



Coherent regional expressions of Holocene summer temperature on Greenland

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15 **Abstract.** Proxy time series suggest that Holocene summer warmth peaked asynchronously across Greenland, but the extent to which these differences reflect coherent regional climate patterns remains unclear. We compile 38 quantitative and qualitative summer-temperature proxy time series from Greenland and adjacent seas and apply an uncertainty-weighted factor analysis framework that accommodates heterogeneous proxy types to identify coherent temporal and spatial patterns from 10 ka to present. We compare these proxy-derived factors with those for five proxy-informed paleoclimate reconstructions and a transient model-only baseline. Two factors best represent the proxy network. The leading temporal mode reaches a maximum by 7.2 ka, reflecting early-to-mid Holocene warmth in northern and eastern Greenland, and contrasts with delayed warming in western Greenland. We interpret this pattern as an insolation-driven summer-temperature response modified by meltwater, ocean stratification, and sea-ice dynamics during retreat of the Laurentide and Greenland ice sheets. The second temporal mode peaks by 5.4 ka and is strongest in southern and southwestern Greenland, which may reflect an increased coastal influence of Atlantic-origin heat transported by the Irminger and West Greenland currents. Climate reconstructions reproduce the temporal evolution of the proxy-derived factors more successfully than their spatial patterns. Two paleoclimate reconstructions provide the closest agreement, but no existing reconstruction reproduces the complete regional structure of the proxy network. Despite proxy and chronological uncertainties, this analysis demonstrates that heterogeneous proxies can resolve coherent spatiotemporal patterns in Holocene Greenland climate, providing a basis for mechanistic interpretation and an empirical benchmark for evaluating paleoclimate reconstructions.

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1 Introduction

Rapid high-latitude warming threatens the stability of the Greenland Ice Sheet (GrIS), a major contributor to global sea-level rise (Box et al., 2022). Summer air temperature governs ice-sheet melt (Shepherd et al., 2020; Trusel et al., 2018), which in turn impacts GrIS surface mass balance, making summer climate a critical variable for past and future GrIS sensitivity. Developing our understanding of this sensitivity using direct climatic measurements is inhibited by the limited duration of the instrumental record. Paleoclimate reconstructions are therefore essential for evaluating GrIS responses to external forcing and internal variability (Dahl-Jensen et al., 2013).

The Holocene is a useful reference period for investigating Greenland climate and GrIS sensitivity because it spans large changes in orbital forcing (Laskar et al., 2004), regional oceanographic conditions (e.g., Hansen et al., 2020; Jennings et al., 2011; Ouellet-Bernier et al., 2014; Perner et al., 2013), sea-ice extent (e.g., Funder et al., 2011; Polyak et al., 2010; Saini et al., 2020), meltwater fluxes (e.g., Allan et al., 2021; Hansen et al., 2020; Lochte et al., 2019), and ice-sheet geometry (e.g., Larsen et al., 2015; Leger et al., 2024; Walcott-George et al., 2026; Young and Briner, 2015). These changes produced substantial summer-temperature variability across Greenland. In northern and eastern Greenland, proxy-derived climate reconstructions indicate strong early Holocene warmth, with summer temperatures in some regions reaching 3–5 °C above than the mid-twentieth century mean (e.g., Axford et al., 2021). In contrast, western and southern Greenland, as well as the Baffin Bay region, experienced more complex climate histories with delayed or broadened peak Holocene warmth (Axford et al., 2021; Briner et al., 2016; Martin et al., 2024). These regional differences may reflect contrasts in insolation sensitivity, sea-ice configuration, ocean circulation and meltwater responses, as well as proxy-specific biases. Interpreting these time series collectively remains challenging. Proxy time series provide valuable constraints on climatic and environmental conditions, but when considered individually, it is difficult to determine whether a given time series reflects isolated local conditions or a coherent spatiotemporal factor of Greenland-wide climate variability.

Distinguishing local and regionally coherent signals is important for characterizing spatial patterns of climate change across Greenland and for evaluating GrIS sensitivity with ice-sheet models (e.g., Briner et al., 2020; Cuzzone et al., 2026; Tabone et al., 2024), which require spatially complete climate fields. Gridded paleoclimate reconstructions, including data-assimilation (DA) products (Badgeley et al., 2020; Erb et al., 2022; Hakim et al., 2016; Osman et al., 2021) and related proxy-informed model reconstructions (e.g., Buizert et al., 2018) help address this need by merging climate-model output with proxy-derived climate estimates to produce physically consistent fields through time. Despite the utility of these paleoclimate reconstructions, they vary substantially in the proxy information and model assumptions used to generate them, producing wide-ranging differences in paleoclimate reconstruction outputs (e.g., Fig. 1). For example, some climate reconstruction products focused on Greenland incorporate ice-core records from the ice-sheet interior only (Badgeley et al., 2020; Buizert et al., 2018), while a global reconstruction product incorporates only proxy estimates of sea surface temperatures (Osman et al., 2021). These decisions are important because different proxies and archives record different environments and seasons, and because interior ice-sheet temperature, offshore marine temperature, and summer temperature near the ice-sheet margin are



not interchangeable climate targets. Temperatures along the GrIS margin are especially important for understanding GrIS sensitivity because they are directly linked to surface melt, ice-margin retreat, and the boundary conditions experienced by land-terminating and marine-terminating glaciers.

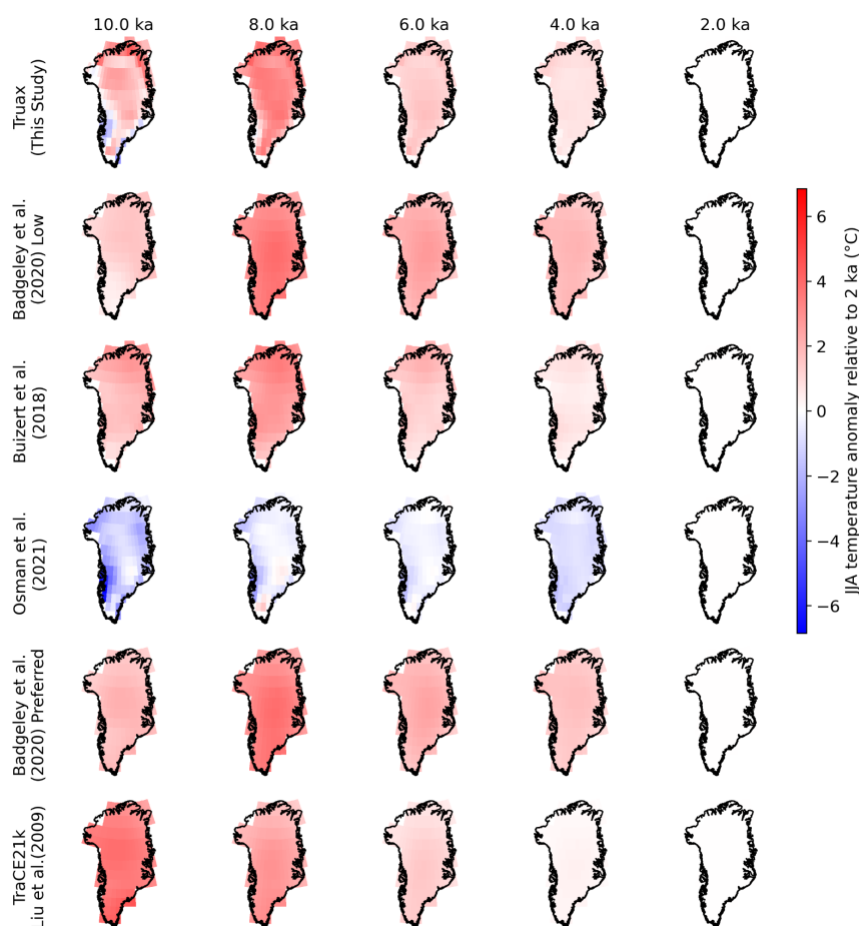


Fig. 1. Summer (June-August) temperature anomalies for climate reconstruction products (rows: Liu et al., 2009; Buizert et al., 2018; Badgeley et al., 2020; Osman et al., 2021; Truax, This Study). Anomalies are calculated at each grid cell relative to the same product's temperature at

Here, we evaluate the agreement between proxies and multiple Holocene climate reconstructions by targeting dominant structures of Greenland summer-temperature variability. We compile a network of published summer-temperature proxy reconstructions from Greenland and apply factor analysis to identify the dominant temporal modes and spatial loading patterns of Holocene summer-temperature variability. We then use the proxy-derived factor structure as a benchmark for evaluating paleoclimate reconstructions, including data-assimilation products, proxy-informed model reconstructions, and a transient model-only baseline.

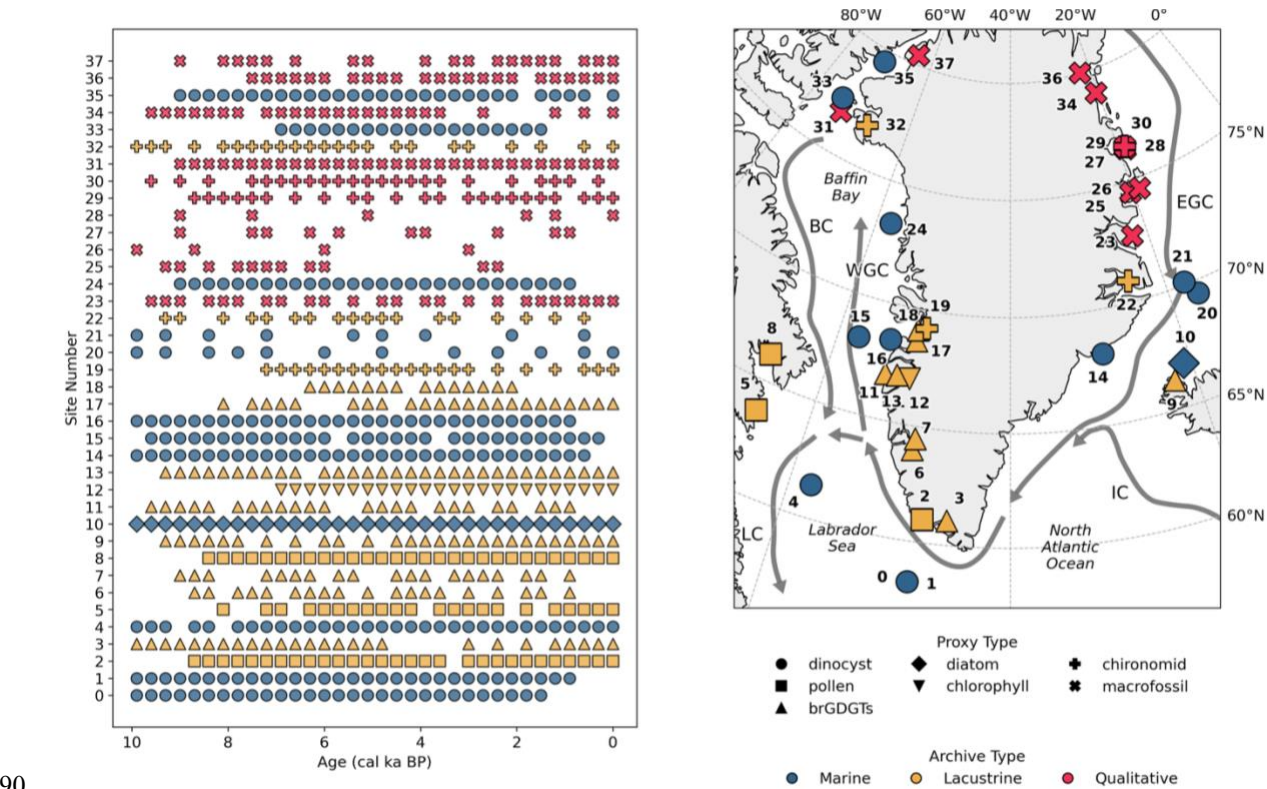


This comparison serves as both an evaluation framework and a tool to identify robust features of Greenland’s Holocene climate history. Shared patterns between the independent proxy network and the reconstructions provide high-
80 confidence interpretations, while discrepancies highlight regions and modes of variability that require further research.

2. Methods

2.1 Proxy Time Series

We identify lacustrine and marine proxy time series primarily from previously published studies and compilations (Fig. 2; Table 1). Time series selection is based on established proxy–temperature relationships, seasonal sensitivity, and temporal resolution. Because our objective is to reconstruct regional patterns of summer climate variability, only proxies with
85 demonstrated sensitivity to summer temperature are included. The marine time series we use are based on dinoflagellate cyst and diatom proxies; terrestrial time series are based on brGDGT, pollen, chlorophyll, and chironomid proxies (Table 1). We use published uncertainties related to proxy–temperature relationships (Table 2). Each time series is required to span at least 4 kyr, with at minimum of 1 observation per 2 kyr.



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Fig. 2. Observations within the 38 proxy time series after binning (300-year intervals), differentiated by proxy type (marker shape) and archive (colour; marine (blue), terrestrial (yellow), and qualitative terrestrial (red)). Time series are indexed (y-axis) to Table 1. Grey arrows indicate ocean circulation: BC, Baffin Current; LC, Labrador Current; WGC, West Greenland Current; IC, Irminger Current; and EGC, East Greenland Current.



95 To address a critical spatial gap in available quantitative time series, we also include qualitative data from temperature-sensitive proxies from northern Greenland (Fig. 2). These include nine macrofossil-based time series (Bennike, 2002; Bennike et al., 2008; Bennike and Wagner, 2012; Bennike and Weidick, 2001; Christiansen et al., 2002; Klug et al., 2009; Schmidt et al., 2011; Wagner et al., 2000, 2008; Wagner and Bennike, 2015) and two chironomid time series (Schmidt et al., 2011) from lake and marine sediment archives.

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Table 1. Metadata for proxy time series included in the Greenland Holocene proxy network. Listed information includes site name, alternate name (where applicable), geographic region, latitude, longitude, archive type, proxy type, data class (quantitative, Qn, or qualitative, Ql), and references.

Index	Site Name	Region	Latitude	Longitude	Archive	Proxy	Class	Reference
0	HU90-013-013	Labrador Sea	58.210	-48.373	Marine	dinocyst	Qn	(de Vernal et al., 2013; de Vernal and Hillaire-Marcel, 2006; Solignac et al., 2004)
1	MD99-2227	Labrador Sea	58.210	-48.370	Marine	dinocyst	Qn	(de Vernal and Hillaire-Marcel, 2006)
2	Lake Qipisarqo	Baffin Island	61.011	-47.754	Lacustrine	pollen	Qn	(Fréchette and de Vernal, 2009)
3	Lake 578	South Greenland	61.080	-45.610	Lacustrine	brGDGTs	Qn	(Schneider et al., 2024)
4	HU2008-029-004TWC	Northwest Labrador Sea	61.464	-58.036	Marine	dinocyst	Qn	(Gibb et al., 2015)
5	Lake Jake	Baffin Island	63.670	-65.150	Lacustrine	pollen	Qn	(Miller et al., 2005)
6	Lake Bullet	West Greenland	63.982	-49.537	Lacustrine	brGDGTs	Qn	(Acharya et al., 2026)
7	Lake Marshall	West Greenland	64.464	-49.431	Lacustrine	brGDGTs	Qn	(Acharya et al., 2026)
8	Lake Iglutalik	Baffin Island	66.140	-66.080	Lacustrine	pollen	Qn	(Kerwin et al., 2004)
9	Lake Skorarvatn	Iceland	66.256	-22.322	Lacustrine	brGDGTs	Qn	(Harning et al., 2020)



10	MD99-2269	Western North Atlantic	66.850	-20.850	Marine	diatom	Qn	(Justwan et al., 2008; Solignac et al., 2006)
11	Lake Rosaea	West Greenland	66.982	-53.718	Lacustrine	brGDGTs	Qn	(Acharya et al., 2026)
12	Lake Braya Sø	Southwest Greenland	66.988	-51.033	Lacustrine	chlorophyll	Qn	(von Gunten et al., 2012)
13	Lake Gus	West Greenland	67.032	-52.427	Lacustrine	brGDGTs	Qn	(Cluett et al., 2023)
14	JM96-1207	Greenland Sea	68.100	-29.350	Marine	dinocyst	Qn	(Solignac et al., 2006)
15	HU2008-029-070	Baffin Bay	68.230	-57.620	Marine	dinocyst	Qn	(Gibb et al., 2015)
16	MSM343300	Disko Bugt	68.472	-54.002	Marine	dinocyst	Qn	(Ouellet-Bernier et al., 2014)
17	Lake N3	West Greenland	68.636	-50.981	Lacustrine	brGDGTs	Qn	(Acharya et al., 2026)
18	Lake Pluto	West Greenland	69.110	-51.032	Lacustrine	brGDGTs	Qn	(Acharya et al., 2026)
19	North Lake	West Greenland	69.242	-50.027	Lacustrine	chironomid	Qn	(Axford et al., 2013)
20	PS1842	Greenland Sea	69.460	-16.530	Marine	dinocyst	Qn	(Van Nieuwenhove et al., 2016)
21	PS1730-1/2	Greenland Sea	70.120	-17.690	Marine	dinocyst	Qn	(Matthiessen and Baumann, 1997; Van Nieuwenhove et al., 2016)
22	Last Chance Lake	East Greenland	70.906	-24.431	Lacustrine	chironomid	Qn	(Axford et al., 2017)
23	Basalt Sø	East Greenland	72.719	-22.225	Lacustrine	macrofossil	Ql	(Wagner et al., 2000)
24	AMD14-204	Northwest Greenland	73.260	-57.900	Marine	dinocyst	Qn	(Caron et al., 2019)



25	Zackenbergt Delta	East Greenland	74.500	-20.500	Lacustrine	macrofossil	Ql	(Christiansen et al., 2002)
26	Germania Havn Sø	Northeast Greenland	74.548	-18.859	Lacustrine	macrofossil	Ql	(Bennike and Wagner, 2012)
27	Hjort Sø	Northeast Greenland	76.333	-18.808	Lacustrine	macrofossil	Ql	(Wagner et al., 2008)
28	Store Koldewey	Northeast Greenland	76.417	-18.750	Lacustrine	macrofossil	Ql	(Klug et al., 2009)
29	Duck Lake	Northeast Greenland	76.417	-18.750	Lacustrine	chironomid	Ql	(Schmidt et al., 2011)
30	Hjort Lake	Northeast Greenland	76.433	-18.767	Lacustrine	chironomid	Ql	(Schmidt et al., 2011)
31	Carey Øer	Northwest Greenland	76.738	-73.208	Lacustrine	macrofossil	Ql	(Bennike et al., 2008)
32	Deltasø	Northwest Greenland	76.760	-67.610	Lacustrine	chironomid	Qn	(Axford et al., 2019)
33	HU91-039-008P+- 007BC	Nares Strait	77.270	-74.330	Marine	dinocyst	Qn	(Levac et al., 2001)
34	Nioghalvfjærdsfjorden and Jøkelbugten	Northeast Greenland	79.000	-20.500	Lacustrine	macrofossil	Ql	(Bennike and Weidick, 2001)
35	AMD14-Kane2B	Nares Strait	79.510	-70.830	Marine	dinocyst	Qn	(Caron et al., 2019)
36	Lille Sneha Sø	Northeast Greenland	80.016	-22.650	Lacustrine	macrofossil	Ql	(Wagner and Bennike, 2015)
37	Washington Land	Northwest Greenland	80.400	-64.000	Lacustrine	macrofossil	Ql	(Bennike 2002)



105 We incorporate the qualitative time series into a unified analytical framework with the quantitative time series as follows. For each site, we create a time series using age constraints and age-depth models reported in the relevant original publication and, for each age, assign the qualitative temperature interpretation given in that publication. The qualitative data are grouped into four ordinal categories: *coldest*, *cold*, *warm*, and *warmest*. The *coldest* and *warmest* categories represent the lowest and highest temperatures inferred for a given site, as described in the original studies. To enable quantitative comparison
 110 with continuous proxy records, the ordinal categories are mapped onto a unitless scale (-1.5, -0.5, 0.5, and 1.5), preserving equal spacing between categories. This transformation allows qualitative records to be incorporated into subsequent analyses while maintaining consistency with quantitative temperature reconstructions.

2.2 Climate Reconstruction Products

We compare our proxy network against five paleoclimate reconstructions that merge networks of paleoclimate records with
 115 climate-model output to generate spatially complete estimates of Greenland temperature evolution during the Holocene. Four of these five products employ an offline ensemble Kalman filter (EnKF) paleoclimate data-assimilation framework, in which an ensemble of prior climate states drawn from model simulations is updated by a network of proxy observations (Hakim et al., 2016). The fifth product, Buizert et al. (2018), uses a related but distinct proxy-informed model merging approach.

We used two products from Badgeley et al. (2020), which reconstructed Greenland mean-annual temperature over
 120 the past 20,000 years by assimilating $\delta^{18}\text{O}$ records from eight Greenland ice cores, using the TraCE-21k simulation as the prior. TraCE-21k (Liu et al., 2009) is a transient climate model simulation, at relatively coarse spatial resolution ($\sim 3.75^\circ$) that used Community Climate System Model 3 (CCSM3) from 21 ka to present. Because TraCE-21k did not include water isotopes, a simple linear proxy-system model was employed to predict the expected $\delta^{18}\text{O}$ value at each ice-core site given the simulated temperature from the model prior. The data-assimilation procedure spreads proxy information across the spatial domain (in
 125 this case, Greenland and surrounding seas) through the prior's covariance structure, producing gridded reconstructions of past surface temperature. Badgeley et al. (2020) also reconstructed precipitation by assimilating accumulation time series from five ice cores. Uncertainty in the ice-flow corrections used to derive accumulation was represented by low-, moderate-, and high-accumulation scenarios, with the low and high scenarios providing sensitivity bounds. Here, we consider the low (Badgeley Low) and high (Badgeley Preferred) boundaries separately. Following previous work (e.g., Hakim et al., 2016), Badgeley et
 130 al. (2020) used a time-agnostic prior, drawing model states without evolving the prior ensemble through time.

Osman et al. (2021) developed a global seasonally resolved surface-temperature reconstruction, known as the Last Glacial Maximum Reanalysis (LGMR), using samples from time-slice simulations (at 16 ka, 14 ka, 12 ka, 9 ka and 6 ka) with the isotope-enabled Community Earth System Model (iCESM; Brady et al., 2019) as the model prior. Osman et al. (2021) constructed a time-evolving prior ensemble, sampling the iCESM time slices according to a probability distribution centred
 135 on each target reconstruction age, allowing the covariance structure of the prior to evolve continuously across the glacial-interglacial transition. Instead of ice-core data, Osman et al. (2021) used marine sea-surface temperature proxies, including alkenones ($\text{U}^{\text{K}_{37}}$), planktic foraminiferal $\delta^{18}\text{O}$ values, planktic foraminifera Mg/Ca ratios, and the tetraether index of isoprenoid

glycerol dialkyl glycerol tetraethers (TEX₈₆), using proxy system models to translate between the iCESM climate variables and the proxy observation space. The LGMR provides global spatial coverage across the full deglacial and Holocene periods.

Buizert et al. (2018) produced Greenland-specific, seasonally resolved surface temperature fields for the last deglaciation using a proxy-informed model-adjustment approach. They used temperature estimates from the GISP2, NGRIP, and NEEM ice cores that were obtained using an inverse firn-densification modelling technique in which temperature and accumulation scenarios are optimized to fit measurements of the $\delta^{15}\text{N}$ of N_2 gas (a proxy for firn thickness) and constraints on the ice-core gas-age–ice-age difference. These temperature estimates were merged with the TraCE-21k transient simulations (Liu et al., 2009). Their approach assumes that TraCE-21k correctly captures the Greenland climate response to orbital and greenhouse gas forcing, but that model-data discrepancies arise from the model's failure to correctly simulate AMOC variability. A spatially and seasonally varying correction for the identified AMOC bias was derived from the model and applied at each time step, producing monthly temperature fields across the full Greenland domain that capture the distinct seasonal responses of temperature to orbital, greenhouse gas, and AMOC forcing through the deglaciation.

Finally, we use a new data-assimilation reconstruction (“Truax”) that builds on previous work in Buizert et al. (2018), Badgeley et al. (2020), and Osman et al. (2021), to produce a Greenland-focused reconstruction of temperature and precipitation amount over the past 20,000 years. This assimilation incorporates both the ice-core $\delta^{18}\text{O}$ time series used in Badgeley et al. (2020) and the ice-core $\delta^{15}\text{N}$ -derived temperature estimates from Buizert et al. (2018) and Martin et al. (2024). We use the same iCESM time-slice simulations as in Osman et al. (2021), but with a different sampling strategy. We identify a systematic mean-state warm bias in the iCESM simulations relative to the ice-core records, most pronounced during the mid-Holocene, which we address by subtracting the offset between the 3 ka and preindustrial simulations from all prior fields prior to assimilation. A time-evolving prior ensemble is then derived by sampling time slices according to a probability distribution centred on each target reconstruction age. We extend the approach of Osman et al. (2021) in two ways. First, we apply time-varying truncation bounds that prevent climatically implausible states from contributing to the prior at any given time step (e.g., model output from the Younger Dryas is not sampled for the Holocene prior). Second, we incorporate iCESM water-hosing experiments as an additional time-slice bin representing Younger Dryas boundary conditions, allowing the prior covariance structure to reflect the abrupt AMOC-driven cooling of that interval.

The five temperature reconstructions differ slightly in their treatment of seasonality. The Buizert et al., (2018), Truax, and two Badgeley et al. (2020) reconstructions use TraCE-21ka to provide the seasonality and spatial pattern. In Badgeley et al. (2020) and Truax, the TraCE-21ka seasonal cycle is low-pass filtered, while Buizert et al., (2018) retained the unsmoothed model seasonal cycle. Osman et al. (2021) reconstructs seasonal change from the iCESM time-slice prior, which is built from discrete time intervals (e.g., 9 ka, 6 ka) but samples continuously from the probability distribution of those slices, such that seasonality evolves through time.



2.3 Factor Analysis

170 To identify coherent signals shared among the proxy time series and the data-assimilation products, we apply Gaussian factor analysis (FA), a latent-variable model closely related to probabilistic principal component analysis (pPCA) (Tipping and Bishop, 1999). Let $X \in \mathbb{R}^{T \times P}$ denote the analysis matrix, where T is the number of time bins and P is the number of temperature time series. For each time bin t , the observation vector x_t is modelled using Eq. (1)

$$x_t = \mu + Z_t \cdot W + \epsilon_t \quad (1)$$

175 where μ is the mean of these observations, $Z \in \mathbb{R}^q$ is the vector of latent factor scores, $W \in \mathbb{R}^{P \times q}$ is a factor loading matrix that maps latent factors into the observation space, and $\epsilon_t \sim N(0, \psi)$. ψ represents the record-specific residual variance not explained by the factors.

The main distinction between pPCA and our FA implementation is the treatment of the residual variance. In pPCA, the residual covariance is isotropic, such that $\psi = \sigma^2 I$, and all of the time series are assumed to have the same unexplained variance (Tipping and Bishop, 1999). In our FA implementation, ψ is diagonal but not isotropic, with a common residual variance assigned to all time series of the same proxy type, $\psi = \text{diag}(\psi_1, \dots, \psi_j)$, where ψ_j is determined by the proxy type represented by time series j . This proxy-specific uncertainty accounts for the unique biases of different proxy systems (e.g., seasonal sensitivity, habitat/ecology, and the archival/recording process). This formulation effectively down weights proxy types with greater uncertainty and gives greater influence to proxy types with lower uncertainties, similar to data assimilation. Similar uncertainty-weighting approaches have been used previously to reduce the influence of high-variance paleoclimate observations (e.g., Hong et al., 2023). For the quantitative proxy time series, we set ψ_j using the errors associated with the proxy-temperature transfer functions used to estimate the temperatures (Table 2). For qualitative proxies, we assign an uncertainty of 1 ordinal unit, allowing deviations of one category. Meanwhile, for each climate reconstruction product, we assign a common isotropic residual variance to all spatial time series. Unlike the proxy network, for which uncertainty can be assigned by proxy type, the time series within a given climate reconstruction product are generated using the same reconstruction framework and do not have directly comparably time-series-specific uncertainty estimates.

185 We evaluate our FA framework and its ability to recover previously identified regional climate structures by applying it to the Sundqvist et al. (2014) proxy network for Greenland and the eastern Canadian Arctic, which Briner et al. (2016) analysed using pPCA. Our FA recovers comparable leading temporal modes (Z_1 and Z_2) and spatial loading patterns (W_1 and W_2), providing a methodological validation and consistency check on our implementation (Text S1, Fig. S1).

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Table 2. Proxy-specific error estimates used to define ψ priors in the factor analysis. Error values represent the assumed uncertainty for each proxy type, reported in interpreted units, with supporting references for each estimate.

Proxy	Error	Units	Reference
brGDGTs	2.85	°C	(Martínez-Sosa et al., 2021)
chironomid	2.00	°C	(Heiri et al., 2011)
chlorophyll	1.71	%	(Von Gunten et al., 2012)
diatom	0.90	°C	(Korhola et al., 2000)
dinocyst	1.50	°C	(Hohmann et al., 2023)
macrofossil	1.00	Qualitative	(This study)
pollen	2.00	°C	(Herzschuh et al., 2023)

2.3.1 Preprocessing

Prior to the factor analysis, each proxy time series is standardized (z-scored) to normalize variance through time, an important step which ensures that higher-variance records do not dominate the identified shared signal obtained from the collective records. Normalization is a common step in PCA, as higher variance records have a larger impact on the covariance structure of the dataset, which results in the higher variance more heavily weighing on the primary axes. Similarly in factor analysis, higher variance records will dominate the covariance structure and more heavily weigh on the leading latent factors. It is important to note that while normalization in a traditional PCA can inflate low variance noise within time series to effectively create a pseudo signal, our handling of the error term, ϵ , negates this phenomenon.

Another important distinction between PCA and factor analysis is that PCA produces the complete set of orthogonal components permitted by the rank of the data, after which diagnostic criteria may be used to determine which components are meaningfully interpretable. In contrast, the number of latent factors (q) in factor analysis is a user-specific model dimension. We therefore fit factor models across a range of candidate values of q , from 2 to the number of proxy records available, and evaluate model support using the Bayesian Information Criterion (BIC; Neath and Cavanaugh, 2012; Schwarz, 1978). BIC



acts to balance improved fit against model complexity and we interpret the q that minimizes BIC as the number of dominant spatiotemporal factors that can be extracted from the network without overfitting.

215 The capability of the factor analysis to handle missing data derives from the model effectively ignoring these absent observations (Tipping & Bishop, 1999). In other words, for time steps with missing data, the model computes the latent factor scores and corresponding loadings from valid observations. As a result, weakly overlapping time series data will produce factor solutions less reflective of the full dataset and more akin to a mosaic of individual records. To address this, we incorporate a two-stage age binning method: the time series are binned at 50-year intervals, and then again at 300 years. This two-stage age binning method is favourable because the first stage effectively accounts for potential sampling biases in the records, providing a more representative estimate of mean values for each 300-year age bin.

220 2.3.2 Interpretation

To summarize how variance is partitioned among retained factors, we sum the squared loadings for each factor across all records and normalize these values by the summed squared loadings across all retained factors. The resulting fractions sum to one and represent the proportion of the factor-attributable variance associated with each factor. These normalized fractions are reported as the variance explained by each factor throughout the manuscript.

225 To aid interpretation of the FA solutions, we group the proxy time series by the sign of their factor loadings (positive vs. negative) for each factor and report the mean and 1σ spread of the centred time series within each group. Using the 300-year time binned values, we centred each proxy time series by subtracting its mean over 10-0ka, isolating shared temporal structure from differences in absolute magnitude or baseline temperature across sites. Quantitative (Fig. 3B, 3G) and qualitative (Fig. 3C, 3H) proxy time series were treated separately throughout, as they differ in scale.

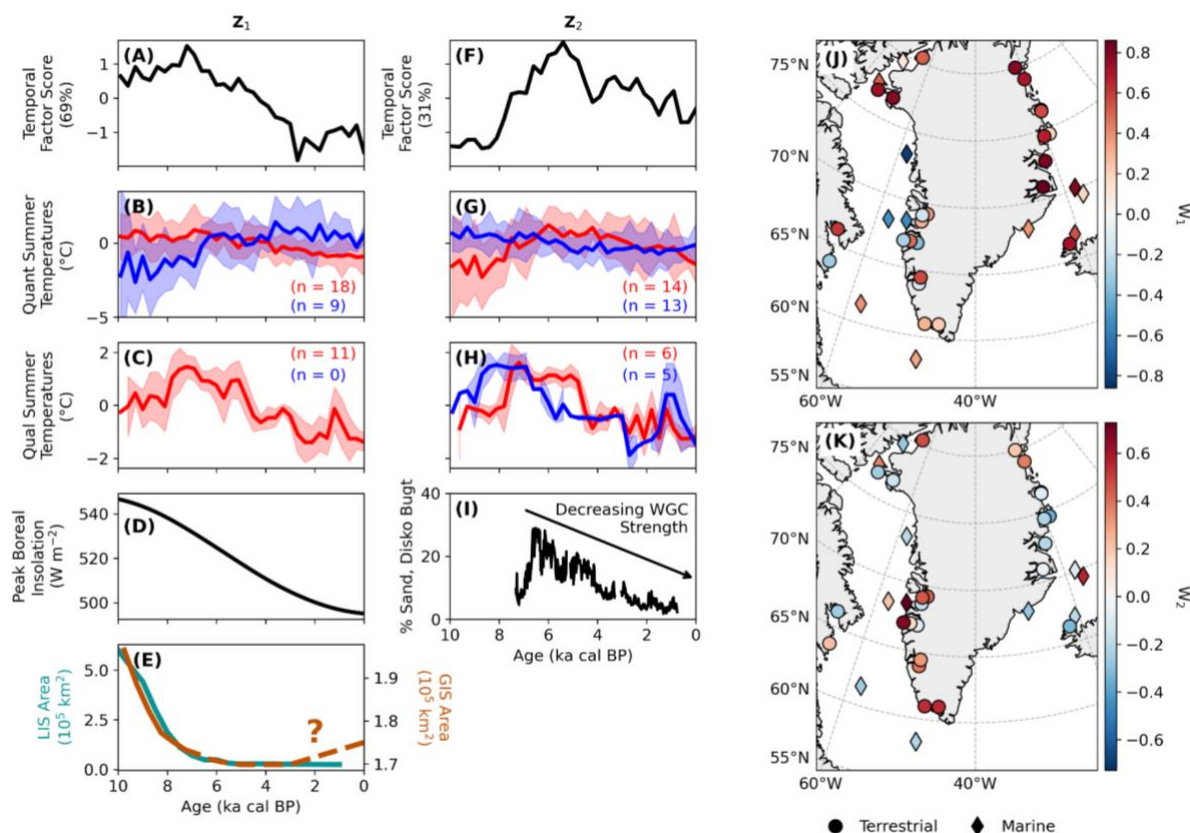


Fig. 3. Factor-analysis solutions for the full Greenland proxy network and comparison with key Holocene forcing and boundary-condition time series. (A) Temporal factor score Z_1 , with the percentage of network variance explained. (B) Quantitative summer-temperature proxy time series grouped by positive (red) and negative (blue) W_1 loadings. Individual time series were centred prior to averaging; lines and shading indicate the group mean and $\pm 1\sigma$, respectively. (C) Qualitative summer-temperature proxy time series grouped by positive (red) and negative (blue) W_1 loadings. (D) Boreal summer insolation at 70°N on June 21 (Laskar et al., 2004). (E) Reconstructed Laurentide Ice Sheet area (green; Dalton et al., 2023) and Greenland Ice Sheet area (orange; Leger et al., 2024). (F) Temporal factor score Z_2 , with the percentage of network variance explained. (G) Quantitative summer-temperature proxy time series grouped by positive (red) and negative (blue) W_2 loadings, with centring, lines, and shading as in B. (H) Qualitative summer-temperature proxy time series grouped by positive (red) and negative (blue) W_2 loadings. (I) Percent sand from a marine sediment core in Disko Bugt, illustrating changes in benthic West Greenland Current strength (Perner et al., 2013). (J) Spatial loadings W_1 . (K) Spatial loadings W_2 .

2.4 Proxy-Climate Reconstruction Product Comparison

Comparing FA solutions across the proxy network with multiple climate reconstruction products requires a common model dimension and reference orientation. We first use BIC to select the number of latent factors (q) and then hold q fixed across the solutions being compared. Still, FA solutions could differ in factor orientation, sign, and scale without changing the reconstructed signal. We therefore apply a Procrustes-style (Schönemann, 1966) least-squares alignment to place the factors from each climate reconstruction product in the same reference space as the proxy factors, prior to computing similarity metrics.



Our alignment calculation considers each factor independently. For each factor k , we compare the climate reconstruction product Z_k to the proxy reference Z_k and determine the best alignment under two constraints. First, we resolve sign indeterminacy by selecting the sign that maximizes the similarity between the climate reconstruction product and proxy score series, based on their inner product. Whenever the sign of Z_k is reversed, the sign of the corresponding W_k is also reversed. Second, we apply a single multiplicative amplitude scaling of the climate reconstruction product Z_k and W_k , bringing the reconstruction product and proxy solutions to comparable amplitudes so that the resulting misfit reflects differences in temporal modes and spatial loading patterns rather than arbitrary scaling. This scaling is bounded to 0.2 to 2 to avoid unrealistically large attenuation or amplification when matching a muted score series.

Once aligned, we quantify similarity using normalized Euclidean misfit for both the temporal scores and spatial loadings. For each factor k , we compute a distance for the score series (dZ_k) and a distance for the loading pattern (dW_k) as the norm of the difference between aligned climate reconstruction products and proxy vectors. These distances are normalized by the norm of the corresponding proxy reference vector and combinedly reported as the mean.

The loading fields of the climate reconstruction products are sampled at the proxy site coordinates. For each proxy site, we identify the nearest climate reconstruction product grid cell and extract the climate reconstruction product loadings from that grid cell. This produces an aligned climate reconstruction product loading matrix on the same spatial footprint as the proxy loadings prior to Procrustes alignment and distance calculation.

To quantify variability in the proxy-derived factor solution due to proxy-network composition, we construct a proxy bootstrap distribution using resampling over proxy records (Diciccio and Romano, 1988). Specifically, we perform 100 bootstrap replicates in which we randomly subsample proxy time series, rerun the factor analysis, and align the resulting factor scores and loading patterns to the full-network proxy reference solution using the same factor-wise alignment procedure described above. This yields bootstrap distributions of dZ_k and dW_k , and the combined distance d_k for each factor. We interpret these distributions as the range of solution variability expected from plausible proxy-network perturbations, providing a baseline against which factor solutions of the climate reconstruction products can be evaluated. To assess the sensitivity of this variance distribution to the proportion of resampled sites, we repeated the bootstrap procedure while resampling 50, 60, 70, and 80% of the proxy network (Fig. S1).

3. Results

3.1 Proxies

Our Greenland proxy network is dominated by two spatiotemporal factors of Holocene climate variability, which explain 69% and 31% of the factor-attributable variance, respectively (Fig. 3A, 3F). The BIC indicates an optimal solution of two factors, supporting this interpretation. The factor scores of the leading latent factor, Z_1 , are high and increase from 10 to ~7 ka, decrease until ~3 ka, and then increase towards modern (Fig. 3A). W_1 has predominantly positive loadings in northern and eastern Greenland and negative loadings concentrated in western Greenland and Baffin Bay (Fig. 3J). Quantitative proxy time series



with positive W_1 loadings display a similar pattern, with high and stable early Holocene summer temperatures decreasing by ~ 7.0 ka (Fig. 3B, red line). However, unlike the late Holocene warming inferred from Z_1 , these quantitative time series indicate stable or slightly decreasing temperatures, after 2.0 ka. The qualitative proxy time series also have positive W_1 loadings and generally indicate relative cool conditions before ~ 8 ka, warmth between ~ 8 and 5 ka, and cooling thereafter (Fig. 3C). In contrast, quantitative proxy time series that have negative W_1 loadings are clustered in western Greenland and Baffin Bay and display warming from 10.0 to 3.0 ka, followed by cooling towards modern (Fig. 3B, blue line). Importantly, both loading-sign groups contain time series derived from multiple proxy types (Table 1), indicating that the east-west contrast in W_1 is due to regionally coherent climate forcing rather than proxy-specific responses.

The second factor score, Z_2 , is consistently low from 10 to ~ 8.3 ka, increases to a maximum by ~ 5.4 ka, and then decreases towards modern (Fig. 3F). The W_2 loading pattern is heterogeneous, with the strongest positive loadings at Lake Rosaea (Site 11, Fig. 3K) and the marine sediment core MSM343300 in Disko Bugt (Site 16, Fig. 3K). Generally positive but variable loadings are also observed across southern Greenland and six sites in northern Greenland (Fig. 3K). Both the quantitative and qualitative proxy sites with positive W_2 loadings indicate general warming during the early Holocene, a mid-Holocene thermal maximum from ~ 6 to 4 ka, and cooling thereafter. Together, the two-factor solution indicates that Holocene summer-temperature variability in Greenland was characterized by two distinct temporal modes with spatially coherent expressions that are not proxy-dependent.

To assess the sensitivity of the FA solution to proxy-network composition, we used bootstrapping to generate variance distributions by repeatedly subsampling the proxy network at different site-retention levels and comparing each subsampled solution to the full-network solution. The temporal score distances for Z_1 and Z_2 , as well as the spatial-loading distances for W_1 and W_2 show a systematic dependence on the proportion of proxy sites retained (Fig. S2). As expected, distances decrease toward 0 as the proportion of retained proxy time series increases, indicating progressively stronger agreement between the subsampled and full-network solutions.

The convergence of Z and W distances to the full-network solution is most pronounced in the second factor pair (Z_2 and W_2). Between the 50% and 80% resampling runs (Fig. S2, top vs. bottom row), the mean distances of Z_2 and W_2 decrease by 0.25 and 0.20, respectively. In contrast, the Z_1 and W_1 distances decrease by 0.07 and 0.03, respectively. This asymmetry suggests that the secondary temporal and spatial factors are more sensitive to the proportion of proxy sites included in the analysis relative to the leading factor solutions.

The largest shift from high- to low-distance solutions occurs between the 60% and 70% resampling runs, while the additional improvement from 70% to 80% is comparatively small (Fig. S2). We therefore use the 70% resampling ensemble as a representative estimate of proxy-solution variance because it captures the dominant temporal and spatial structure of the full-network solution while still imposing substantial perturbations to proxy-network composition.



3.2 Reconstruction-Product--Proxy Data Comparisons

When comparing the mean $Z_{1,2}$ and mean $W_{1,2}$ of FA solutions, the solutions of the reconstruction products generally fall within the expected distribution of proxy variance (Fig. 4). Considering the mean of the temporal and spatial distances, the
 315 Truax, Badgeley Preferred, and Badgeley Low reconstructions yield the most similar results to the full proxy network, with mean distances of 0.72, 0.73, and 0.77, respectively. Meanwhile, the Osman product yields the largest mean distance of 0.95 (Table S1).

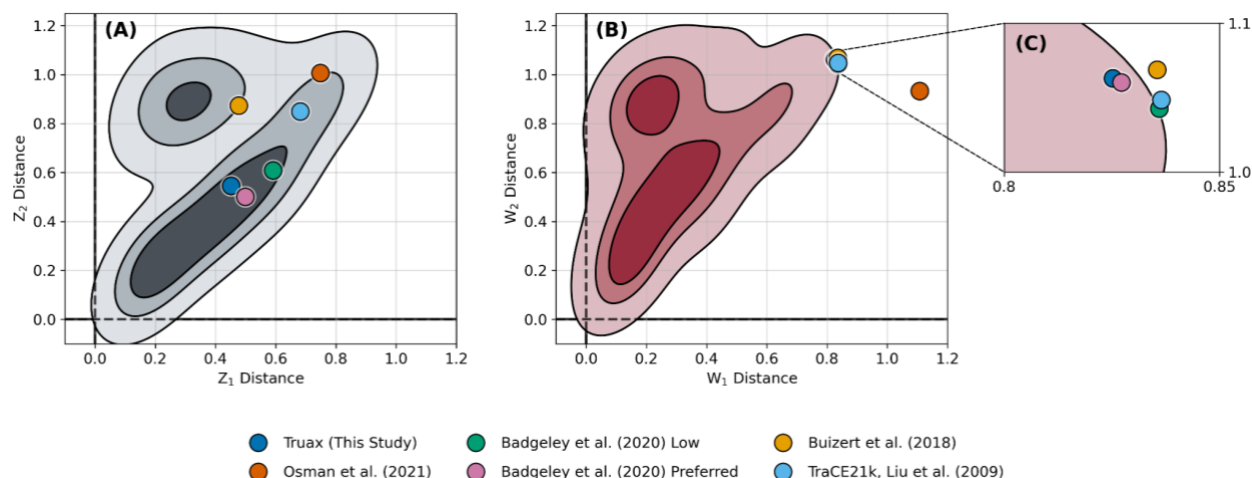


Fig 4. Evaluations of the temporal, Z, and spatial, W, similarities between spatial climate fields from climate reconstruction products and independent proxy data from coastal Greenland and adjacent seas. (A) Procrustes distances between the climate reconstruction products (circles) and full proxy network, considering mean temporal-score distances, dZ , and (B) considering spatial loading-pattern distances, dW . (C) Expanded view of the loading-pattern distances distinguishes products with similar distances. In all panels, variance within the proxy network is represented by a distribution, computed via Monte Carlo resampling (70% resampling, without replacement) (shaded regions; lightest, medium, and darkest regions span 95th, 68th, and 34th percentiles of resampled data, respectively).

All of the temporal distances computed for the reconstruction products fall within the 95th percentile of proxy Z -distance variance, indicating general agreement with the proxy-derived temporal modes (Fig. 4A). However, the reconstruction products separate into two groups. Truax, Badgeley Preferred, and Badgeley Low yield lower Z distances that fall within the 34th percentile of proxy Z -distance variance, indicating closer agreement with the full-network proxy solution. In contrast,
 330 TraCE21k, Buizert, and Osman yield higher Z distances, particularly for Z_2 , that are more similar to 50% resampling distribution (Fig. S2A). As such, although all reconstruction products produce temporal modes that fall within the proxy-derived variance envelope, the Truax and Badgeley reconstruction products more closely reproduce the full-network temporal modes.

In comparison, the spatial loading distances are more discriminatory. Considering W_1 and W_2 together, Badgeley
 335 Low, TraCE21k, Buizert, and Osman fall outside the 95th percentile of proxy W -distance variance, whereas Truax and Badgeley Preferred fall within the 95th percentile (Fig. 4B). Even the Truax and Badgeley Preferred spatial loading-pattern distances lie near the edge of the proxy-derived variance envelope (Fig. 4C) and no reconstruction product falls within the 68th



percentile, suggesting that the reconstruction products better capture the temporal modes than the spatial loading patterns of the proxy network.

340 4. Discussion

4.1 Proxy Factor Analysis

The two spatiotemporal patterns identified in the proxy network indicate regional differences in the timing of Holocene summer-temperature change across Greenland (Fig. 3). These regional patterns are inconsistent with a single, spatially uniform summer-temperature history across Greenland, as might be expected from a climatic system responding solely and uniformly to declining boreal summer insolation. Meanwhile, the geographic organization of similar proxy histories suggests that the observed variability reflects shared regional climate signals rather than site-specific variability or noise. This two-factor structure is consistent with previous interpretations of a dominant two-factor Holocene climate structure across Greenland and northern Canada (Fig. S1; Briner et al., 2016).

The resampling analysis indicates that the stability of the FA solution depends on proxy-network completeness (Fig. S2). Both the temporal scores and spatial loadings become increasingly similar to the full-network solution (i.e., closer to the origin in Fig. S2) as the proportion of retained sites increases. This behaviour is expected for EOF- or FA-like decompositions, which can be sensitive to sampling density, spatial network geometry, and the separation between factors (Hannachi et al., 2007; North et al., 1982). Therefore, if more summer-temperature proxy time series were to become available, especially in locations with currently few constraints (e.g., over the ice sheet, in southeastern Greenland, Melville Bugt, northernmost Greenland, and much of the surrounding seas), the temporal scores and spatial loadings could be refined.

4.2 Comparison of Proxies and Climate Reconstruction Products

For the leading temporal mode, all reconstruction products broadly reproduce the mid- to late Holocene cooling trend identified as Z_1 in the proxy network (Fig. 5A). However, the main difference among products occurs during the early Holocene, when the full proxy-network Z_1 solution increases to peak values at ~ 7.2 ka. The Osman and TraCE21k products yield comparatively high Z_1 distances and indicate maximum values before ~ 9.5 ka, followed by cooling toward modern (Fig. 4A, 5A). The Buizert reconstruction displays a similar Z_1 trend to these reconstruction products, but with stable Z_1 values from 8 to 6 ka, coincident with peak proxy values (Fig. 5A). Consequently, the Buizert reconstruction yields a smaller Z_1 distance (Fig. 4A). In contrast, the Truax, Badgeley Low, and Badgeley Preferred reconstructions yield lower Z_1 distances and display warming from 10 to ~ 7 ka before cooling toward modern. This agreement supports the interpretation of Z_1 as capturing an early Holocene increase in summer temperatures, which peaked at ~ 7.2 ka, rather than a monotonic cooling trend following an early Holocene maximum temperature at 10 ka.

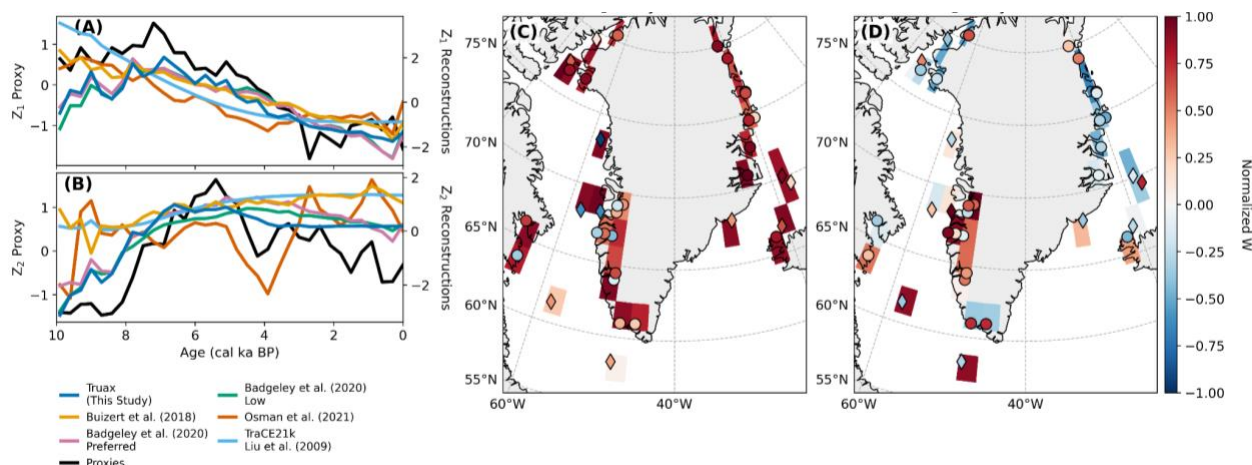


Fig. 5. Comparison of temporal factor scores and spatial loading patterns between the full proxy network and climate reconstruction products. Temporal factor scores, Z , are shown for the first (A) and second (B) factors of the proxy network and climate reconstructions. Spatial loading patterns, W , compare proxy-site loadings with corresponding grid-cell loadings from the Truax reconstruction for the first (C) and second (D) factors. Proxy loadings are shown as circles for terrestrial sites and diamonds for marine sites, while Truax loadings are shown as underlying grid cells. All spatial loadings are normalized to the maximum absolute value to facilitate comparison between the proxy network and reconstruction product.

The differences in W_1 distances are consistent with the greater spatial homogeneity of the loading patterns in the reconstruction products than in the proxies (Figs. 5C, S4). In contrast to the proxy network solutions, the reconstruction products generally show positive W_1 loadings across all grid cells and do not reproduce the negative loading pattern observed in western Greenland-Baffin Bay in the proxies (Fig. 5C).

The second temporal mode, Z_2 , provides a clearer separation among reconstruction products. Osman, TraCE21k, and Buizert yield relatively high Z_2 distances (Fig. 4A) and do not reproduce the proxy-derived mid-Holocene thermal maximum (Fig. 5B). TraCE21k and Buizert instead show more monotonic Holocene warming, while Osman displays a less coherent Z_2 trend (Fig. 5B). These results indicate that although these reconstruction products can reproduce aspects of the leading proxy temporal mode, they do not accurately capture the secondary temporal mode associated with mid-Holocene warmth.

The Truax, Badgeley Low, and Badgeley Preferred climate reconstruction products more closely reproduce the proxy-derived mid-Holocene thermal maximum, yielding lower Z_2 distances than the Osman, TraCE21k, and Buizert reconstructions (Fig. 4A, 5B). Although the Badgeley Preferred product has a slightly lower Z_2 score than the Truax product, the latter produces the combination of Z_2 and W_2 solutions that is most similar to the independent proxy network. The Truax product indicates early Holocene warming followed by a mid-Holocene thermal maximum that is similar in timing and duration to the proxy-derived Z_2 temporal mode (Fig. 5B), whereas the Badgeley Preferred product indicates a more prolonged interval of mid-Holocene warmth. The Truax product also produces a uniquely positive heavy loading pattern in southwestern Greenland, similar to the proxy network solution (Fig. 5D, S5). This pattern suggests that the positively loaded grid cells in southwestern Greenland and the adjacent North Atlantic region capture a mid-Holocene warmth signal that is not clearly expressed in the other reconstruction products but is an important feature of the proxy network. On the other hand, the Truax product is not



fully consistent with the proxy-derived spatial structure. In particular, the Truax product produces strong negative W_2 loadings in northern Greenland, whereas the proxy network shows positive loadings in this region (Fig. 5D). These negatively loaded grid cells appear to reflect a more stable early Holocene thermal maximum followed by cooling toward modern, rather than the mid-Holocene thermal maximum represented by positive W_2 loadings (Fig. S3B, Factor 2). The cause of this discrepancy is unclear, but it indicates that even the reconstruction product that is most similar to the independent proxy network still fails to fully reproduce the spatial structure of the proxy-derived second mode.

The Badgeley Preferred product slightly outperforms the Truax product in the Z_2 and W_2 Procrustes comparisons (Fig. 4A, 4C). However, this agreement is not reflected as clearly in the underlying temporal structure. Both Badgeley products suggest a longer interval of mid-Holocene warmth, with maximum values occurring near $\sim 4\text{ka}$, approximately 1.5 kyr later than in the proxy network (Fig. 5B). In the raw temperature time series, grid cells with positive and negative W_2 loadings show relatively limited separation (Fig. S3D, S3E), suggesting that the second factor of the Badgeley products does not distinguish clearly between regions where the proxies suggest there were different Holocene temperature histories. This muted spatial differentiation is also reflected in the loading patterns, where the limited number of positively loaded grid cells are not concentrated in southern and southwestern Greenland (Fig. S5). As such, while the Badgeley products capture a similar second temporal mode, its W_2 spatial loading pattern is inconsistent with the proxy network.

Broadly, all reconstruction products capture some aspects of the dominant proxy-derived temporal trends, but none fully reproduce the combined temporal and spatial structure of the proxy network, because the W_1 patterns in the products all tend to be quite homogenous (Fig. S3, left column; Fig. S4). While the reconstruction products provide useful approximations of broad Holocene temperature variability, they do not fully recover the regional climate structure expressed by the independent Greenland proxy network. Accordingly, we use these products to assess broad-scale agreement and reconstruction limitations, whereas our interpretations of regional Holocene summer-climate dynamics and their forcing mechanisms are based on the proxy time series and the paired temporal modes and spatial loading patterns identified by the factor analysis.

4.3 Using Factor Analysis to Interpret Greenland-wide Holocene Climate Patterns and Processes

The factor analysis identifies coherent spatiotemporal factors in the proxy network but cannot determine the underlying mechanisms that produced these patterns. Any proposed paleoclimatic interpretation of these results must therefore account for three principal features of the two-factor solution: early Holocene warmth followed by cooling across northern and eastern Greenland, a delayed and opposing response in western Greenland and Baffin Bay, and a middle-Holocene thermal maximum in southern and southwestern Greenland. We therefore use these spatiotemporal constraints to evaluate the potential climatic and oceanographic mechanisms based on their ability to explain the timing and geographic expression of the identified patterns.

The regional summer-temperature patterns identified by the factor analysis may reflect the combined influences of orbital forcing, greenhouse-gas concentrations, and oceanographic processes (e.g., Axford et al., 2021; Briner et al., 2016). Orbital and greenhouse-gas forcing would have similar temporal impacts across the entire study region. The regional contrasts in the timing of summer warmth therefore suggest that regional processes modified the temperature response to these shared



influences. Oceanographic dynamics provide a framework for exploring these contrasts because differences in freshwater input, sea-ice cover, stratification, and Atlantic-water influence could have distinct spatial expressions (e.g., Gibb et al., 2015; Hansen et al., 2020). We distinguish two oceanographic regimes around Greenland that may have influenced the proxy network: (1) a freshwater-dominated, strongly stratified surface-water regime in Baffin Bay and (2) an Atlantic-influenced regime associated with the Irminger Current (IC) and West Greenland Current (WGC) along southern and southwestern Greenland.

In Baffin Bay, cold, low-salinity Arctic-origin water and sea ice are transported southward in the near-surface Baffin Island Current, whereas the northward-flowing WGC carries a mixture of Polar Water and warmer, more saline Atlantic-origin water along the eastern margin of the Basin (e.g., Münchow et al., 2015; Tang et al., 2004). During the early Holocene, meltwater from the retreating Greenland and Laurentide ice sheets further freshened the surface ocean in Baffin Bay, strengthening stratification and limiting vertical exchange with the warmer water below (Hansen et al., 2020; Jennings et al., 2011; Lloyd et al., 2005; Münchow et al., 2015; Saini et al., 2020). Marine proxy time series document extensive sea ice and strong meltwater influence in eastern Baffin Bay between ~9.2 and 7.9 ka, followed by declining meltwater influence, reduced sea-ice cover, and a greater subsurface Atlantic-water influence during the middle Holocene (Gibb et al., 2015; Hansen et al., 2020). Baffin Bay surface conditions could therefore remain comparatively cold even when Atlantic-origin water was present at depth.

A different oceanographic expression characterizes southern and southwestern Greenland, which lie closer to the pathway through which Atlantic-origin IC water enters the WGC. Here, the climatic influence of Atlantic water also depends both on the amount or temperature of the inflowing water and on whether freshwater stratification restricts its exchange with surface waters. Reconstructions along the West Greenland margin distinguish between subsurface Atlantic-water conditions and surface environments affected by meltwater and sea ice, demonstrating that the two can vary independently (Krawczyk et al., 2017; Moros et al., 2016; Ouellet-Bernier et al., 2014). These reconstructions also indicate enhanced Atlantic-water influence during parts of the middle Holocene, when reduced meltwater influence accompanied a smaller Greenland Ice Sheet and allowed Atlantic-origin water to exert a stronger regional influence (Kvorning et al., 2025; Perner et al., 2013). As such, a later expression of Atlantic-derived heat may reflect a reduction in the freshwater barrier separating subsurface Atlantic water from the environments recorded by marine and coastal proxies.

These two regimes – freshwater and sea-ice suppression of surface warming in Baffin Bay and the changing surface and coastal expression of Atlantic-origin heat around southern and southwestern Greenland – provide the oceanographic framework used below to interpret the leading proxy factors.

4.3.1 Peak early Holocene warmth in northern and eastern Greenland contrasts with a late Holocene peak warmth in western Greenland (Z_1 and W_1)

The leading proxy mode, Z_1 , is paired with positive W_1 loadings across northern and eastern Greenland (Fig. 3J). Both the quantitative and qualitative time series associated with Z_1 indicate warm early Holocene conditions followed by cooling



through the middle Holocene (Fig. 3B, 3C, red curves). This pattern provides quantitative support for a previous synthesis suggesting that summer warmth was strongest in the early Holocene and was particularly pronounced in northwestern Greenland (Axford et al., 2021). Our analysis additionally highlights that this strong early Holocene warmth extends across northeastern Greenland from the Skallingen area to Scoresby Sund and includes the Greenland Sea.

Independent glacial geologic evidence from northwestern Greenland is consistent with the early Holocene warmth represented by Z_1 . In northwestern Greenland, retreat of the Greenland Ice Sheet from Inglefield Land between ~ 9 and 7 ka reflects rapid early Holocene deglaciation, potentially aided by both atmospheric warmth and ice-margin reorganization following the opening of Nares Strait, while regrowth of Prudhoe Dome at ~ 7 ka indicates subsequent cooling (e.g., Blake et al., 1992; Mason, 2010; Nichols, 1969; Søndergaard et al., 2020; Walcott-George et al., 2026). Farther east, the Northeast Greenland Ice Stream retreated tens of kilometres behind its present margin by ~ 7.8 ka, while the 79°N ice shelf was reduced or absent from ~ 8.5 to 4.4 ka (Bennike and Weidick, 2001; Larsen et al., 2018; Smith et al., 2023). An early Neoglacial moraine at Sanddlen indicates that a different northeastern sector had already returned to approximately its modern ice-margin configuration by 4.6 ± 0.4 ka (Smith et al., 2023; Briner and Bennike, 2025). Collectively, these indicators are consistent with strong early Holocene warmth followed by cooling, as represented by Z_1 , but their differing retreat and readvance chronologies emphasize spatially variable ice-dynamic responses rather than a systematically longer minimum phase in northeastern Greenland.

Ice-core-based mean annual temperature reconstructions provide context for the timing of the Z_1 maximum, with Holocene thermal maximum conditions occurring between ~ 9.9 and 6.9 ka across GISP2, NGRIP, and NEEM (Martin et al., 2024). Although Martin et al. (2024) identify a later onset of thermal maximum conditions toward northern Greenland, they suggest that the delayed onset at NEEM may reflect regional influences from Baffin Bay, elevation effects, lapse-rate changes, and near-surface inversions, rather than a simple latitudinal response to insolation. The range of Holocene thermal maximum timing in the ice-core reconstructions is therefore comparable to the ~ 7 ka maximum in Z_1 . Following this maximum, the decline in Z_1 is consistent with a response to orbitally forced summer cooling, as peak boreal summer insolation decreased after ~ 11 ka (Fig. 3D; Laskar et al., 2004), a relationship previously invoked to explain regional cooling (e.g., Axford et al., 2021; Briner et al., 2016). However, the Z_1 signal indicates two intervals of asynchrony between summer climate and boreal summer insolation. First, the increase of Z_1 from 10 to 7 ka departs from the expected response to declining summer insolation (Fig. 3A). Second, Z_1 values increase after ~ 3 ka, which is also inconsistent with simultaneously decreasing peak boreal insolation (Fig. 3A).

The increase in Z_1 from 10 to 7 ka is most apparent in the positively loaded qualitative time series (Fig. 3C). In contrast, the positively loaded quantitative temperature time series indicate warm but comparatively stable conditions during the early Holocene (Fig. 3B, red line). This divergence suggests that the rising Z_1 pattern reflects a delayed or regionally modified response to early Holocene warmth in northern and northeastern Greenland. Indeed, several original site interpretations described cold or nutrient-poor lake assemblages and delayed vegetation development following deglaciation, with warmer conditions established after ~ 8 ka (e.g., Bennike and Wagner, 2012; Klug et al., 2009; Schmidt et al., 2011;



Wagner et al., 2008; Wagner and Bennike, 2015). Despite high early Holocene insolation, regional warming may have been muted by cold surface waters, sea ice, and meltwater transported by the East Greenland Current (EGC) and associated Polar Water. This mechanism has been proposed to explain the delayed thermal maximum at sites in northeastern Greenland (Fig. 2, Sites 27, 29, 30), particularly along the outer coast, where climate was strongly influenced by the cold EGC (e.g., Schmidt et al., 2011; Wagner et al., 2008). Oceanographic reconstructions from the Northeast Greenland Shelf and Fram Strait support this interpretation, indicating cold, sea-ice-influenced surface waters and strong EGC/Polar Water influence during the early Holocene, followed by reduced Polar Water influence and higher surface temperatures after 8.2 ka (Davies et al., 2022; Pados-Dibattista et al., 2022). As such, the rise in Z_1 from 10 to 7 ka could reflect the delayed regional expression of early Holocene warmth in parts of northern and northeastern Greenland, rather than a Greenland-wide summer-temperature trend.

In comparison, northwestern Greenland ($> 75^\circ\text{N}$, $>60^\circ\text{W}$) presents a more heterogeneous spatial pattern. Previously suggested factors such as ice-sheet dynamics, meltwater injection, sea-ice variability, ocean circulation changes, and proxy-specific sensitivities may have influenced the proxy network compiled here (Axford et al. 2021; Briner et al., 2016). Two of the three southernmost time series in northwestern Greenland extend into the early Holocene and have large positive W_1 loadings. The terrestrial time series from Carey Øer (Fig 2, Site 31; Bennike et al., 2008) indicates summer temperatures greater than modern values from ~ 9.3 to 7.4 ka, while the Deltasø (Fig 2, Site 32; Axford et al., 2019) time series indicates the highest Holocene summer temperatures from ~ 10 to 6.2 ka. The third time series, HU91-039-008P+-007BC (Fig. 2, Site 33; Levac et al., 2001) begins after ~ 7 ka and therefore does not constrain earlier conditions. The northernmost marine site (Fig 2, Site 35; Caron et al., 2019) does not have large loadings on either W_1 or W_2 , but the raw temperature time series indicates warmest conditions prior to 8.6 ka, suggesting that some sea-surface temperature time series are compatible with early warmth. The northernmost terrestrial site (Fig 2, Site 37; Bennike, 2002) has positive W_1 and W_2 loadings and suggests a mid-Holocene thermal maximum. A similar disparity was previously noted by Axford et al. (2021), identifying strong evidence for early-Holocene warmth in this region, including a high abundance of summer melt layers from the Agassiz Ice Cap between 11 and 8 ka (Lecavalier et al., 2017). However, other time series indicate later peak warmth, which may reflect the influence of ocean circulation, sea ice, meltwater discharge, and ocean gateway changes in Baffin Bay and Nares Strait (e.g., Gibb et al., 2015; Hansen et al., 2020; Jennings et al., 2011; Okuma et al., 2023; Ouellet-Bernier et al., 2014; Saini et al., 2020). Thus, northern Greenland likely experienced strong early Holocene summer warmth, but the timing and expression of that warmth varied among archives, sites, and proxy types. This in turn impacted the timing of peak warmth, which manifested as a lag between the primary forcing mechanism, peak summer insolation, and the peak temperature in this region.

The dominant northern expression of Z_1 is also evident among six time series in northern Greenland ($> 76^\circ\text{N}$). Six time series have positive W_1 and W_2 loadings, but larger loadings on W_1 (Fig. 2, Sites 28, 29, 30, 34, 36, and 37). The original interpretations of these time series generally describe broad early-to-middle Holocene warm intervals with warmest conditions that occurred closer to the ~ 7 ka Z_1 maximum but persisted sufficiently long enough to covary secondarily with Z_2 . At Store Koldewey in northeastern Greenland, lake bioproductivity reached a long-term maximum during the early Holocene, whereas chironomid assemblages from Duck and Hjort lakes indicate relatively warm conditions from ~ 8 to 5 ka (Klug et al., 2009;



Schmidt et al., 2011). The delayed development and subsequent termination of these conditions were attributed partly to the influence of the cold East Greenland Current and associated sea ice (Klug et al., 2009; Schmidt et al., 2011). Farther north, Lille Sneha Sø records a thermal maximum extending from ~7.7 to 4.4 ka, but its highest temperatures occurred by ~7 ka, while evidence from Nioghalvfjærdsfjorden indicates glacier-free and seasonally open-water conditions during much of the same interval (Bennike and Weidick, 2001; Wagner and Bennike, 2015). In northwestern Greenland, the Washington Land time series constrains warm marine conditions by ~7.3 ka rather than a discrete middle-Holocene temperature maximum (Bennike, 2002). Regional syntheses likewise identify early warmth in northern and eastern Greenland, while showing that local deglaciation and changes in coastal sea ice could delay or prolong its ecological expression (Briner et al., 2016; Funder et al., 2011). Thus, the secondary positive W_2 loadings at these northern sites most likely reflect the persistence of early Holocene warmth into the middle Holocene and broadly synchronous cooling after ~5 ka.

The second departure of Z_1 from the insolation trend – an increase after ~3 ka – may reflect a limitation of EOF-style analyses when time series with different Holocene trends are forced into a single shared mode. When the quantitative proxy time series are separated by the loading sign of W_1 (Fig. 3B), the sites with positive W_1 loadings (red curve) show stable early-Holocene warmth followed by cooling after ~7 ka and stabilization after ~2 ka. The positively loaded qualitative time series indicates a similar pattern, with warming during the early Holocene followed by general cooling towards modern after ~7 ka (Fig. 3C). In the Baffin Bay and western Greenland region, independent climate reconstructions display a similar cooling trend after ~3 ka (e.g., Kvorning et al., 2025; Limoges et al., 2020; Moros et al., 2016). Indeed, the negatively loaded sites on Z_1 (blue curve, Fig. 3B) are clustered in the Baffin Bay and western Greenland region and also suggest cooling over this late-Holocene interval. However, because these records load negatively on Z_1 , their late-Holocene cooling is expressed in the factor score as an increase after ~3 ka. As such, the increasing Z_1 trend after 3 ka is likely an artefact of how cooling in the negatively loaded proxy time series is projected onto the shared Z_1 mode. More broadly, the spatial coherence of proxy temperature histories represented by the positive and negative W_1 loadings suggest that a spatially homogenous temperature history does not fully capture Holocene summer-climate variability across Greenland.

The proxy time series that are anti-correlated with Z_1 also provide meaningful regional climate information. These sites are concentrated in the Baffin Bay region (Fig. 3J), suggesting a coherent regional response to insolation. Within the oceanographic framework outlined above, we hypothesize that early-Holocene warmth, expressed across the positively loaded Z_1 sites, enhanced melting of the Greenland and Laurentide ice sheets (Fig. 3E), leading to meltwater discharge of cold fresh water, stronger stratification, and associated sea-ice feedbacks that muted or delayed marine warming in Baffin Bay (Gibb et al., 2015; Hansen et al., 2020; Jennings et al., 2011, 2014; Lloyd et al., 2005; Ouellet-Bernier et al., 2014; Perner et al., 2013; Saini et al., 2020). As insolation-driven summer warmth weakened through the middle Holocene, ice-sheet retreat slowed and the western Greenland Ice Sheet approached its minimum Holocene extent between ~5 and 3 ka (Briner et al., 2014; Young and Briner, 2015). As such, the dipolar W_1 spatial loading pattern (Fig. 3J) reflects two regional responses to the same peak summer-insolation forcing. In eastern and northern Greenland, high summer insolation produced a relatively direct warming response. Conversely, climate in the Baffin Bay region was dominated by the influence of ice-sheet freshwater inputs to Baffin



Bay, which were strong during warm periods when the icesheet retracted from its Holocene maxima and weakest when meltwater fluxes were weakest.

This negatively loaded Baffin Bay signal is not reproduced by any of the climate reconstruction products, which we attribute primarily to the model priors on which those products depend. In a data-assimilation framework, proxy information is extended beyond assimilated sites through the covariance structure of the prior (Hakim et al., 2016). A regional pattern that the prior does not generate therefore cannot be recovered at unconstrained locations, even where the assimilated network is dense elsewhere. The relevant pattern here – early-Holocene surface cooling in Baffin Bay driven by ice-sheet meltwater, stratification, and sea-ice feedbacks, occurring while summer temperatures elsewhere in Greenland were high – is not represented in the priors used by any of the products evaluated. Although both TraCE-21k and iCESM include coupled ocean–sea-ice physics and have been used to simulate meltwater forcing (Brady et al., 2019; Wan et al., 2026), that forcing is based on ICE-6G and not consistent with the most up-to-date information about GrIS deglacial evolution. Further, because the Osman, Badgeley, and Truax products draw priors from a mixture of meltwater and non-meltwater states across this interval, any regional meltwater signal would be further muted. We therefore interpret the absence of negative W_1 loadings in Baffin Bay as a limitation of the priors and note that it is reinforced by the fact that none of the products assimilate the Baffin Bay marine proxy time series in which the signal is expressed. Future model experiments that impose early-Holocene, glaciologically informed meltwater forcing in Baffin Bay would provide a direct test of whether these feedbacks can suppress local temperature while broader Greenland summer temperatures remain high.

South of the Davis Strait, within the Labrador Sea, three sea-surface temperature time series (Fig. 2, Sites 0, 1, and 4) have small loadings on both W_1 and W_2 , with larger positive loadings on W_1 (Fig. 3J) than on W_2 (Fig. 3K). All three time series indicate an early Holocene thermal maximum followed by generally lower temperatures through the middle and late Holocene (Gibb et al., 2015; Solignac et al., 2004; de Vernal et al., 2013; de Vernal and Hillaire-Marcel, 2006). This early Holocene thermal maximum could have resulted from LIS meltwater creating a shallow, low-salinity surface layer that warmed under peak boreal insolation (Gibb et al., 2014, 2015; Solignac et al., 2004; de Vernal and Hillaire-Marcel, 2006). These time series are therefore consistent with the insolation-driven temporal mode represented by Z_1 and indicate that ice-sheet meltwater had regionally distinct effects on summer temperatures, contributing to a muted temperature signal in Baffin Bay but promoting strong seasonal surface warming in the Labrador Sea.

4.3.2 Coherence between ocean circulation patterns and regional southern/southwestern Greenland summer climate

The cold early Holocene and substantial warming to peak summer temperatures at ~5.4 ka represented by Z_2 is consistent with previous syntheses identifying a later Holocene thermal maximum in southern and western Greenland relative to the northern and eastern region (Axford et al., 2021; Briner et al., 2016; Gajewski, 2015). Independent evidence from local glacier systems in southern Greenland further support a delayed thermal maximum in this region as glaciers document substantially smaller footprints between ~7.1 to 5.5 ka (Larocca et al., 2020). The broad temporal structure of Z_2 resembles the secondary temporal mode identified by Briner et al. (2016), which is characterized by mid-Holocene warmth and cool conditions during the early



and late Holocene, with strongest expression in southern Greenland and the Canadian Arctic archipelago. However, this thermal maximum in Z_2 occurs approximately 1-2 kyr earlier than in the Briner et al. (2016), suggesting a distinct timing of mid-Holocene thermal maxima in Greenland summer-temperature reconstructions compared to the broader Greenland-eastern Canadian Arctic region.

The generally positive but variable W_2 loadings in southern and southwestern Greenland (Fig. 3 K) indicate a regional mid-Holocene temperature pattern. This pattern is not described by the negative W_1 loadings in western Greenland and Baffin Bay, which may capture the suppression of marine warming associated with early-Holocene meltwater influence (Section 4.3.1). As such, the southern and southwestern Greenland thermal maximum may reflect auxiliary regional modulation. One potential explanation is that Atlantic-derived heat from the Irminger Current (IC) and its contribution to the West Greenland Current (WGC) became more effective at warming southern and southwestern Greenland during the middle Holocene. This oceanographic influence could help explain the later thermal maximum relative to the insolation-driven early Holocene thermal maximum in northern and eastern Greenland. The strongest positive W_2 loadings at Lake Rosaea and in Disko Bugt are consistent with this interpretation. Lake Rosaea's coastal position and predominantly southerly winds allow conditions in the Labrador Sea and Davis Strait to influence local temperatures through ocean-atmosphere heat exchange (Acharya et al., 2026), while the Disko Bugt sediment core records sea-surface conditions in a region influenced by the WGC (Ouellet-Bernier et al., 2014; Moros et al., 2016). Differences in exposure to ocean conditions could therefore explain the strong expression of Z_2 at these two sites and weaker associations elsewhere in the region.

Previous work provides a consistent picture of this regional oceanographic influence. In Disko Bugt, Perner et al. (2013) show that the WGC re-established by ~7.3 ka but remained fresh and meltwater-influenced until 6.2 ka, as indicated by changes in the proportion of sand (Fig. 3I). Thermal optimum-like conditions were interpreted in bottom waters from ~5.5 to 3.5 ka, owing to (1) reduced meltwater influence as the ice sheet became increasingly land-based (Young and Briner, 2015) and (2) an enhanced contribution of warm, saline IC waters to the WGC (Perner et al., 2013). Summer surface-water reconstructions from the same area also indicate warm conditions ($>10^\circ\text{C}$) were established by ~6 ka (Ouellet-Bernier et al., 2014). This timing suggests that surface-water warming preceded the benthic optimum, which is consistent with the increase in Z_2 prior to its ~5.4 ka maximum (Fig. 3F) and the ~6.0 ka onset of peak temperatures in quantitative time series with positive W_2 loadings (Fig. 3G). A multiproxy synthesis of the late Holocene change in Disko Bugt also identifies a similar warm interval from ~5.5 to 3.5 ka, characterized by reduced meltwater influence and strong IC contribution to the WGC, followed by cooling after 3.5 ka (Moros et al., 2016).

Farther south in Nuup Kangerlua, the fjord system near Nuuk, a distinct oceanographic regime was identified from ~7 to 3 ka, during which reduced GIS extent – near its Holocene minimum from ~5 to 3 ka (Larsen et al., 2015; Young and Briner, 2015) – coincided with enhanced entrainment of Atlantic-origin subpolar mode water into a fjord (Kvorning et al., 2025). Radiocarbon-dated subglacial methane provides independent evidence that the western Greenland Ice Sheet margin lay inland of its present position during the mid- to late Holocene, with ages of 4.2 to 1.4 ka in southwestern Greenland indicating subsequent ice-margin readvance over formerly exposed terrain (Hatton et al., 2026). Together, these data support the



interpretation that the Z_2 peak at ~ 5.4 reflects the time of maximum spatial coherence across the southern and southwestern Greenland coastal proxy network, which occurred during a broader thermal optimum extending from ~ 6 to 3.5 ka. The correspondence between the period of minimum GIS extent in western Greenland and the peak influence of Atlantic-derived water in Disko Bugt and Nuup Kangerlua suggests a positive feedback whereby reduced ice-sheet extent decreased meltwater discharge and increased the coastal sensitivity to Atlantic heat delivered by the WGC/IC system.

The rapid decline of Z_2 between 5 and 4 ka is consistent with the progressive decrease of boreal summer insolation, the onset of Neoglacial cooling, and a shift toward stronger East Greenland Current (EGC) influence along the West Greenland margin. Perner et al. (2013) and Moros et al. (2016) document cooling of WGC conditions from ~ 3.5 ka associated with increasing EGC contribution and coinciding with early glacier readvances in the Disko Bugt area, while Jennings et al. (2011) identify a similar cooling and Irminger Current weakening from ~ 3.5 ka on the southeastern Greenland Shelf (Fig. 3I). The timing of Z_2 cooling in our proxy network is broadly consistent with these records, though the Z_2 decline begins somewhat before the most pronounced neoglacial cooling recognized in individual records. This difference in timing may reflect the factor score capturing the period of maximum coherent warmth rather than warmth at the longest-persisting sites. We therefore interpret the second factor, Z_2/W_2 , as a regional coastal-ocean climate response in which the thermal response of southern and western Greenland was shaped by the evolving balance between GIS-derived meltwater discharge and Atlantic-origin heat delivered via the WGC and IC systems.

5. Conclusions

We use factor analysis on a Greenland proxy network to identify two dominant spatiotemporal factors of Holocene summer climate variability. The leading factor contrasts strong early Holocene warmth across northern and eastern Greenland from later warmth in western Greenland and Baffin Bay. We interpret the early warmth in northern and eastern Greenland as a response to peak early Holocene summer boreal insolation. In western Greenland and Baffin Bay, meltwater-driven surface stratification and sea-ice cover could have suppressed the temperature response to this insolation forcing, resulting in a delayed warming signal. The second factor captures a mid-Holocene thermal maximum in southern and southwestern Greenland, with its strongest expression in Disko Bugt. We interpret this pattern as reflecting a stronger surface and coastal influence of Atlantic-origin heat as the influence of proximal freshwater declined. Together, these factors suggest that Greenland's Holocene climate history comprised spatially coherent regional responses to interacting atmospheric, oceanographic, and cryospheric processes.

The climate reconstruction products generally reproduce the broad temporal modes associated with these factors more successfully than their spatial expression. Although the Truax and Badgeley et al. (2020) reconstructions provide the closest overall agreement with the proxy network, no product captures the spatial structure, particularly the contrasting Baffin Bay response and spatial pattern of middle-Holocene warmth. These results show that inter-reconstruction agreement in Greenland-wide temperature trends does not guarantee accurate representation of climate near the ice-sheet margin. Incorporating



660 additional coastal and marine proxies and improving the representation of freshwater, sea-ice, and ocean-circulation processes should therefore be priorities for future Greenland climate reconstructions and the ice-sheet simulations that use them.

Code and data availability

All code and data are available in a GitHub repository (<https://github.com/GerardOtiniano/Otiniano-et-al.-CotP-Repo>), which will be archived using Zenodo.

665 Author contributions

AdV, EJS, EKT, GAO, GJH, and OJT conceptualized this project, developed methodology, performed analysis, contributed to investigation and software. AdV, JPB, OJT, and SA contributed resources. AdV, EJS, EKT, GJH, JKC, and JPB acquired funding, offered project administration, and supervision. All author contributed to validation, visualization, and revision and editing of the manuscript.

670 Competing interests

The authors declare that they have no conflict of interest.

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