet stratigraphy. **a** From left to right - optical and Computed Tomography (CT) imagery of the sedimentary record, (sub)units classification and position of ^{14}C samples. **b** organic content—reflected by Loss on Ignition (LOI) % and XRF incoherent/coherent scattering (Inc./Coh.) ratios, minerogenic indicators – MS and Titanium (Ti) counts, density – reflected by Dry Bulk Density (DBD) values (bottom) and CT grayscale values (top), and provenance indicator Calcium (Ca). **c** End-Member Modelling Analysis (EMMA) results for Berglivatnet. **d** Grain size distributions of samples from each unit, color-coded accordingly, with

individual samples shown in the background and an average per unit shown bolded in the foreground. The grain size means of each meaningful EM are also marked. **e** Cumulative coefficients of determination (R^2 : goodness-of-fit) for the ten EM model that we ran, highlighting the three significant EMs. **f** Grain size distribution of catchment samples shown in grey and grain size means of each meaningful EM displayed above. See Suppl. Figure 5 for a more detailed overview of the EMMA results. Unit divisions are marked with black stippled lines, while subunits are separated with grey dotted lines.

of the main inlet of Berglivatnet (Fig. 5f) have similar bi-modal distributions. The finer mean matches that of EM 2 and that of unit 4 samples, while the coarser mean overlaps with EM 3. Based on these observations, we argue that EM 2 reflects the input of fluvially-eroded ice-proximal glacial deposits from the catchment and attribute the coarser grain size of EM 3 to sorting during runoff peaks.

Finally, **unit 1** covers the last 1.3 cal. ka BP (top 108.5 cm) in Berglivatnet or the last ~1.2 cal. ka BP (uppermost 4.1 cm) in Lakssjøen, and consists of light brown to orange silty clay with well-preserved mm-scale laminations (Fig. 5a). The increase in Ti counts and DBD values compared to unit 2 shows a gradual transition to more minerogenic sedimentation. This is further supported by very low LOI values (3–4%) and co-varying Inc./Coh. scattering ratios. A muted increase in Ca input suggests that the provenance of clastic input in unit 1 is different than in unit 4. Compositionally, we also note a sharp increase in redox-sensitive elements Iron (Fe) and Manganese (Mn)⁶¹. As highlighted by the ordination diagram of Fig. 6, unit 1 samples cluster around Mn and Fe with progressively stronger loadings on PC1 from the base (subunit 1c) towards the top (subunit 1a). Grain size analysis shows a shift from the poorly sorted medium silts that dominate unit 2 to very fine silts that are dominated (70%) by EM 1 (Fig. 5c). As field observations suggest that modern-day sediment input is dominated

by glacial input, we argue that EM 1 reflects glacial flour input. This notion is further supported by suspended sediment samples from other contemporaneously glaciated catchments^{24,66,67}, which reveal a near-identical fine silt-dominated grain size. Accordingly, we argue that variations in fine-grained minerogenic sediment fluxes in unit 1 reflect changes in glacier activity.

Ice margin oscillations and Younger Dryas retreat during deglaciation

The presented CT imagery, our granulometric data, measures of organic content (LOI, Inc./Coh.), as well as minerogenic proxies (MS, Ti, Ca), all indicate highly variable depositional conditions in the undisturbed ice-contact lake sediments of unit 4 in Lake Lakssjøen (see Suppl. Figure 3 and the Late glacial- Holocene lake evolution in Results & Discussion). We specifically observe alternations between 1) comparatively coarse (EM 3-dominated) sediments with chaotic beds that are indicative of deposition from ice-proximal sub-aqueous fans in similar settings⁶³, and 2) silty (EM 1-dominated) laminated sequences that are associated with more ice-distal conditions⁶⁰.

Following from the above, we argue that the ~14–10 cal. ka BP period was marked by regular ice margin oscillations in our study area, although the

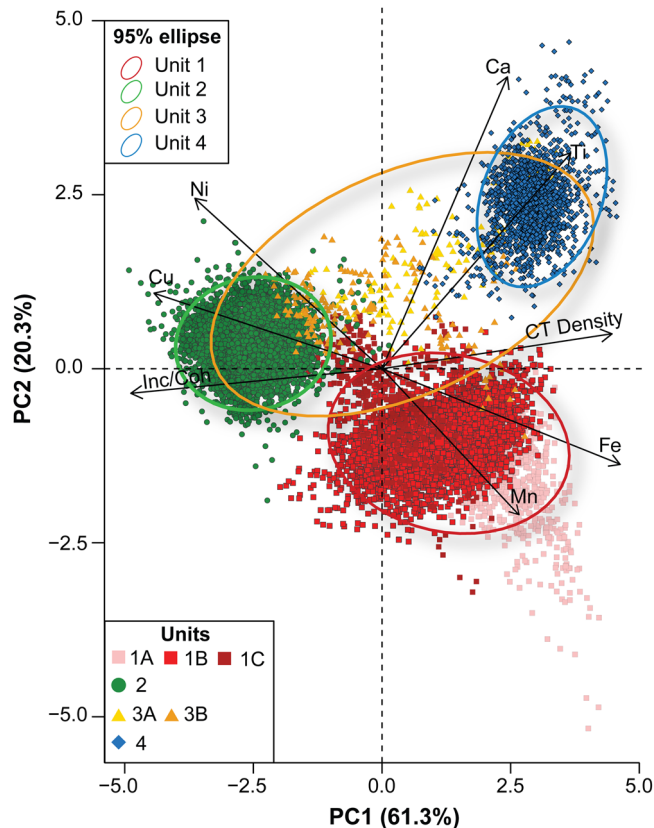


Fig. 6 | Berglivatnet ordination diagram. Down-core samples and 95% prediction ellipses are coded in colors matching those of the (sub)unit classification shown in Figs. 4 and 5. The vectors (arrows) reflect the scores of selected variables on the first (PC 1) and second (PC 2) principal components (PCs) with the greatest explanatory power (listed on both axes).

comparatively poor chronological control on unit 4 does not allow us to ascertain on what timescales (Fig. 4 and Suppl. Figure 2). The unique Ca-rich provenance of minerogenic input in unit 4 (see Fig. 6), which is not a major constituent of the metamorphic bedrock that underlies the investigated catchments, suggests a non-local source of glacial input. Based on the ubiquitous presence of Carboniferous carbonates of the Campbellryggen subgroup in inner Wijdefjorden⁴⁷, which contain very high Ca concentrations of up to 21%⁶⁸, we argue that there was regional ice terminating in Berglivatnet and Lakssjøen during deposition of unit 4, sourced from the south. In support of this interpretation, provenance studies of erratic boulders in our study area also indicate that ice drained north at the time⁶⁹. Contemporaneously, the moraines observed in Berglivatnet (see Late glacial–Holocene lake evolution paragraph in Results & Discussion and Fig. 3b), were deposited by local ice flow from the east.

While we cannot attribute the aforementioned ice margin oscillations between ~14–10 cal. ka BP to either (internal) ice dynamics or (external) climate forcing, it is worth noting that the undisturbed unit 4 stratigraphy of Lakssjøen reveals a shift towards more ice-distal (laminated) conditions during the Younger Dryas (12.9 – 11.7 cal. ka BP) stadial (see Suppl. Figure 3). Indeed, organic content peaks at 10% LOI values also observed during the (Early) Holocene, while the terrestrial origin of the plant macrofossil dated ~12.5 cal. ka BP in this deposit also suggests climatic amelioration (see Suppl. Table 1). These findings contribute to a decades-long debate about the behavior of the GICs of Svalbard during the Younger Dryas⁷⁰, and support recent evidence from glacially-derived (Fe)-oxide fluxes that show this stadial was marked by rapid ice retreat in Wijdefjorden⁴³. From a paleoclimate perspective, the presented evidence for land vegetation (terrestrial plant fossils) and high productivity (LOI) helps explain why thermophilous species were already present at the onset of the

Holocene in nearby Ringhornsdalen⁷¹. Beyond our study area, the above (growing season) data complement a growing body of evidence that suggests that the Younger Dryas stadial was characterized by mild summers and severe winters^{72,73}, rather than year-round cooling.

Postglacial neotectonic seismicity

The identification of the diagnostic SSDS and MTDs of lacustrine seismites relies on the CT image-based approach applied here and pioneered by Oswald et al. (2021)⁷⁴. Based on this analysis unit 3 provides evidence of an earthquake-triggered deposit on Svalbard. While this sequence is directly underlain by the deglacial ice-contact lake deposit of unit 4 (see the Ice margin oscillations and Younger Dryas retreat during deglaciation section in Results & Discussion), we do not think seismicity was related to a Glacial Lake Outburst Flood (GLOF). Notably, because our XRF-PCA analysis in Fig. 6 shows that basal subunit 3b is compositionally similar to the comparatively organic overlying unit 2 sediments. Therefore, the seismic event that deposited unit 3 must have been triggered when both sites were no longer ice-contact lakes. We argue that seismic activity was likely caused by a combination of 1) rapid emergence of the study area⁷⁵ – local uplift rates peaked at 2 m/century when unit 3 was deposited ~9.5 cal. ka BP³⁵, and 2) ensuing fault activation – our study area is adjoined by two fault zones: the Billefjorden fault zone (BFZ) to the West⁷⁶, and the Lomfjorden fault zone (LFZ) to the East (see Fig. 1b)⁷⁷. In this regard, we should note that movement along an extension of the LFZ triggered a strong (M_w 6.1) earthquake in 2008⁷⁸, which attests to its activity. While future uplift rates will undoubtedly be lower than during the Early Holocene, our findings suggest that isostatic adjustments can trigger seismic activity in the region.

Glacier survival during the Holocene Thermal Maximum

Based on our observation-validated interpretation that EM 1 captures the influx of glacier flour and therefore records changes in glacier variability (see our Late glacial–Holocene lake evolution section in Results & Discussion), Fig. 7 shows that Åsgardfonna likely survived 1) the culmination of the Holocene Thermal Maximum (HTM) on Svalbard ~9.5 ka BP^{18,46,79,80}, and 2) the subsequent local Holocene glacier minimum ~7 cal. ka BP^{29,81} (Fig. 7b). And while Fig. 7d shows that EM 1 concentrations are negligible in 8 out of 260 samples, we stress that this only indicates that Berglibreen retreated out of the sub-glacial catchment shown in Fig. 2a, rather than complete disappearance. At this threshold, ~4 km from the present-day glacier margin (see Fig. 2a), a sizeable (~50%) ice cap would still sit on the highest (most resilient) part of the plateau. Importantly, given a response time of at least 880 years for a GIC the size of Åsgardfonna⁸², the observed centennial-scale EM 1 minima which vary between 70 – 280 years, are too brief for complete melt (Fig. 7d), even under on-going and predicted rates of change³⁰. In support of this evidence, the modelled response of Åsgardfonna to RCP 8.5 forcing – which predicts summer temperatures that are ~8 °C higher than today and on-par with HTM estimates on Svalbard^{18,31} – reveals that ice remains in the sub-glacial catchments of Berglibreen as well as Lakssjøen by 2100 CE (Fig. 2 and Suppl. Fig. 9). Analytically, we should also highlight the uncertainties of replicate EM measurements (see Fig. 7d), and stress that log transformation reveals that the observed variations do not stem from the impacts of closed-sum effects on our compositional EM data (see Suppl. Figure 6)⁸³. In further support of our claim that Åsgardfonna survived the HTM, we also note broad similarities between Early Holocene advances (EM 1 increases) and Ice Rafted Debris (IRD) fluxes from similarly sized marine-terminating glaciers on western Spitsbergen with a comparable response time as Åsgardfonna⁸² (see Fig. 7f). Even more so, the most prominent of these advances around 8 and 9 cal. ka BP are also captured by changes in glacially derived (Fe)-oxide input in Wijdefjorden⁴³. In conclusion, we argue that these findings provide terrestrial evidence of continuous GIC survival throughout the warmest part of the HTM on Svalbard (Fig. 7a, b)⁸¹.

Following from the above, the variations in EM 1-derived glacier input to Berglivatnet allow us to investigate the drivers of glacier–climate change on Svalbard between the culmination of the Holocene Thermal Maximum

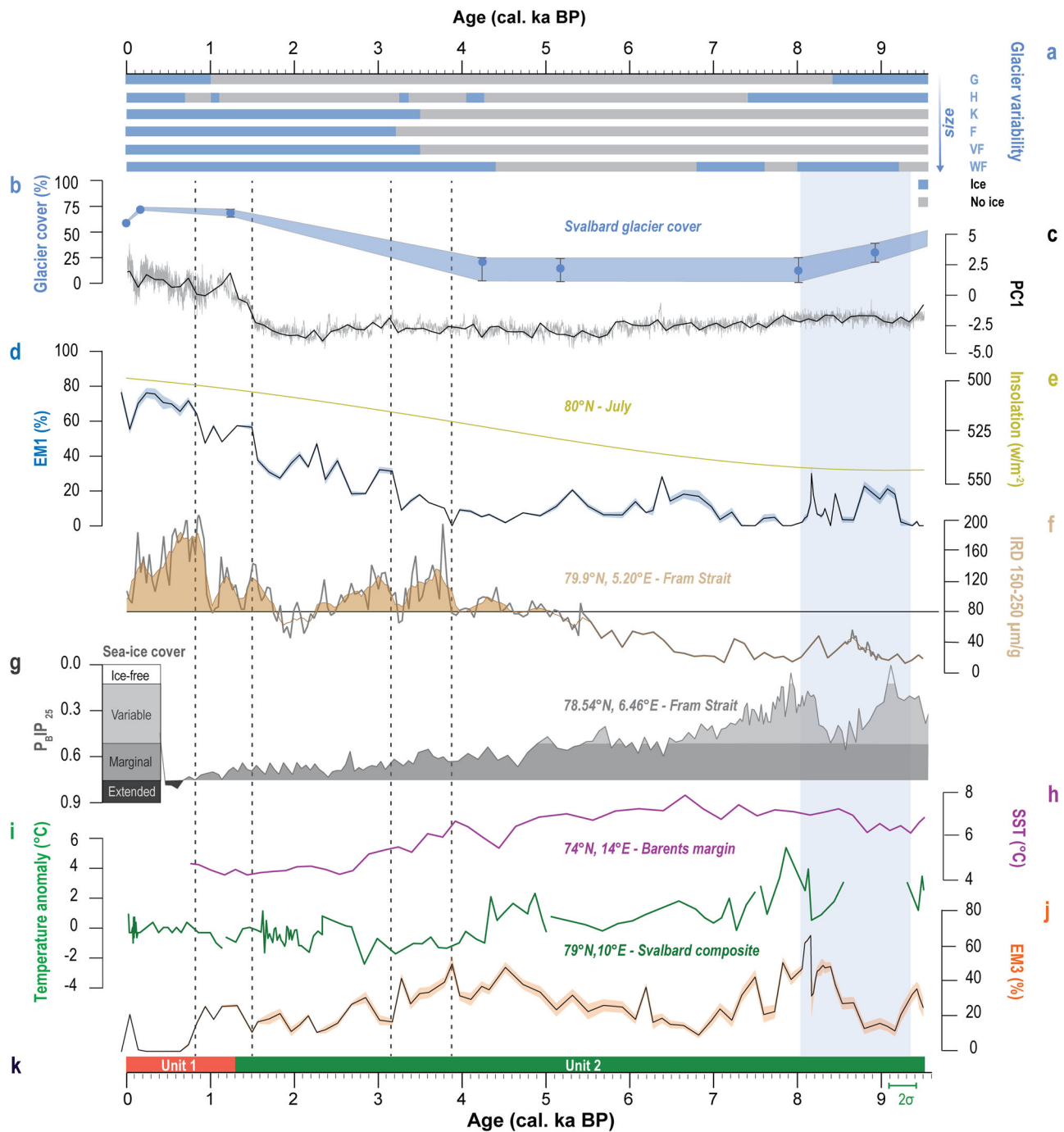


Fig. 7 | Selected regional paleoclimate records. **a** Timeline of Holocene glacier activity from near-by studies (WF - Wijdefjorden⁴², V - Vårflusjøen⁴⁷, F - Femmilsjøen³⁴, K - Kløsa⁹⁶, H - Hajeren⁹³, G - Gjøavatnet⁹⁴). **b** estimated percentage of glacier cover on Svalbard (modified after Fjeldskaar et al.²⁹ and Farnsworth et al.⁸¹). **c** XRF-based PC1 glacier activity reconstruction - light gray, and 100-year averages - black (this study), **d** Grain Size (GS) End-Member Modelling Analysis (EMMA)-based glacier activity reconstruction (this study: EM 1) - blue shading highlights the analytical uncertainty (2σ) of measurements. **e** Summer insolation at 80°N ¹²⁴ **f** Ice-rafted debris (IRD) records from core MSM05/5-723-2⁸⁴ in the Fram Strait, showing 5-point smoothed averages (brown), a mean (horizontal line), and curve-filling values exceeding the calculated mean. **g** P_{IP25} -derived sea ice extent from the Fram Strait. The composite record shows spliced data from core MSM05/5-712-2⁵² between 0 - 8.7 cal. ka BP and core MSM05/5-723-2⁸⁴ between 8.7 - 9.5 cal. ka BP. Gray shadings are in accordance with estimates of sea ice conditions

($P_{IP25} > 0.1$ variable, > 0.5 marginal, > 0.75 extended ice cover) after Müller et al.¹³³ **h** U_{37}^K -based summer sea surface temperature (SST) reconstruction from the Barents Sea margin¹³¹. **i** compilation of U_{37}^K -based preindustrial surface summer temperatures from a compilation of non-glacial Svalbard lakes (Hakluyvatnet: core AMP112 between 0-5 ka BP and 7.5 - 9.5 cal. ka BP, Gjøavatnet: core GJP0114 between 5 - 7.5 cal. ka BP)^{18,134}. **j** GS EMMA-based runoff intensity (this study: EM 3) - orange shading highlights the analytical uncertainty (2σ) of measurements. **k** units classification of the Berglivaatnet record (color coding corresponds with that of Fig. 5). The light blue vertical rectangle in the background highlights the extent of Early Holocene Berglivaatnet advances. Vertical black stippled lines mark identified phases of glacier (re-)growth based on Change Point Detection (CPD: see Methods). The green horizontal bar marks the average calibrated 2σ age uncertainty range for the Early Holocene interval of unit 2 based on the presented age-depth model (see Fig. 4).

(HTM) and onset of Neoglaciation (Fig. 7b). As outlined in the introduction, the former period is of particular relevance by providing a glimpse into the future, as summer surface temperatures were up to 9 °C higher than today (Fig. 7i)¹⁸. As such pronounced warming greatly enhances melt rates, we here investigate whether an increase in accumulation – the other major climatic driver of glacier change¹⁰ – helped offset mass loss. To do so, we first cross-correlate glacial EM 1 and our EM 3-based runoff indicators, to account for offsets due to the afore-mentioned multi-centennial response time of Åsgardfonna⁸². In contrast, hydroclimate proxies like EM 3 have a near-instant response to changes in precipitation and/or runoff. As can be seen in Suppl. Fig. 10, we find a robust correlation ($\rho = 0.72$, $p < 0.0001$), with a mean lag of 1.6 ka for EM1. This offset corresponds to the expected response times for GICs the size of Berglibreen (~5 km²) to NW Åsgardfonna (~50 km²)⁸², as shown by the inset of Suppl. Fig. 10. Beyond these site-specific findings, we also compare our EM data to biomarker-based (P_BIP₂₅) summer sea ice data from the adjacent Fram Strait^{52,84} (Fig. 1a). While direct proxies of Arctic hydroclimate change remain ambiguous^{85,86}, observations identify sea-ice cover as the primary control on regional moisture fluxes²⁷. As can be seen in Fig. 7 and Suppl. Fig. 12, EM 3 indeed increases during phases of less extensive sea-ice conditions (and vice versa) between 9.5–7 ka BP, further suggesting that hydrological intensification aided HTM glacier survival.

We cannot directly attribute EM 3-derived runoff changes to either increases in precipitation and/or melt rates, although other local lake proxy data provide useful insight. Notably, an excursion towards more negative leaf wax hydrogen isotope ratios in nearby Austre Nevlingen³³, as well as an increase in snowmelt layer thickness in Vårflusjøen on adjacent Andree Land⁸⁷ (Fig. 1b). Therefore, both records indicate that winter precipitation increased between 9.5 and 7 cal. ka BP, in tandem with more variable sea ice conditions (Fig. 7g). To help explain why this relation decouples after 7 ka BP, we invoke changes in runoff seasonality, linked to insolation changes. Specifically, a gradual lengthening of the snow (winter) season, as summer insolation decreased from the Mid-Holocene onwards (Fig. 7e). We argue that, as a result, EM 3-derived runoff progressively captures a runoff signal reflective of melt rather than precipitation towards the Late Holocene. To support this argument, we calculate the inflection points of spring, summer and autumn insolation changes at 80° N, to derive a weighted average. As shown in Suppl. Fig. 11, this analysis reveals a shift in seasonality occurred around 7 ka BP. Considering the above, we hypothesize that Early Holocene changes of Åsgardfonna were driven by increases in snowfall.

To further investigate the role of hydrological intensification on the inferred HTM survival of Berglibreen and Åsgardfonna, and place these findings in a modern context, we compare our results with simulations of the Equilibrium Line Altitude (ELA) or snowline. Unlike modelled glacier dimensions, ELAs respond directly to a change in climate forcing (no response time)⁸⁸. Available RCP 8.5 scenario simulations for 2071–2100 CE suggest the snowline of Åsgardfonna will rise above the modern topography⁸⁹, ensuing disappearance of the ice cap. This contrasts with the presented EM 1-derived proxy evidence, which suggests glacier survival under similar (~8 °C) levels of summer (melt season) warming. Therefore, our results leave the possibility open that recent (CMIP5 and 6) generations of climate models still under-estimate changes in the phase and magnitude of Arctic precipitation change as previously suggested by i.e. McCrystall et al.⁴, Bintanja & Selten⁶.

In conclusion, End-Member Modelling Analysis (EMMA) enabled us to isolate the (diluted) sedimentological imprint of HTM glacial input (EM 1) and surface runoff (EM 3). By unmixing these processes, our evidence suggests that Åsgardfonna not only survived warmer-than-present conditions during the culmination of the HTM, but also experienced brief advances following periods of hydrological intensification (see Fig. 7). Comparison to other climate reconstructions from Svalbard suggests that phases of glacier growth occurred in tandem with sea ice loss – the dominant control on surface moisture fluxes²⁷, and an increase in winter precipitation – a key constraint on ice accumulation¹⁰. As both of these processes will

determine the poorly constrained magnitude and phase of amplified future Arctic hydroclimate change, our findings raise the possibility that, in regions close to the sea-ice margin, snowfall increases might slow the retreat of glaciers and ice caps. Such areas include the periphery of Greenland, the Northern Canadian Arctic, the Russian Arctic, and Svalbard, which together hold around half of the sea-level potential locked in glaciers when considering the Northern Hemisphere (~16 cm)⁹⁰.

Step-wise Neoglacial re-growth and pre-Little Ice Age glacier maximum

As can be seen in Fig. 7, our EM 1-derived glacier indicator suggests Neoglacial regrowth of Berglibreen commenced sometime during the Late Holocene. While broadly consistent with other regional GICs²⁰, and tracking the millennial-scale decrease of summer insolation as is common in the higher latitudes of the Northern Hemisphere⁹¹, we observe step-wise changes superimposed on this gradual trend. To help attribute these tipping points, we identified them objectively using Change Point Detection (CPD) after Reeves et al.⁹².

Before discussing our Neoglacial CPD results (Fig. 7), we should elaborate on the sensitivity difference between our granulometric (EMMA) and geochemical (XRF-PCA) glacier indicators. As can be seen in Fig. 7c, d, the former suggests a gradual advance after 3.9 cal. ka BP, whereas the latter identifies rapid re-growth around 1.5 cal. ka BP. While similar offsets between these approaches have been observed in previous studies⁴⁹, XRF core scanning has been favoured in the field over the past decade. Indeed, all Holocene-length lacustrine sediment-based glacier reconstructions published on Svalbard since 2015 include XRF data in the interpretation^{34,87,93–96}. Methodologically, the analytical convenience and high resolution of XRF core scans are advantageous over laborious and comparatively coarse grain size sampling protocols⁹¹. And sedimentologically, it is oft-argued that glacial flour overprints the granulometric signature of other sediment sources, and renders glacier-fed lakes insensitive to variations in grain size⁹⁷. Although this might be true for some systems, or when assessing composite particle size distributions, our application of EMMA on glacier-fed lake sediments calls for caution. As can be seen in Fig. 7c, d, agreement between EM 1 and XRF-PCA input is restricted to unit 1 ($\rho = 0.5$, $p = 0.04$) where sedimentation is dominated by glacial flour input (see our Late glacial-Holocene lake evolution section in Results & Discussion). Finally, while it is beyond the purview of this study to attribute the observed difference in proxy sensitivity, we can preclude two likely candidates: 1) dilution – our XRF data has been scatter-normalized after Saunders et al.⁹⁸, and 2) matrix effects – centered log-ratios of included XRF element counts reveal identical trends (see Methods and Suppl. Figure 7)⁹⁹. Following from the above, we encourage future workers to combine established XRF scanning approaches with grain size EMMA to better detect diluted glacial flour signals.

As shown in Fig. 7d, CPD analysis shows that Berglibreen experienced three phases of glacier growth around 3.2, 1.5, and 0.8 cal. ka BP, following the onset of Neoglaciation around 3.9 cal. ka BP. This timing of regrowth is consistent with an observation-constrained isostatic model for the entire archipelago (Fig. 7b)^{29,81}, and other large GICs on Svalbard with a similar response time: notably large marine-terminating glaciers on the west coast (Fig. 7f)⁸⁴, and in northern Wijdefjorden⁴³. The timing of growth at ~3.2 cal. ka BP agrees well with the findings of Allaert et al.³⁴, who report that Longstaffbreen – a lower-lying (~500–700 m a.s.l.) northern outlet of Åsgardfonna – advanced into the Femmilsjøen catchment at this time (Fig. 1b). Across the fjord, Røthe et al.⁸⁷ report that Lake Vårflusjøen started receiving flour from glaciers with a similar altitude distribution ~3.5 cal. ka BP. Because glacial flour started to dominate sedimentation in Berglivatnet as Berglibreen expanded, subsequent advances around 1.5 and 0.8 cal. ka BP are captured by both XRF-PCA and grain size-EMMA glacier indicators (see Fig. 7c–d) – as outlined in the previous paragraph. Dated moraines, vegetation kill dates, as well as lacustrine sediment evidence show that the first step marked the advance of lower-altitude valley glaciers on the west coast (see

Fig. 7a)^{96,100–103}. As shown in the same panel, the final CPD-inferred advance around 0.8 cal. ka BP coincides with the reformation of coastal cirque glaciers on Svalbard^{93,94}. Following from the above, the identified growth phases mark a progressive lowering of the snowline that first triggered the expansion of high-altitude GICs and ended in the reformation of low-lying cirque glaciers. Glacier expansion might also help explain the observed increases in redox-sensitive Iron (Fe) and Manganese (Mn) input throughout unit 1 (see the Late glacial–Holocene lake evolution section in Results & Discussion and Fig. 6). As described by i.e. Ashley (1995)⁶³, Carrivick & Tweed⁶⁰ and Sabatier et al.⁵⁸, the increased inflow of heavy (cold and sediment-laden) meltwater perturbs thermal stratification in distal glacier-fed lakes by oxygenating bottom waters – a prerequisite for the precipitation of Fe and Mn. Finally, we note that both Principal Component Analysis (PCA) and End-Member Modelling Analysis (EMMA) evidence suggests that around 0.8 cal. ka BP, Berglivatnet experienced the most prominent period of growth observed during the late Holocene (see Fig. 7c, d). In doing so, our findings support a growing body of regional evidence that suggests the classical Little Ice age (LIA: 0.7–0.1 cal. ka BP) did not mark the culmination of Arctic North Atlantic Neoglaciation²⁰ and references therein^{95,104}.

Methods

Mapping

To map the bathymetry of the investigated basins and identify flat coring sites to minimize the risk of sediment disturbance, we surveyed Lakes Berglivatnet and Lakssjøen with a Lowrance Elite Ti2echo sounding system. The point data were exported and gridded at a 10×10 m resolution using Petrel v. 2020.6 in UTM zone 33 N (datum: WGS84). Prior to coring, the sedimentary infill of Lake Berglivatnet was investigated using a sub-bottom profiler to locate areas with the thickest package of undisturbed sediments for. For this purpose, we used an Edge Tech 3100-P CHIRP equipped with an SB-424 Tow Vehicle. A 4–16 kHz source sweep was transmitted at 10 pulses a second. This frequency range ensures a vertical resolution of ~8 cm. A sound velocity of 1500 m/s was used for depth conversion.

To identify topographic thresholds that restrict the inflow of meltwater to our lakes when ice retreats behind modern margins, we mapped the sub-glacial landscape of our study area (Fig. 2a). To do so, we relied on the globally modeled ice thickness distribution by Farinotti et al.⁹⁰, which was validated against the measurement-corrected local results by Fürst et al.¹⁰⁵. Using ESRI ArcGIS Pro version 2.9.5, we subtracted these independent ice thickness estimates from the high-resolution (2 m) Arctic digital elevation model (ArcticDEM) data for our study area¹⁰⁶, to obtain a sub-glacial topography model (Fig. 2a). We then applied the hydrology toolset in ArcGIS Pro to delineate sub-glacial catchments.

Coring

Relying on bathymetric and seismic surveys, six cores were collected from the two lakes during a field campaign in August 2021. To ensure retrieval of surface sediments that overlap with instrumental observations, we paired piston cores—extracted with a Nesje corer¹⁰⁷, with surface cores—taken with a UWITEC gravity coring system. In Berglivatnet, gravity core 602-21-10 (132.2 cm) and 175.3 cm long piston core 602-21-11 (Fig. 3d), were extracted from the western part of the basin at 56 m water depth (Fig. 3a–c). In Lakssjøen, the following four cores were retrieved from the south-western basin at 70 m water depth (Suppl. Figure 1b–c): gravity cores 602-21-1 (95.4 cm) and 602-21-3 (97.7 cm), as well as piston cores 602-21-2 (68.5 cm) and 602-21-4 (224.7 cm) (Suppl. Figure 1c).

Sedimentological analyses

Prior to analysis, all extracted cores were split lengthwise – identifying reference and working halves – and stored at 4 °C in the EARTHLAB facility at the University of Bergen. For the sake of sediment preservation, non-destructive (scanning) analyses were carried out first. To correlate the overlapping sections of gravity and piston cores, and pinpoint intervals dominated by allochthonous minerogenic input¹⁰⁸, we measured magnetic

susceptibility (MS) at 2 mm intervals with a Bartington MS2E sensor mounted on a GEOTEK multi-sensor core logger (MSC-L). To identify sub-millimeter scale changes in (glacigenic) minerogenic input, the elemental composition of the cores was mapped with an ITRAX XRF scanner from Cox Analytics¹⁰⁹. Scanning was performed at 200 µm intervals, using a molybdenum (Mo) tube with a voltage of 30 kV and a current of 30 µA, at 10 s dwell-time. Our choice for a Mo tube ensured the highest sensitivity for the detection of heavier lithogenic elements that characterize glacigenic input⁹³. We furthermore characterized sediment structure in 3-D using a ProCon-X-Ray CT-Alpha computed tomography (CT) scanner, set at 135 kV and 950 µA to generate 16-bit scans with a 53.1 µm resolution for lake Berglivatnet and 47.4 µm resolution for lake Lakssjøen. Processing, visualization, and analysis of the X-ray CT scans was performed with version 9.1.1 of the Thermo Scientific Avizo software. Scans were visualized in 3-D using the *Volume Rendering* module, the *SplineProbe* was used to map down-core variations in grayscale values at 200 µm intervals¹¹⁰, and we also relied on threshold-based segmentation with the *Colormap editor* to highlight particles or structures¹¹¹.

Composite records were generated for each lake using one set of gravity and piston cores. In Lakssjøen, where more than one set was retrieved, cores 602-21-3 and 602-21-4 were selected as they capture the full sedimentary sequence (Suppl. Figure 1a). To obtain a common depth scale, signal matching was performed using the QAnalySeries software version 1.5.1^{112,113}. Next, we measured physical parameters sequentially at 0.5–2 cm intervals on both composite records ($n = 247$ for 602-21-10/11, $n = 154$ for 602-21-3/4). Visibly laminated sediments were more closely sampled than more homogenous sections. First, we extracted 0.5 cm³ samples using a 2 mL syringe. Sediment was subsequently transferred, weighed, and dried overnight at 105 °C to measure Dry Bulk Density (DBD) and water content. The dried samples were then combusted at 550 °C for 4 h and cooled in a desiccator to determine organic content by measuring Loss on Ignition (LOI)¹¹⁴. Finally, we performed grain size analysis on the minerogenic residue using a Malvern Mastersizer 3000 with an LV Hydro dispersion unit¹¹⁵. For this purpose, each sample was measured in triplicate to warrant reproducibility.

Geochronology

Following wet sieving through a 250 and 125 µm mesh, 16 plant macrofossil samples were identified and submitted for Accelerator Mass Spectrometer (AMS) radiocarbon dating (Suppl. Table 1) either at the Tandem Laboratory at Uppsala University (Ua: 11 samples) or the Poznan Radiocarbon Laboratory (Poz: 5 samples). To avoid freshwater reservoir effects¹¹⁶, we picked terrestrial plant material where possible, and paired these with aquatic plants at one core depth to assess offsets (Suppl. Table 1, samples: Poz-157383, Poz-157384). To rigorously determine (changes in) sediment accumulation rates, our sampling strategy primarily focused on unit transitions.

Statistical analyses

We performed a range of statistical analyses to cluster, unmix, as well as transform data, and capture shared gradients of change or detect change points. To allow multivariate statistics, measured physical and scanning parameters were resampled to their highest (0.5 cm) common resolution using a moving average followed by regular interpolation in Past version 4.04¹¹⁷. To objectively divide our records into distinct units, visual assessment was supported by hierarchical clustering using the Rioja package version 1.0-5 in Rstudio 2023.03¹¹⁸. For this purpose, dissimilarity was calculated as Euclidean distance and expressed in total mean square values. To identify gradients of change shared between parameters, we performed a Principal Component Analysis (PCA) with the help of R Stats Package version 4.2.2. To avoid spurious correlations, all data were centered and standardized prior to analysis¹¹⁹. Following Blott & Pye (2011)¹²⁰, we relied on the GRADISTAT package to calculate grain size statistics using metric Folk and Ward measures. Furthermore, particle size distribution (PSD) data were unmixed using End-Member Modelling Analysis (EMMA), to help

disentangle the imprint of different depositional processes¹²¹. For this purpose, we followed the recommendations of van Hateren et al. (2017)¹²², using the non-parametric HALS-NMF algorithm in version 12 of the AnalySize software¹²³, which was run in Matlab R2021a. To account for closed-sum effects and dilution, compositional EMMA output was log-transformed following the recommendations of Bertrand et al.⁹⁹. To assess the offsets between different end-members, cross-correlation was performed using Qanalyses software, version 1.5.1^{112,113}. To identify key inflection points in Holocene seasonal insolation data¹²⁴, we calculated the first derivative of insolation for spring (MAM), summer (JJA), and autumn (SON) using the dplyr package¹²⁵ in RStudio 2023.03. Finally, to objectively identify phases of glacier growth (see the Step-wise Neoglacial re-growth section in Results & Discussion), we applied two Change Point Detection (CPD) algorithms⁹². First, the binseg method of the changepoint package in Rstudio 2023.03, specifying a maximum of 10 change points, and following the recommendations of van Hateren et al.^{122,126}. And secondly, the python (3.11.3) roufCP package, using a rough-fuzzy algorithm that is optimized for the detection of more gradual transitions after Bhaduri et al.¹²⁷. Change points were detected using the Kolmogorov-Smirnov test (kstest), after applying a 500-year moving average on the data, in light of chronological uncertainties.

Modelling input

To incorporate the modeled response of Åsgardfonna under RCP 8.5 forcing, the predicted glacier area for 2100 CE was extracted from the netCDF-4 dataset outputs provided by Rounce et al.³¹, using the ncdf4 package¹²⁸ in Rstudio 2023.03. To obtain spatially explicit results, this data was processed in ESRI ArcGIS Pro version 2.9.5, in conjunction with glacier outlines from version 7 of the Randolph Glacier Inventory¹²⁹, to calculate and plot the projected ice margin for 2100 CE under the RCP 8.5 scenario.

Data availability

The authors declare that all data produced for this study and shown in the manuscript figures, is provided in Supplementary Data and has been uploaded to the DataverseNO repository where the files can be accessed using the following DOI - <https://doi.org/10.18710/VDCUJE>.

Received: 19 February 2024; Accepted: 24 January 2025;

Published online: 11 February 2025

References

1. Serreze, M. C. & Barry, R. G. Processes and impacts of Arctic amplification: A research synthesis. *Glob. Planet. Change* **77**, 85–96 (2011).
2. Kopec, B. G., Feng, X., Michel, F. A. & Posmentier, E. S. Influence of sea ice on Arctic precipitation. *Proc. Natl. Acad. Sci. USA* **113**, 46–51 (2016).
3. England, M. R., Eisenman, I., Lutsko, N. J. & Wagner, T. J. W. The Recent Emergence of Arctic Amplification. *Geophys. Res. Lett.* **48**, e2021GL094086 (2021).
4. McCrystall, M. R., Stroeve, J., Serreze, M., Forbes, B. C. & Screen, J. A. New climate models reveal faster and larger increases in Arctic precipitation than previously projected. *Nat. Commun.* **12**, 6765 (2021).
5. Serreze, M. C., Barrett, A. P. & Stroeve, J. Recent changes in tropospheric water vapor over the Arctic as assessed from radiosondes and atmospheric reanalyses. *J. Geophys. Res. Atmos.* **117**, n/a–n/a (2012).
6. Bintanja, R. & Selten, F. M. Future increases in Arctic precipitation linked to local evaporation and sea-ice retreat. *Nature* **509**, 479–482 (2014).
7. Screen, J. A. & Simmonds, I. The central role of diminishing sea ice in recent Arctic temperature amplification. *Nature* **464**, 1334–1337 (2010).
8. Gardner, A. S. et al. A reconciled estimate of glacier contributions to sea level rise: 2003 to 2009. *science* **340**, 852–857 (2013).
9. Hugonnet, R. et al. Accelerated global glacier mass loss in the early twenty-first century. *Nature* **592**, 726–731 (2021).
10. Oerlemans, J. Extracting a Climate Signal from 169 Glacier Records. *Science* **308**, 675–677 (2005).
11. Pörtner, H.-O. et al. The ocean and cryosphere in a changing climate. *IPCC special report on the ocean and cryosphere in a changing climate* (2019).
12. Walsh, J. E., Kattsov, V., Portis, D. & Meleshko, V. Arctic precipitation and evaporation: Model results and observational estimates. *J. Clim.* **11**, 72–87 (1998).
13. Masson-Delmotte, V. et al. Climate change 2021: the physical science basis. *Contribution working group I sixth Assess. Rep. intergovernmental panel Clim. change* **2**, 2391 (2021).
14. Cai, Z. et al. Assessing Arctic wetting: Performances of CMIP6 models and projections of precipitation changes. *Atmos. Res.* **297**, 107124 (2024).
15. Lecavalier, B. S. et al. High Arctic Holocene temperature record from the Agassiz ice cap and Greenland ice sheet evolution. *Proc. Natl. Acad. Sci. USA* **114**, 5952–5957 (2017).
16. McFarlin, J. M. et al. Pronounced summer warming in northwest Greenland during the Holocene and Last Interglacial. *Proc. Natl. Acad. Sci. USA* **115**, 6357–6362 (2018).
17. Thomas, E. K. et al. A Wetter Arctic Coincident With Hemispheric Warming 8,000 Years Ago. *Geophys. Res. Lett.* **45**, (2018).
18. van der Bilt, W. G. M., D'Andrea, W. J., Werner, J. P. & Bakke, J. Early Holocene Temperature Oscillations Exceed Amplitude of Observed and Projected Warming in Svalbard Lakes. *Geophys. Res. Lett.* **46**, 14732–14741 (2019).
19. Briner, J. P., Axford, Y., Forman, S. L., Miller, G. H. & Wolfe, A. P. Multiple generations of interglacial lake sediment preserved beneath the Laurentide Ice Sheet. *Geology* **35**, 887–890 (2007).
20. Larocca, L. J. & Axford, Y. Arctic glaciers and ice caps through the Holocene: a circumpolar synthesis of lake-based reconstructions. *Climate* **18**, 579–606 (2022).
21. Karlén, W. Lacustrine Sediment Studies. *Geografiska Annaler: Ser. A, Phys. Geogr.* **63**, 273–281 (1981).
22. Hallet, B., Hunter, L. & Bogen, J. Rates of erosion and sediment evacuation by glaciers: A review of field data and their implications. *Glob. Planet. Change* **12**, 213–235 (1996).
23. Herman, F., De Doncker, F., Delaney, I., Prasicek, G. & Koppes, M. The impact of glaciers on mountain erosion. *Nat. Rev. Earth Environ.* **2**, 422–435 (2021).
24. Leemann, A. & Niessen, F. Holocene glacial activity and climatic variations in the Swiss Alps: reconstructing a continuous record from proglacial lake sediments. *Holocene* **4**, 259–268 (1994).
25. Briner, J. P., Stewart, H. A. M., Young, N. E., Philipps, W. & Losee, S. Using proglacial-threshold lakes to constrain fluctuations of the Jakobshavn Isbræ ice margin, western Greenland, during the Holocene. *Quat. Sci. Rev.* **29**, 3861–3874 (2010).
26. van der Bilt, W. G., Bakke, J., Vasskog, K., Røthe, T. & Støren, E. W. Glacier-fed lakes as palaeoenvironmental archives. *Geol. Today* **32**, 213–218 (2016).
27. Huang, J., Yu, H., Dai, A., Wei, Y. & Kang, L. Drylands face potential threat under 2 °C global warming target. *Nat. Clim. Change* **7**, 417–422 (2017).
28. Onarheim, I. H., Eldevik, T., Smedsrud, L. H. & Stroeve, J. C. Seasonal and Regional Manifestation of Arctic Sea Ice Loss. *J. Clim.* **31**, 4917–4932 (2018).
29. Fjeldskaar, W., Bondevik, S. & Amantov, A. Glaciers on Svalbard survived the Holocene thermal optimum. *Quat. Sci. Rev.* **199**, 18–29 (2018).
30. Geyman, E. C., JJ Van Pelt, W., Maloof, A. C., Aas, H. F. & Kohler, J. Historical glacier change on Svalbard predicts doubling of mass loss by 2100. *Nature* **601**, 374–379 (2022).

31. Rounce, D. R. et al. Global glacier change in the 21st century: Every increase in temperature matters. *Science* **379**, 78–83 (2023).
32. Luncke, B. Luftkartlegningen På Svalbard 1936. *Norsk Geografisk Tidsskrift Norwegian J. Geogr.* **6**, 145–154 (1936).
33. Kjellman, S. E. et al. Holocene precipitation seasonality in northern Svalbard: Influence of sea ice and regional ocean surface conditions. *Quat. Sci. Rev.* **240**, 106388 (2020).
34. Allaart, L. et al. Glacial history of the Åsgardfonna Ice Cap, NE Spitsbergen, since the last glaciation. *Quat. Sci. Rev.* **251**, 106717 (2021).
35. Salvigsen, O. Österholm H. Radiocarbon dated raised beaches and glacial history of the northern coast of Spitsbergen, Svalbard. *Polar Res.* **1982**, 97–115 (1982).
36. Farnsworth, W. R. et al. Persistence of Holocene ice cap in northeast Svalbard aided by glacio-isostatic rebound. *Quat. Sci. Rev.* **331**, 108625 (2024).
37. Hagen, J. O., Liestøl, O., Roland, E., & Jørgensen, T. *Glacier atlas of Svalbard and Jan Mayen*. Norsk Polarinstitutt Oslo (1993).
38. Bouchayer, C., Aiken, J. M., Thøgersen, K., Renard, F. & Schuler, T. V. A Machine Learning Framework to Automate the Classification of Surge-Type Glaciers in Svalbard. *J. Geophys. Res.: Earth Surf.* **127**, (2022).
39. Norwegian, P. I. Satellittbildmosaikk av Svalbard; Sentinel2 - 2020 (2023).
40. Martín-Moreno, R., Allende Álvarez, F. & Hagen, J. O. Little Ice Age' glacier extent and subsequent retreat in Svalbard archipelago. *Holocene* **27**, 1379–1390 (2017).
41. Allaart, L. et al. Late Quaternary glacier and sea-ice history of northern Wijdefjorden, Svalbard. *Boreas* **49**, 417–437 (2020).
42. Jang, K. et al. Glacial and environmental changes in northern Svalbard over the last 16.3 ka inferred from neodymium isotopes. *Glob. Planet. Change* **201**, 103483 (2021).
43. Jang, K. et al. Non-linear response of glacier melting to Holocene warming in Svalbard recorded by sedimentary iron (oxyhydr)oxides. *Earth Planet. Sci. Lett.* **607**, 118054 (2023).
44. Fujii, Y. et al. 6000-Year Climate Records in an Ice Core from the Høghetta Ice Dome in Northern Spitsbergen. *Ann. Glaciol.* **14**, 85–89 (1990).
45. Dowdeswell, J. A., Drewry, D. J. & Simões, J. C. Comments on: 6000-year climate records in an ice core from the Høghetta ice dome in northern Spitsbergen. *J. Glaciol.* **36**, 353–356 (1990).
46. Mangerud, J. & Svendsen, J. I. The Holocene Thermal Maximum around Svalbard, Arctic North Atlantic; molluscs show early and exceptional warmth. *Holocene* **28**, 65–83 (2017).
47. Dallmann, W. K. Geoscience atlas of Svalbard. (2015).
48. Witt-Nilsson, P., Gee, D. G. & Hellman, F. J. Tectonostratigraphy of the Caledonian Atomfjella Antiform of northern Ny Friesland, Svalbard. *Nor. Geologisk Tidsskr.* **78**, 67–80 (1998).
49. van der Bilt, W. G. M. et al. Novel sedimentological fingerprints link shifting depositional processes to Holocene climate transitions in East Greenland. *Glob. Planet. Change* **164**, 52–64 (2018).
50. Spielhagen, R. F. et al. Enhanced Modern Heat Transfer to the Arctic by Warm Atlantic Water. *Science* **331**, 450–453 (2011).
51. Onarheim, I. H., Smedsrud, L. H., Ingvaldsen, R. B. & Nilsen, F. Loss of sea ice during winter north of Svalbard. *Tellus A: Dyn. Meteorol. Oceanogr.* **66**, 23933 (2014).
52. Müller, J. et al. Holocene cooling culminates in sea ice oscillations in Fram Strait. *Quat. Sci. Rev.* **47**, 1–14 (2012).
53. Førland, E. J., Benestad, R., Hanssen-Bauer, I., Haugen, J. E. & Skaugen, T. E. Temperature and Precipitation Development at Svalbard 1900–2100. *Adv. Meteorol.* **2011**, 1–14 (2011).
54. Klimaservicesenter, N. Klimaprofil Ny-Ålesund (2023).
55. Schuler, T. V. & Østby, T. I. Sval_imp: a gridded forcing dataset for climate change impact research on Svalbard. *Earth Syst. Sci. Data* **12**, 875–885 (2020).
56. Reimer, P. J. et al. The IntCal20 Northern Hemisphere Radiocarbon Age Calibration Curve (0–55 cal kBP). *Radiocarbon* **62**, 725–757 (2020).
57. Blaauw, M. et al. Package 'rbacon' (2021).
58. Sabatier, P. et al. A Review of Event Deposits in Lake Sediments. *Quaternary* **5**, 34 (2022).
59. Millan, R., Mougnot, J., Rabatel, A. & Morlighem, M. Ice velocity and thickness of the world's glaciers. *Nat. Geosci.* **15**, 124–129 (2022).
60. Carrivick, J. L. & Tweed, F. S. Proglacial lakes: character, behaviour and geological importance. *Quat. Sci. Rev.* **78**, 34–52 (2013).
61. Davies, S. J., Lamb, H. F. & Roberts, S. J. Micro-XRF core scanning in palaeolimnology: recent developments. *Micro-XRF studies of sediment cores: Applications of a non-destructive tool for the environmental sciences*, 189–226 (2015).
62. Löwemark, L. et al. Normalizing XRF-scanner data: A cautionary note on the interpretation of high-resolution records from organic-rich lakes. *J. Asian Earth Sci.* **40**, 1250–1256 (2011).
63. Ashley, G. Glaciolacustrine environments. *Glacial Environ.* **1**, 417–444 (1995).
64. Trottier, A. P., Lajeunesse, P., Gagnon-Poiré, A. & Francus, P. Morphological signatures of deglaciation and postglacial sedimentary processes in a deep fjord-lake (Grand Lake, Labrador). *Earth Surf. Process. Landf.* **45**, 928–947 (2020).
65. van Loon, A. T., Pisarska-Jamroz, M., Nartišs, M., Krievāns, M. & Soms, J. Seismites resulting from high-frequency, high-magnitude earthquakes in Latvia caused by Late Glacial glacio-isostatic uplift. *J. Palaeogeogr.* **5**, 363–380 (2016).
66. Otto J.-C. Proglacial Lakes in High Mountain Environments. (Springer International Publishing, 2019).
67. Zhang, T. et al. Warming-driven erosion and sediment transport in cold regions. *Nat. Rev. Earth Environ.* **3**, 832–851 (2022).
68. Ottesen, R. et al. Geochemical atlas of Norway, Part 2: Geochemical Atlas of Spitsbergen. *Geological Survey Norway* (NGU, 2010).
69. Hormes, A., Gjermundsen, E. F. & Rasmussen, T. L. From mountain top to the deep sea–Deglaciation in 4D of the northwestern Barents Sea ice sheet. *Quat. Sci. Rev.* **75**, 78–99 (2013).
70. Farnsworth, W. R. et al. Vedde Ash constrains Younger Dryas glacier re-advance and rapid glacio-isostatic rebound on Svalbard. *Quat. Sci. Adv.* **5**, 100041 (2022).
71. Voldstad, L. H. et al. A complete Holocene lake sediment ancient DNA record reveals long-standing high Arctic plant diversity hotspot in northern Svalbard. *Quat. Sci. Rev.* **234**, 106207 (2020).
72. Bromley, G. et al. Lateglacial Shifts in Seasonality Reconcile Conflicting North Atlantic Temperature Signals. *J. Geophys. Res.: Earth Surface* **128**, e2022JF006951 (2023).
73. Schenk, F. et al. Warm summers during the Younger Dryas cold reversal. *Nat. Commun.* **9**, 1634 (2018).
74. Oswald, P., Strasser, M., Hammerl, C. & Moernaut, J. Seismic control of large prehistoric rockslides in the Eastern Alps. *Nat. Commun.* **12**, 1059 (2021).
75. Forman, S. et al. A review of postglacial emergence on Svalbard, Franz Josef Land and Novaya Zemlya, northern Eurasia. *Quat. Sci. Rev.* **23**, 1391–1434 (2004).
76. Koehl, J.-B. & Allaart, L. The Billefjorden Fault Zone north of Spitsbergen: a major terrane boundary? *Polar Res.* **40**, 7668 (2021).
77. Piepjohn, K., Dallmann, W. K. & Elvevold, S. The Lomfjorden fault zone in eastern Spitsbergen (Svalbard). In *Circum-Arctic Structural Events: Tectonic Evolution of the Arctic Margins and Trans-Arctic Links with Adjacent Orogens* (eds Piepjohn, K., Strauss, J. V., Reinhardt, L. & McClelland, W. C.) Vol. 541, <https://doi.org/10.1130/SPE541> (2019).
78. Ottemöller, L., Kim, W.-Y., Waldhauser, F., Tjåland, N. & Dallmann, W. The Storfjorden, Svalbard, Earthquake Sequence 2008–2020: Transtensional Tectonics in an Arctic Intraplate Region. *Seismological Res. Lett.* **92**, 2838–2849 (2021).

79. Risebrobakken, B. et al. Early Holocene temperature variability in the Nordic Seas: The role of oceanic heat advection versus changes in orbital forcing. *Paleoceanography* **26**, PA4206 (2011).
80. Pierikowski, A. J. et al. Seasonal sea ice persisted through the Holocene Thermal Maximum at 80 N. *Commun. Earth Environ.* **2**, 1–10 (2021).
81. Farnsworth, W. R. et al. Holocene glacial history of Svalbard: Status, perspectives and challenges. *Earth-Sci. Rev.* **208**, 103249 (2020).
82. Raper, S. C. & Braithwaite, R. J. Glacier volume response time and its links to climate and topography based on a conceptual model of glacier hypsometry. *Cryosphere* **3**, 183–194 (2009).
83. Aitchison J. The statistical analysis of compositional data. (1986).
84. Werner, K. et al. Holocene sea subsurface and surface water masses in the Fram Strait—Comparisons of temperature and sea-ice reconstructions. *Quat. Sci. Rev.* **147**, 194–209 (2016).
85. Linderholm, H. W. et al. Arctic hydroclimate variability during the last 2000 years: current understanding and research challenges. *Climate* **14**, 473–514 (2018).
86. McFarlin, J. M., Axford, Y., Masterson, A. L. & Osburn, M. R. Calibration of modern sedimentary $\delta^2\text{H}$ plant wax-water relationships in Greenland lakes. *Quat. Sci. Rev.* **225**, 105978 (2019).
87. Røthe, T. O., Bakke, J., Støren, E. W. & Bradley, R. S. Reconstructing Holocene glacier and climate fluctuations from lake sediments in Vårfluesjøen, Northern Spitsbergen. *Front. Earth Sci.* **6**, 91 (2018).
88. Braithwaite, R. J. From Doktor Kurowski's Schneegrenze to our modern glacier equilibrium line altitude (ELA). *Cryosphere* **9**, 2135–2148 (2015).
89. Hanssen-Bauer, I. et al. Climate in Svalbard 2100. In: *A knowledge base for climate adaptation*. Norwegian CCS Research Centre (NCCS) (2019).
90. Farinotti, D. et al. A consensus estimate for the ice thickness distribution of all glaciers on Earth. *Nat. Geosci.* **12**, 168–173 (2019).
91. Solomina, O. N. et al. Holocene glacier fluctuations. *Quat. Sci. Rev.* **111**, 9–34 (2015).
92. Reeves, J., Chen, J., Wang, X. L., Lund, R. & Lu, Q. Q. A Review and Comparison of Changepoint Detection Techniques for Climate Data. *J. Appl. Meteorol. Climatol.* **46**, 900–915 (2007).
93. van der Bilt, W. G. M. et al. Reconstruction of glacier variability from lake sediments reveals dynamic Holocene climate in Svalbard. *Quat. Sci. Rev.* **126**, 201–218 (2015).
94. de Wet, G. A. et al. Holocene glacier activity reconstructed from proglacial lake Gjøvatnet on Amsterdamøya, NW Svalbard. *Quat. Sci. Rev.* **183**, 188 (2018).
95. van der Bilt, W. G., Born, A. & Haaga, K. A. Was Common Era glacier expansion in the Arctic Atlantic region triggered by unforced atmospheric cooling? *Quat. Sci. Rev.* **222**, 105860 (2019).
96. Røthe, T. O. et al. Arctic Holocene glacier fluctuations reconstructed from lake sediments at Mitrahålvøya, Spitsbergen. *Quat. Sci. Rev.* **109**, 111–125 (2015).
97. Bakke, J. et al. A complete record of Holocene glacier variability at Austre Okstindbreen, northern Norway: an integrated approach. *Quat. Sci. Rev.* **29**, 1246–1262 (2010).
98. Saunders, K. M. et al. Holocene dynamics of the Southern Hemisphere westerly winds and possible links to CO₂ outgassing. *Nat. Geosci.* **11**, 650–655 (2018).
99. Bertrand S. et al. Inorganic geochemistry of lake sediments: A review of analytical techniques and guidelines for data interpretation. *Earth-Sci. Rev.* 104639 (2023).
100. Svendsen, J. I. & Mangerud, J. Holocene glacial and climatic variations on Spitsbergen, Svalbard. *Holocene* **7**, 45–57 (1997).
101. Miller, G. H., Landvik, J. Y., Lehman, S. J. & Southon, J. R. Episodic Neoglacial snowline descent and glacier expansion on Svalbard reconstructed from the 14 C ages of ice-entombed plants. *Quat. Sci. Rev.* **155**, 67–78 (2017).
102. Reusche, M. et al. 10Be surface exposure ages on the late-Pleistocene and Holocene history of Linnébreen on Svalbard. *Quat. Sci. Rev.* **89**, 5–12 (2014).
103. Philipps, W. et al. Late Holocene glacier activity at inner Hornsund and Scottbreen, southern Svalbard. *J. Quat. Sci.* **32**, 501–515 (2017).
104. Kjær, K. H. et al. Glacier response to the Little Ice Age during the Neoglacial cooling in Greenland. *Earth-Sci. Rev.* **227**, 103984 (2022).
105. Fürst, J. J. et al. The Ice-Free Topography of Svalbard. *Geophys. Res. Lett.* **45**, 11,760–11,769 (2018).
106. Porter, C. et al. Arctic DEM. *Harvard Dataverse* **1**, (2018).
107. Nesje, A. A piston corer for lacustrine and marine sediments. *Arct. Alp. Res.* **24**, 257–259 (1992).
108. Sandgren P. & Snowball I. Application of mineral magnetic techniques to paleolimnology. *Tracking environmental change using lake sediments: Physical and geochemical methods*, 217–237 (2001).
109. Croudace, I. W., Rindby, A. & Rothwell, R. G. ITRAX: description and evaluation of a new multi-function X-ray core scanner. *Geol. Soc., Lond., Spec. Publ.* **267**, 51–63 (2006).
110. van der Bilt, W. G. M., Cederström, J. M., Støren, E. W. N., Berben, S. M. P. & Rutledal, S. Rapid Tephra Identification in Geological Archives With Computed Tomography: Experimental Results and Natural Applications. *Front. Earth Sci.* **8**, 622386 (2021).
111. van der Bilt, W. G. & Lane, C. S. Lake sediments with Azorean tephra reveal ice-free conditions on coastal northwest Spitsbergen during the Last Glacial Maximum. *Sci. Adv.* **5**, eaaw5980 (2019).
112. Paillard, D., Labeyrie, L. & Yiou, P. Macintosh program performs time-series analysis. *Eos, Trans. Am. Geophys. Union* **77**, 379–379 (1996).
113. Kotov, S. & Pälke, H. QAnalyzeSeries—a cross-platform time series tuning and analysis tool. In: *AGU Fall Meeting Abstracts* (2018).
114. Dean, W. E. Determination of carbonate and organic matter in calcareous sediments and sedimentary rocks by loss on ignition; comparison with other methods. *J. Sediment. Res.* **44**, 242–248 (1974).
115. van der Bilt, W. G. M. et al. Late Holocene canyon-carving floods in northern Iceland were smaller than previously reported. *Commun. Earth Environ.* **2**, 86 (2021).
116. Philippsen, B. The freshwater reservoir effect in radiocarbon dating. *Herit. Sci.* **1**, 24 (2013).
117. Hammer, O. PAST: Paleontological statistics software package for education and data analysis. *Palaeontol. electron* **4**, 9 (2001).
118. Juggins, S. & Juggins, M. S. Package 'rioja'. RCRAN (2019).
119. Šmilauer, P. & Lepš, J. *Multivariate analysis of ecological data using CANOCO 5*. Cambridge University Press (2014).
120. Blott, S. J. & Pye, K. GRADISTAT: a grain size distribution and statistics package for the analysis of unconsolidated sediments. *Earth Surf. Process. Landf.* **26**, 1237–1248 (2001).
121. Prins, M. A. & Weltje G. J. End-member modeling of siliciclastic grain-size distributions: the late Quaternary record of eolian and fluvial sediment supply to the Arabian Sea and its paleoclimatic significance. (1999).
122. van Hateren, J., Prins, M. & Van Balen, R. On the genetically meaningful decomposition of grain-size distributions: A comparison of different end-member modelling algorithms. *Sediment. Geol.* **375**, 49–71 (2018).
123. Paterson, G. A. & Heslop, D. New methods for unmixing sediment grain size data. *Geochem. Geophys. Geosyst.* **16**, 4494–4506 (2015).
124. Laskar, J. et al. A long-term numerical solution for the insolation quantities of the Earth. *Astron. Astrophys.* **428**, 261–285 (2004).
125. Wickham, H., Francois, R., Henry, L. & Müller, K. dplyr. *A Grammar of Data Manipulation 2020 [Last accessed on 2020 Aug 12]* Available from, Rproject (2014).
126. Killick, R. & Eckley, I. changepoint: An R package for changepoint analysis. *J. Stat. Softw.* **58**, 1–19 (2014).

127. Bhaduri, R., Roy, S. & Pal, S. K. Rough-Fuzzy CPD: a gradual change point detection algorithm. *J. Data, Inf. Manag.* **4**, 243–266 (2022).
128. Pierce D. & Pierce M. D. Package ‘ncdf4’ (2019).
129. Pfeffer, W. T. et al. The Randolph Glacier Inventory: a globally complete inventory of glaciers. *J. Glaciol.* **60**, 537–552 (2014).
130. Fetterer, F., Knowles, K., Meier, W. N., Savoie, M. & Windnagel, A. K. Sea Ice Index (G02135, Version 3). (ed Center NSaID) (2017).
131. Marchal, O. et al. Apparent long-term cooling of the sea surface in the northeast Atlantic and Mediterranean during the Holocene. *Quat. Sci. Rev.* **21**, 455–483 (2002).
132. Jakobsson M. International bathymetric chart of the Arctic Ocean (IBCAO). *Encycl. Earth Sci.* 365–367 (2016).
133. Müller, J. et al. Towards quantitative sea ice reconstructions in the northern North Atlantic: A combined biomarker and numerical modelling approach. *Earth Planet. Sci. Lett.* **306**, 137–148 (2011).
134. van der Bilt, W. G. M. et al. Alkenone-based reconstructions reveal four-phase Holocene temperature evolution for High Arctic Svalbard. *Quat. Sci. Rev.* **183**, 204–213 (2018).

Acknowledgements

This research was funded by a Starting Grant (TMS2021STG01) from the Trond Mohn Research Foundation (TMF), and also received support from the Norwegian Research Council (RCN) via the ExPal project (grant nr. 337291). We thank Sysselimesteren for allowing us to collect our samples under permit 11689 in the Research in Svalbard (RiS) database. We also thank the three reviewers for providing constructive feedback that helped improve the manuscript. Finally, we express our gratitude to Gaetane Demyttenaere and Ali Bolkar Kiliç for their help with labwork.

Author contributions

W.v.d.B designed the study and obtained funding for it. The applied methodology was developed by W.v.d.B and analyses were performed by A.G.A. Bathymetry and CHIRP data were acquired and pre-processed by J.B, E.S and J.M.C. Catchment samples and cores were collected by W.v.d.B, A.S, J.B, E.S, J.M.B, J.M.C and S.v.d.P. The original draft was written by A.G.A and W.v.d.B, and all authors contributed to the final manuscript through multiple rounds of editing.

Funding

Open access funding provided by University of Bergen (incl Haukeland University Hospital).

Competing interests

The authors declare no competing interests.

Additional information

Supplementary information The online version contains supplementary material available at <https://doi.org/10.1038/s43247-025-02064-z>.

Correspondence and requests for materials should be addressed to Andreea Gabriela Auer.

Peer review information *Communications Earth & Environment* thanks Jan Kavan and the other, anonymous, reviewer(s) for their contribution to the peer review of this work. Primary Handling Editors: Rachael Rhodes and Carolina Ortiz Guerrero. A peer review file is available.

Reprints and permissions information is available at <http://www.nature.com/reprints>

Publisher's note Springer Nature remains neutral with regard to jurisdictional claims in published maps and institutional affiliations.

Open Access This article is licensed under a Creative Commons Attribution 4.0 International License, which permits use, sharing, adaptation, distribution and reproduction in any medium or format, as long as you give appropriate credit to the original author(s) and the source, provide a link to the Creative Commons licence, and indicate if changes were made. The images or other third party material in this article are included in the article's Creative Commons licence, unless indicated otherwise in a credit line to the material. If material is not included in the article's Creative Commons licence and your intended use is not permitted by statutory regulation or exceeds the permitted use, you will need to obtain permission directly from the copyright holder. To view a copy of this licence, visit <http://creativecommons.org/licenses/by/4.0/>.

© The Author(s) 2025