



Greenland temperature and precipitation over the last 20,000 years using data assimilation

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Abstract. Reconstructions of past temperature and precipitation are fundamental to modeling the Greenland Ice Sheet and assessing its sensitivity to climate. Paleoclimate information is sourced from proxy records and climate-model simulations; however, the former are spatially incomplete while the latter are sensitive to model dynamics and boundary conditions. Efforts to combine these sources of information to reconstruct spatial patterns of Greenland climate over glacial-interglacial cycles have been limited by assumptions of fixed spatial patterns and a restricted use of proxy data. We avoid these limitations by using paleoclimate data assimilation to create independent reconstructions of temperature and precipitation for the last 20,000 years. Our method uses information from long ice-core records and extends it to all locations across Greenland using spatial relationships derived from a transient climate-model simulation. Our reconstructions evaluate well against independent ice-core records. In addition, we find that the relationship between precipitation and temperature is frequency dependent and spatially variable, suggesting that thermodynamic scaling methods commonly used in ice-sheet modeling are overly simplistic. Our results demonstrate that paleoclimate data assimilation is a useful tool for reconstructing the spatial and temporal patterns of past climate on timescales relevant to ice sheets.

1 Introduction

Predicting the future behavior of the Greenland Ice Sheet requires understanding the sensitivity of the ice sheet to changes in temperature and precipitation (Bindschadler et al., 2013). One important constraint on this sensitivity is the response of the paleo ice-sheet to changing climate in the past (Alley et al., 2010). On glacial-interglacial timescales, temperature, not precipitation, appears to be the dominant control on the size of the Greenland Ice Sheet (Alley et al., 2010), as evidenced by the fact that the ice sheet is largest during cold and arid glacial periods and smallest during warm and wet interglacials. On these timescales, precipitation over the Greenland Ice Sheet scales positively with temperature (Robin, 1977), as anticipated by the Clausius-Clapeyron relation between temperature and saturation vapor pressure. However, ice-core records show that this thermodynamic relation is a poor approximation on annual to multi-millennial timescales (Kapsner et al., 1995; Fudge et al., 2016). For example, the GISP2 ice core from central Greenland shows that cooling coincided with increased snowfall between the early Holocene and present (Cuffey and Clow, 1997). Despite such evidence, ice-sheet modeling experiments typically



assume that precipitation consistently follows the thermodynamic, Clausius-Clapeyron relationship (e.g., Huybrechts et al.,
25 1991; Greve et al., 2011; Pollard and DeConto, 2012).

Ice-core records provide the best empirical estimates of climate history over the Greenland Ice Sheet. For temperature,
important proxies include oxygen and hydrogen isotopes of ice (e.g., Jouzel et al., 1997), nitrogen isotope ratios of gas trapped
in ice (e.g., Buizert et al., 2014; Kobashi et al., 2017), and borehole thermometry (e.g., Cuffey et al., 1995; Dahl-Jensen
30 et al., 1998). For precipitation, the thickness of annual layers of accumulated ice, corrected for thinning, is used (e.g., Dahl-
Jensen et al., 1993). Ice-core records, however, are limited in their spatial coverage. In contrast, climate-model simulations are
spatially-complete estimates of past climate, but they are subject to uncertainty in model dynamics and boundary conditions.

Efforts to combine information from proxy data and climate models have long been a part of ice-sheet modeling. The
most common approach is to scale the modern spatial pattern of temperature and precipitation using data from a single ice
core (e.g. Huybrechts et al., 1991; Huybrechts, 2002; Greve, 1997; Greve et al., 2011; Nielsen et al., 2018). This assumes a
35 fixed spatial pattern through time, which is unlikely to be valid. Recently, Buizert et al. (2018) used the average of the three
best-understood Greenland ice-core records to adjust the results of a transient climate-model simulation (the transient climate
evolution experiment, TraCE21ka; Liu et al., 2009; He et al., 2013). This approach allows for possible changes in spatial
relationships, but is restricted to a single climate-model realization, makes relatively little use of the available data from ice
cores outside northern Greenland, and provides no information on precipitation. Other attempts to incorporate more proxy data
40 have been limited to short time periods (e.g., Simpson et al., 2009; Lecavalier et al., 2014).

In this study we apply a novel method, paleoclimate data assimilation (paleo DA), to obtain a new, spatially-complete
reconstruction of Greenland temperature and precipitation. We focus on the last 20,000 years, which includes the end of the
last glacial period, the glacial to interglacial transition, and the Holocene thermal maximum (HTM), when temperatures at
the Greenland Ice Sheet summit reached 1-2 °C warmer than present (Cuffey and Clow, 1997; Dahl-Jensen et al., 1998). The
45 climate history, and the corresponding changes in the size of the ice sheet, are well-documented over this time period (Kaufman
et al., 2004; Young and Briner, 2015).

Paleo DA combines spatial information from a climate-model simulation and temporal data from proxy records to produce
a climate “reanalysis”, where the term is taken from the modern climate reanalysis methods on which paleo DA is based (e.g.,
Kalnay et al., 1996). We adapt the paleo DA framework developed by Hakim et al. (2016), who reconstructed annual two-
50 meter air temperature and 500 hPa geopotential height over the last millennium using a global network of temperature and
precipitation-sensitive proxy records. Here, we use oxygen isotopes of ice and layer thickness from ice cores to reconstruct
temperature and precipitation, respectively. We choose these proxies for their high temporal resolution, direct relationships
to climate over the ice sheet, and availability from multiple ice cores. For the climate-model simulation, we use TraCE21k
(Liu et al., 2009; He et al., 2013), which was run with the Community Climate System Model version 3 (CCSM3) to simulate
55 the last 21,000 years. We compare the resulting reanalysis to previously-published climate reconstructions (Sects. 3 and 5),
and assess the precipitation-temperature relationship (Sect. 3). We evaluate the reanalysis with independent proxy records and
sensitivity tests (Sect. 4).



2 Methods

Our paleoclimate reconstruction method assimilates oxygen isotope ratios and accumulation from ice cores with a transient
60 climate-model simulation to reconstruct the last 20,000 years of Greenland temperature and precipitation. In the following
subsections we discuss the ice-core data, the climate-model simulation, and the details of our paleoclimate data assimilation
approach.

2.1 Ice-core data

We use proxy records from eight ice cores from the Greenland Ice Sheet and nearby ice caps (Fig. 1, Table 1). As a proxy
65 for temperature, we use previously-published measurements of oxygen isotope ratios from the ice, which we discuss using
the conventional $\delta^{18}\text{O}$ nomenclature (Dansgaard, 1964). We note that while other proxies (such as borehole thermometry or
 $\delta^{15}\text{N}$ of N_2) have been used to produce temperature estimates (e.g., Cuffey et al., 1995; Dahl-Jensen et al., 1998; Buizert et al.,
2014; Kobashi et al., 2017), this has not been done at all our core locations; we instead rely on such data to obtain independent
estimates of error in the $\delta^{18}\text{O}$ -temperature relationship (see Sect. 2.3.3) and as comparisons to our resulting reanalysis (see Sect.
70 3). The accumulation history has been estimated for five of these cores from measured layer thickness corrected for vertical
ice-thinning due to dynamical strain in the ice sheet. We rely on previously-published accumulation histories for the GISP2
and NEEM cores (Cuffey and Clow, 1997; Rasmussen et al., 2013), and we estimate accumulation for the Dye3, GRIP, and
NGRIP cores using available layer-thickness data and simple ice-thinning calculations (see below and Sect. S1). We do not use
accumulation records from the Agassiz, Camp Century, or Renland cores because the ice-thinning history at these sites is not
75 adequately constrained. Most of the ice-core data are available at 50-year or higher resolution and have been synchronized to
a common depth-age scale (the Greenland ice core chronology 2005, GICC05; Andersen et al., 2006; Rasmussen et al., 2006;
Svensson et al., 2006; Vinther et al., 2006, 2008). All of these ice-core records extend from the beginning of the Holocene to
the present. Five $\delta^{18}\text{O}$ and four accumulation records also include the last glacial period. To evaluate the impact of the differing
lengths of these records, we produce a sensitivity reanalysis for which we assimilate just the fixed proxy-network (i.e., only
80 those data that span the full reanalysis time period, the last 20,000 years).

To extract the accumulation signal from measured layer-thickness, the layers must be destrained using assumptions about
the history of ice flow. For the Dye3, GRIP, and NGRIP cores, we use a one-dimensional ice-flow model (Dansgaard and
Johnsen, 1969) to calculate the cumulative vertical strain the layers have experienced at each core site. The Dansgaard-Johnsen
model requires specifying the vertical velocity at the surface and a kink height which determines the shape of the vertical
85 velocity profile. The velocity profile below the kink height approximates the influence of greater deformation rates in deeper
ice due to increased deviatoric stress and warmer ice temperature. For sites at the pressure-melting point at the bed, such
as NGRIP, we also implement the basal melt-rate (e.g., Dahl-Jensen et al., 2003). Previous work on many of the Greenland
ice cores has estimated cumulative vertical thinning assuming that the accumulation history scales with $\delta^{18}\text{O}$ and then found
optimal parameter values by comparing the modeled and measured depth-age relationships (Dahl-Jensen et al., 1993, 2003;
90 Rasmussen et al., 2014). Here, we wish to maintain independent determinations of the $\delta^{18}\text{O}$ and accumulation proxy records.



To do this, we select reasonable ice-flow parameters independently, based on the glaciological setting of each site; specifically, we use kink-height values of 0.1-0.2 for flank flow and 0.4 for ice flow near ice divides where the deviatoric stress is low (Raymond, 1983; Conway et al., 1999). Where available, we use published values or kink-height values that result in a good match to published accumulation records (Dahl-Jensen et al., 2003; Gkinis et al., 2014). Based on this range of plausible ice-
95 flow parameters, we develop three scenarios for each site: "low", "moderate", and "high", where the names reflect the relative magnitude of accumulation in the glacial and early Holocene. We assimilate the intermediate-value ("moderate") accumulation records to produce our main precipitation reanalysis, while we assimilate the high and low accumulation records into high and low sensitivity scenarios, respectively, to provide a conservative estimate of uncertainty. Descriptions of the rationale for the parameter choices at each site are given in Sect. S1. Our method to estimate accumulation should be most accurate for
100 the interior ice cores (i.e., GISP2, GRIP, NEEM, NGRIP); these sites are thicker and have lower accumulation rates such that layers of the same age have experienced less cumulative strain than for the more coastal cores (i.e., Agassiz, Camp Century, and Renland). We do not attempt to reconstruct accumulation from these coastal cores because the layers cannot be destrained with sufficient accuracy. Dye3 suffers from the same challenges; however, it is the only ice core with long-term climate data south of 70°N (Fig. 1). Thus, we include the Holocene Dye3 accumulation rates despite the greater uncertainty relative to the
105 interior cores.

Because records from the Dye3 ice core are our only source of information in southern Greenland, we take the following steps to increase the data available for assimilation. The Dye3 record has not been previously assigned a depth-age scale beyond 11.7 ka (throughout this paper, "ka" refers to thousands of years before 1950 CE). We extend the depth-age scale to 20 ka to take advantage of the glacial portion of the $\delta^{18}\text{O}$ record. To do this, we match the $\delta^{18}\text{O}$ record from Dye3 to the
110 $\delta^{18}\text{O}$ record from NGRIP using the cross-correlation maximization procedure from Huybers and Wunsch (2004) (Sect. S2). We interpolate the glacial $\delta^{18}\text{O}$ record from Dye3 (which, as measured, has an average resolution of only 85 years) to the same 50-year resolution used for our other ice-core records. Extension of the Dye3 depth-age scale also provides a layer-thickness record from 20 ka to present; however, we do not use accumulation data from Dye3 for the period 20-11.7 ka because the low resolution impedes our ability to estimate accumulation variations from layer thickness. Using this depth-age scale extension
115 for Dye3 may introduce error that is difficult to quantify; however, we find that including Dye3 has an important impact on the resulting reanalysis of southern Greenland climate (Sect. S3).

Where possible, we account for non-local effects on the ice-core records. The global-mean $\delta^{18}\text{O}$ of seawater fluctuates with global ice-volume, while on the regional scale, horizontal advection brings ice from other elevations and latitudes. We correct for changes in the oxygen-isotope composition of seawater following the methods of Stenni et al. (2010), using the benthic
120 foraminifera dataset from Bintanja et al. (2005). For ice cores in regions of high horizontal advection, we correct for elevation and latitude differences between the site of snow deposition and the ice-core site. Following the methods from Dahl-Jensen et al. (2013), we apply corrections for advection-caused elevation changes in the $\delta^{18}\text{O}$ records from Camp Century, Dye3, and NEEM and for advection-caused latitude changes in the $\delta^{18}\text{O}$ record from NEEM (Vinther et al., 2009; Dahl-Jensen et al., 2013). We do not correct the accumulation records for advection from upstream because the elevation-accumulation
125 relationship is complicated by the prevailing wind direction (Roe and Lindzen, 2001) and the thinning function uncertainties



are likely to be larger than the effects of ice advection. We also do not correct $\delta^{18}\text{O}$ or accumulation for changing elevation at the ice-core site; our goal is to reconstruct conditions at the surface, rather than at a constant reference elevation. We take the anomaly of each corrected $\delta^{18}\text{O}$ record and the ratio of each accumulation record relative to the mean of all data in the record that falls within the time period 1850-2000 CE. We then average each record to 50-year resolution, the lowest resolution in these records (with the exception of $\delta^{18}\text{O}$ from the glacial period in the Dye3 core). It is these corrected, averaged records that we use in the data assimilation (Figs. 2 and 3).

2.2 Climate-model simulation

We use TraCE21ka, a simulation of the last 21,000 years of climate, which was run with the fully-coupled CCSM3 and transient ice-sheet, orbital, greenhouse gas, and meltwater flux forcings (Liu et al., 2009, 2012; He et al., 2013). For paleoclimate data assimilation, it is important that the climate simulation capture a range of possible climate states over the time period of interest. By design, TraCE21ka captures the major glacial-to-Holocene temperature changes, as well as short-term, rapid climate changes (i.e., the transitions into and out of the Bølling-Allerød and Younger Dryas events). In contrast, many other transient climate simulations cover only the last millennium and have too little variability to capture the range of climate states across the glacial-interglacial transition. Furthermore, TraCE21ka includes changes in orbital forcing and therefore changes in the seasonality of temperature and precipitation, which strongly influence the relationship between $\delta^{18}\text{O}$ and temperature (Steig et al., 1994; Werner et al., 2000). TraCE21ka also includes transient ice sheets, which is important for capturing the influence of elevation change on the ice-core records. We note, however, that the ice sheets in TraCE21ka are independent of climate, updated only every 500 years during the simulation, and taken from ICE-5G (Peltier, 2004), a now outdated ice-sheet reconstruction (Roy and Peltier, 2018). From TraCE21ka, we use two-meter air temperature for temperature (T) and the sum of large-scale stable precipitation and convective precipitation for precipitation (P). We take the anomaly of temperature and the fraction of precipitation relative to the mean for 1850-2000 CE. We then average to 50-year resolution, as for the ice-core records.

2.3 Paleoclimate data assimilation

To combine the ice-core data and climate-model data, we use an offline data assimilation method similar to that described in Hakim et al. (2016). “Offline” refers to the absence of a forecast model that evolves the climate state between time steps, such that in offline data assimilation the same initial climate state is used for every time step. The offline method is appropriate when characteristic memory in the system is significantly shorter than the time step (Hakim et al., 2016, and references therein). In our case, each time step is an average over 50 years, as dictated by the resolution to which we average the proxy records.

Our paleoclimate data assimilation framework uses ensembles for the initial (prior) and final (posterior) estimates of the climate state, providing a probabilistic framework for interpreting and evaluating the results. To compute the posterior ensemble, we apply the Kalman update equation (Whitaker and Hamill, 2002), which spreads new information gained from proxy records



to all locations and variables in the prior ensemble:

$$\mathbf{x}_a = \mathbf{x}_b + \mathbf{K}(\mathbf{y} - \mathcal{H}(\mathbf{x}_b)) \quad (1)$$

where bold lowercase letters are vectors, bold capital letters are matrices, and script capital letters are mapping functions. The
160 posterior ensemble is \mathbf{x}_a , \mathbf{x}_b is the prior ensemble, \mathbf{y} is the proxy data, and \mathcal{H} is the function that maps from the prior variables
to the proxy variables. \mathbf{K} is the Kalman gain matrix:

$$\mathbf{K} = \mathbf{B}\mathbf{H}^T(\mathbf{H}\mathbf{B}\mathbf{H}^T + \mathbf{R})^{-1} \quad (2)$$

where T indicates a matrix transpose, \mathbf{B} is the covariance matrix computed from the prior ensemble, \mathbf{H} is the linearization of
165 \mathcal{H} about the mean value of the prior, and \mathbf{R} is a diagonal matrix containing the error variance for each proxy record, the use of
which requires an assumption that error covariances between proxy records are negligible.

To compute the new information gained from the proxy records, the prior ensemble must first be mapped into proxy space
to get prior estimates of the proxy ($\mathcal{H}(\mathbf{x}_b)$). This mapping (\mathcal{H}) is the proxy system model (PSM). Our PSM for the $\delta^{18}\text{O}$ -
temperature relationship is a linear function and for accumulation is a direct comparison between ice-core-derived accumulation
and precipitation from the prior, which is interpolated from the climate-model grid to the geographic location of the ice core.
170 These PSMs are detailed in Sects. 2.3.1 and 2.3.2. Comparing the prior estimates of the proxy ($\mathcal{H}(\mathbf{x}_b)$) to the proxy data (\mathbf{y})
yields the innovation ($\mathbf{y} - \mathcal{H}(\mathbf{x}_b)$), the new information gained from the proxy records.

The Kalman gain (\mathbf{K} , Eq. 2) weights the innovation by the relative magnitude of the ensemble covariance of the prior
estimates of the proxy ($\mathbf{H}\mathbf{B}\mathbf{H}^T$), and the error covariance of the proxy records (\mathbf{R}). The Kalman gain spreads the weighted
innovation to all locations and variables in the prior ensemble, using the covariance structure ($\mathbf{B}\mathbf{H}^T$) from the prior ensemble.

175 The prior ensemble is an initial estimate of possible climate states, which we form using 100 randomly-chosen 50-year
averages from the TraCE21ka simulation. Proxy records are assimilated into the prior using the Kalman update (Eq. 1), which
produces the posterior ensemble, a new estimate of possible climate states. We assimilate $\delta^{18}\text{O}$ to reconstruct temperature
and separately assimilate accumulation to reconstruct precipitation. This process is repeated over multiple iterations, with each
iteration using one of ten different 100-member prior ensembles and excluding one proxy record. Each proxy record is excluded
180 from a total of ten iterations, where each of these iterations uses a different one of the ten prior options. For a reanalysis, the
total number of iterations is thus ten times the number of proxy records, such that for temperature we have 80 iterations and
for precipitation we have 50 iterations. A reanalysis is a compilation of the posterior ensembles from these iterations, resulting
in a temperature reanalysis having 8,000 ensemble members and a precipitation reanalysis having 5,000 ensemble members.

The proxy record that is excluded from an iteration is independent of that iteration's posterior ensemble, such that we can
185 evaluate the posterior against this record. With our PSMs, we convert the posterior ensemble into predictions of the independent
record using the mapping \mathcal{H} and compare these predictions to the record along the time axis. We use four skill metrics to
evaluate different aspects of the predictions. The correlation coefficient (corr ; Eq. 3) measures the relative timing of signals in
the predictions and the proxy record:

$$\text{corr} = \frac{1}{n-1} \sum_{i=1}^n \left(\frac{(y_i - \bar{y})(v_i - \bar{v})}{\sigma_y \sigma_v} \right) \quad (3)$$



190 where v is the ensemble mean of the predicted values, y is the proxy record value, an overbar indicates a time mean, n is the number of time steps, and σ is the standard deviation of the variable in the subscript. The coefficient of efficiency (CE; Eq. 4) (Nash and Sutcliffe, 1970) is affected by signal timing as well as signal amplitude and mean bias:

$$CE = 1 - \frac{\sum_{i=1}^n (v_i - y_i)^2}{\sum_{i=1}^n (v_i - \bar{v})^2} \quad (4)$$

The root mean squared error (RMSE; Eq. 5) gives an intuitive sense for the magnitude of the differences between the predictions
195 and proxy record:

$$RMSE = \left(\frac{1}{n} \sum_{i=1}^n (y_i - v_i)^2 \right)^{1/2} \quad (5)$$

The ensemble calibration ratio (ECR; Eq. 6) indicates whether the ensemble has enough spread (uncertainty) given the error in the ensemble mean (e.g. Houtekamer et al., 2005):

$$ECR = \frac{1}{n} \sum_{i=1}^n \left(\frac{(y_i - v_i)^2}{var(\mathbf{v}_i) + r} \right) \quad (6)$$

200 where \mathbf{v} is a vector of the ensemble members of the predicted values, r is the error variance for the proxy record (y), and var indicates the variance. Accordingly, if the ensemble variance is appropriate for the amount of error, then the ensemble calibration ratio is near unity.

We compute all four skill metrics for both the posterior and prior ensembles, which shows whether assimilation of proxy records results in an improved estimate of the climate state over our initial estimate. We define improvement as correlation
205 coefficient closer to 1, CE closer to 1, RMSE closer to 0, and ensemble calibration ratio closer to 1. We anticipate that our reanalysis will show improvement over the prior because the prior is constant in time and contains no information about temporal climate variations; however, improvement is not guaranteed, especially if proxy records contain highly-localized climate signals or if the prior covariance structure is unable to appropriately spread information from the proxy sites to other locations.

210 We evaluate results over three time periods: 1) the full overlap between the reanalysis time period and the proxy record, 2) a time representative of the glacial, 20-15 ka, and 3) a time representative of the Holocene, 8-3 ka. Some proxy records overlap the full reanalysis period, 20-0 ka, while others overlap just the Holocene, 11.7-0 ka. The skill metrics computed for these two groups should be considered separately.

To remove mean bias from temperature, we subtract out the reference-period mean. For precipitation, we divide by the
215 reference-period mean. It is these bias-corrected results that are referred to unless noted otherwise.

2.3.1 Proxy system model: $\delta^{18}\text{O}$

The isotopic composition of precipitation, as recorded in ice cores, is highly correlated with temperature at the time and location of deposition, but is also sensitive to conditions at the moisture source region (i.e. sea surface temperature and relative humidity at the ocean surface; e.g., Johnsen et al., 1989). Moisture-source conditions primarily affect the deuterium excess,



220 which we do not use here, and the influence on $\delta^{18}\text{O}$ is comparatively weak (e.g., Armengaud et al., 1998). For our $\delta^{18}\text{O}$ PSM, we use a linear relationship with temperature at the ice-core drill site (T_{site}) that has a slope of $0.67 \pm 0.02 \text{‰} \text{°C}^{-1}$, which was calibrated using modern, spatial data (Johnsen et al., 1989). It is well known that this modern, spatially-derived slope does not necessarily apply to temporal $\delta^{18}\text{O}$ - T_{site} relationships, which have effective slopes that are time, frequency, and location dependent. Temporal changes in precipitation seasonality, inversion-layer thickness, and source region conditions introduce
225 nonlinearity into the effective $\delta^{18}\text{O}$ - T_{site} relationship (e.g., Jouzel et al., 1997; Pausata and Löfverström, 2015). Diffusion in the firn column also affects this relationship, but it is negligible for annual and longer timescales at the locations of the ice cores we use (Cuffey and Steig, 1998). Borehole thermometry at the GISP2 and GRIP sites show that for the low-frequency changes associated with the last glacial-interglacial transition, the temporal slope is less than half the modern, spatial slope (Cuffey and Clow, 1997; Jouzel et al., 1997). Numerous studies have suggested that precipitation seasonality is the largest source of
230 nonlinearity in the $\delta^{18}\text{O}$ - T_{site} relationship (e.g., Steig et al., 1994; Pausata and Löfverström, 2015); changes in precipitation seasonality are thought to be the primary reason that the effective $\delta^{18}\text{O}$ - T_{site} relationship for the glacial-interglacial transition has such a low slope (Werner et al., 2000).

We rely on TraCE21ka to estimate the site-specific effects of precipitation seasonality on the $\delta^{18}\text{O}$ - T_{site} relationship. Site-specific effects can also be estimated using independent temperature reconstructions, e.g. from borehole thermometry or $\delta^{15}\text{N}$
235 measurements; however, such independent reconstructions for the last 20,000 years exist only for a few of the long ice-core records, GISP2, GRIP, and NGRIP (Buizert et al., 2018; Dahl-Jensen et al., 1998; Gkinis et al., 2014), limiting the utility of such records to capture spatial variability.

To incorporate estimates of the site-specific effects of precipitation seasonality from TraCE21ka, we adjust the linear $\delta^{18}\text{O}$ PSM by replacing T_{site} with T_{site}^* , the precipitation-weighted temperature at the ice-core drill site. We compute T^* across
240 Greenland using T and P from TraCE21ka:

$$T^* = \sum_{i=1}^{n=12} \left(T_{mon} \frac{P_{mon}}{P_{ann}} \right) \quad (7)$$

With T_{site}^* in our PSM, we find that the $\delta^{18}\text{O}$ - T_{site} slope is spatially variable, but on average it is about 75% of modern. Thus, in using TraCE21ka, we capture the mean of the modern (i.e. high-frequency) relationship and that of the glacial-interglacial (i.e. low-frequency), effective temporal-slope.

245 To evaluate the sensitivity of our results to the choice of PSM, we produce four other reconstructions (S1-S4). The S1 scenario uses the PSM, $\delta^{18}\text{O} = 0.67T_{site}$, which is the modern (high-frequency) relationship and does not account for precipitation seasonality. The S2 scenario uses $\delta^{18}\text{O} = 0.5T_{site}$, the mean of the modern and glacial-interglacial temporal slopes. The S3 scenario uses $\delta^{18}\text{O} = 0.335T_{site}$, which is similar to published estimates of the glacial-interglacial temporal slope (half the modern slope) (Cuffey and Clow, 1997; Jouzel et al., 1997). By lowering the slope in the S2 and S3 scenarios, we implicitly account
250 for precipitation seasonality; however, in these scenarios and S1, the $\delta^{18}\text{O}$ - T_{site} relationship is spatially uniform, whereas it is spatially variable in the main reanalysis because we include the spatial pattern of precipitation seasonality. The S4 scenario uses the same PSM as in the main reanalysis, $\delta^{18}\text{O} = 0.67T_{site}^*$, but we adjust the strength of the precipitation seasonality in TraCE21ka such that the average $\delta^{18}\text{O}$ - T_{site} slope around Greenland is approximately $0.335 \text{‰} \text{°C}^{-1}$. The S4 scenario thus



has the same spatially-variable $\delta^{18}\text{O}-T_{site}$ relationship as in the main reanalysis, but with a greater influence of precipitation
255 seasonality.

2.3.2 Proxy system model: Accumulation

Accumulation is closely related to total precipitation at our ice-core sites, which have limited surface melting. Simulations from
the regional climate model HIRHAM5 show that for modern climate at the GISP2, GRIP, NEEM, and NGRIP ice-core sites,
surface mass balance, snowfall, and precipitation are all within 1.6 cm water equivalent (w.e.) when averaged over the years
260 1989 to 2012 CE (Langen et al., 2015, 2017). For this reason, and because TraCE21ka lacks process-based ablation variables,
our PSM is a direct-comparison between ice-core accumulation and simulated precipitation at the model grid-cell closest to
the ice-core site.

This direct-comparison PSM may be an incomplete model of the accumulation-precipitation relationship at the Dye3 ice-
core site; regional climate simulations show that modern surface mass balance is lower than precipitation due to melt rates that
265 average 84 cm w.e. year⁻¹. Significant melt rates would cause the spatial covariance structure of accumulation across these
sites to differ from that of precipitation; however, both models and observations lack the necessary variables or duration to
show the extent to which this difference exists for 50-year timescales through the last glacial-interglacial cycle. We emphasize
that we use relative, rather than absolute changes in the data assimilation, to account for the mean bias between precipitation
and accumulation.

270 2.3.3 Proxy error variance estimation

In the Kalman filter (Eq. 1), the diagonal elements of \mathbf{R} contain the error variance of each proxy record, which includes how
we model the proxy (the PSM). We compute representative error variances for $\delta^{18}\text{O}$ and accumulation, and apply them to all
records and time slices. We do not include error associated with corrections applied to the ice-core data (Sect. 2.1) or associated
with the accumulation PSM (Sect. 2.3.2) because we cannot characterize the statistical properties of these errors.

275 A universal, but typically small, error source is from the measurement of proxies. For $\delta^{18}\text{O}$, measurement error is equivalent
to laboratory precision. We compute a representative measurement error from the GISP2 ice core, for which a single mea-
surement of $\delta^{18}\text{O}$ has a laboratory precision (variance) of 0.024 ‰² (Stuiver and Grootes, 2000). Assuming independent error
and annual measurements, the 50-year average error variance reduces to 0.0034 ‰², which is insignificant compared to other
sources of error. For accumulation, the measurement error is from the measurement of layer thickness, which is related to the
280 error in annual-layer counts per unit depth. Again, we assume GISP2 is a representative core and estimate the layer-thickness
error from Table 3 in Alley et al. (1997), which provides repeat annual-layer counts for several depth intervals. From this table,
we find the standard deviation of counted years in each depth interval, divide by the average number of years, average across all
depth intervals, and square the result. This computation results in a layer-thickness error variance of 0.0015, a unitless number
due to our use of fractional accumulation records.

285 Another source of error is the extent to which a model grid-cell may misrepresent a point proxy-measurement. In the
innovation, there is an implicit assumption that the proxy and the prior estimate of the proxy are representative of the same



processes. However, an ice core is about 100 cm^2 , an area that is affected by processes at all scales, from regional change to local, sub-meter-scale topography, while a model grid-cell in TraCE21ka can cover tens of thousands of square kilometers, an area that is affected by only the largest scales, from global to regional. Thus, there is an inherent inability of the prior to represent local processes at the ice-core site, which we refer to as the spatial representation error. To estimate this error, we compute the variance of the local noise (e.g., Reeh and Fisher, 1983) using the GISP2 and GRIP ice-core records, which are located about 30 km apart within the same model grid cell. For $\delta^{18}\text{O}$, our estimate is 0.21 ‰^2 which is about half the value determined by Fisher et al. (1985) at several locations around Greenland. Our estimate is relatively conservative, considering that we are using 50-year averages rather than annual averages as in Fisher et al. (1985). For accumulation, we estimate a spatial representation error variance of 0.0023.

A third source of error is the extent to which the $\delta^{18}\text{O}$ PSM may be an inaccurate model of the $\delta^{18}\text{O}$ - T_{site} relationship. The less accurate the PSM, the less weight that should be given to the innovation. We estimate PSM error variance by calculating the mean squared error (MSE) between a $\delta^{18}\text{O}$ record and an independent temperature record mapped to $\delta^{18}\text{O}$ using the PSM, $\delta^{18}\text{O} = 0.67T_{\text{site}}$. We use independent datasets taken from the GISP2 ice core: the $\delta^{18}\text{O}$ record and three $\delta^{15}\text{N}$ -derived temperature estimates for the Holocene, a mean estimate and the two-standard-deviation uncertainty bounds (Kobashi et al., 2017). From these datasets, we estimate three PSM error variances that range from 0.56 to 1.1 ‰^2 , from which we choose the largest.

For the assimilation of each $\delta^{18}\text{O}$ record, we use an estimated total error variance of 1.3 ‰^2 , which is a sum of the measurement, spatial representation, and PSM error variances. For the assimilation of each accumulation record, we use an estimated total error variance of 0.0038, which is a sum of the measurement and spatial representation error variances.

3 Results

Through the assimilation of ice-core data with a prior ensemble that is constant in time, we produce a spatially-complete Greenland temperature and precipitation reanalysis (Figs. 4 and 5). Here we focus on results relevant to the evolution and sensitivity of the Greenland Ice Sheet, including the late glacial anomaly, the Holocene thermal maximum (HTM), and the relationship between temperature and precipitation.

In our reanalysis, late glacial (20-15 ka) mean-temperature anomalies range from about -20 °C in northern Greenland to less than -10 °C in southern Greenland (Fig. 4c). At the GRIP and GISP2 ice-core sites, the reanalysis has a -14 °C anomaly with a standard deviation of 2 °C . This is in excellent agreement with the mean-temperature anomaly of -14 °C for the same period at the GISP2 site, which was derived from $\delta^{18}\text{O}$ calibrated with borehole thermometry (Cuffey et al., 1995; Cuffey and Clow, 1997). Average late-glacial precipitation in the reanalysis ranges from a third to half of modern with the highest values on the coasts around southern Greenland (Fig. 4d).

Our reanalysis shows warmest temperatures occurred across Greenland between 7 and 3 ka, reaching a temperature maximum around 5 ka (Fig. 5). Although this timing tends to be later than many estimates of the HTM, it lies within the ranges reported in the literature; for example, a summary of proxy records from around Greenland shows peak warmth usually oc-



320 curring around 9-5 ka, but also as early as 10.8 ka and as late as 3 ka (Kaufman et al., 2004). Borehole thermometry shows
that temperatures peaked around 6-7.7 ka at Summit and 4.5 ka at the Dye3 ice-core site (Cuffey and Clow, 1997; Dahl-Jensen
et al., 1998; Kaufman et al., 2004). Temperature estimates from $\delta^{15}\text{N}$ of N_2 show an earlier HTM peak at Summit around 8
ka (Kobashi et al., 2017). In northwest Greenland, $\delta^{18}\text{O}$ measurements from lake sediments show the HTM starting before 7.7
325 et al., 2018).

Mean-annual HTM temperature anomalies in our reanalysis range from nearly 2 °C in northern Greenland to about +1 °C in
southern Greenland (Fig. 4a). Similar to the late-glacial temperature anomalies, the pattern of the HTM is dominated by a north-
south trend that has the greater temperature changes to the north, especially in northwest Greenland. While this spatial pattern
agrees well with previous studies which have noted especially warm Holocene temperature anomalies in northwest Greenland
330 (Lasher et al., 2017; Lecavalier et al., 2017; McFarlin et al., 2018), many estimates of HTM anomalies around Greenland are
higher than our reanalysis indicates. Our low temperature estimates (compared to previous work) may be in part due to our
reconstructing the annual mean rather than the summer mean; the greatest temperature anomalies in the HTM are thought
to have occurred in the summer months and many proxies are more sensitive to summer than annual temperature (Kaufman
et al., 2004). Importantly, in our reanalysis, the higher HTM temperatures do not translate to a marked increase in precipitation
335 as would be expected from a thermodynamically-scaled relationship between temperature and precipitation. Instead, we find
fractional precipitation within 2% of modern values (1.0 ± 0.02) during the HTM (Fig. 4b), with slightly higher-than-modern
precipitation in central Greenland and slightly lower-than-modern precipitation in northwestern Greenland.

Our results allow us to investigate the relationship between temperature and precipitation. To facilitate comparison with
thermodynamic scalings widely used by ice-sheet models, we define the relationship as exponential and find the best fit for our
340 reanalysis:

$$P_{fraction} = \frac{P_{past}}{P_{modern}} = e^{\beta(T_{past} - T_{modern})} \quad (8)$$

where P is the precipitation rate, T is the temperature, and β is a scaling factor (Greve et al., 2011). For a given temperature
change, a higher value of β results in a larger change in precipitation (orange in Fig. 6a). In ice-sheet models that use this
scaling, it is commonly applied with a uniform β value for all locations (e.g., Huybrechts et al., 1991; Huybrechts, 2002;
345 Greve, 1997; Greve et al., 2011; Pollard and DeConto, 2012; Cuzzone et al., 2019). We find that our best-fit scaling-factors
(β) center on the Greve et al. (2011) value of 0.07 for locations around Greenland, but our scaling factors tend to be larger
($\beta > 0.10$) where late-glacial precipitation is lowest and smaller ($\beta < 0.05$) where late-glacial precipitation is highest (Figs. 4d
and 6a).

The precipitation-temperature relationship in our reanalysis is driven by the assimilated ice-core records, though the spatial
350 pattern of this relationship is also influenced by the spatial covariance structures of the prior ensembles. Previous work with
ice-core records has found that the relationship between temperature and precipitation is frequency dependent, with a stronger
relationship at lower frequencies (Cuffey and Clow, 1997); as expected, there is a similar frequency-dependence in our reanaly-
sis. We find that an exponential scaling captures the low-frequency glacial to Holocene precipitation change; however, this fails



at higher frequencies (Fig. 6b-e). To evaluate this frequency-dependence, we filtered our results using 6th order, low-pass and
355 high-pass Butterworth filters with $5,000 \text{ year}^{-1}$ cutoff frequencies. The low-pass filtered dataset shows the same precipitation-
temperature relationship as the unfiltered dataset (Fig. 6c-d), while the high-pass filtered dataset shows that precipitation is
less sensitive to changes in temperature (i.e., the value of β is lower) at these higher frequencies (Fig. 6e). This decoupling of
temperature and precipitation is apparent in the amplitude difference of high-frequency signals in the glacial and the Holocene.
In our temperature reanalysis, we find that high frequencies in the glacial have a greater amplitude than those in the Holocene,
360 while in our precipitation reanalysis, we find the opposite. A single scaling, as is typically used in ice-sheet modeling, cannot
capture this difference.

4 Evaluation

4.1 Independent proxy evaluation

Here we evaluate our results against proxy records that are excluded from ten of the iterations that make up the temperature
365 and precipitation reanalyses. For this evaluation, we use the raw results (without a mean-bias correction). We find, however,
that the mean biases are small relative to climate changes over the last 20,000 years; there is little difference between our
bias-corrected and uncorrected results and it is unlikely that the mean bias has a large affect on our evaluation.

Evaluation against independent proxy data shows that our reanalysis captures the timing and magnitude of low-frequency
climate changes (Figs. 7 and 8) and is an improvement over the prior ensemble (Figs. S3 and S4). Evaluation over the full 20,000
370 years of the temperature and precipitation results shows high, positive correlation coefficients (ranging from a minimum of 0.97
to maximum of 0.99), which indicate that the reanalysis captures both the timing and sign of climate events, while high CE
(0.87-0.98) and low RMSE values (0.62-1.2‰ for $\delta^{18}\text{O}$ and 0.04-0.08 for accumulation) indicate that the reanalysis captures
the magnitude of these events. Our skill during this longest evaluation period is primarily due to the presence of low-frequency
climate changes, which tend to be coherent across Greenland, such that evaluation over this full 20,000-year period shows more
375 skill than evaluation over the full Holocene, which shows more skill than evaluation over just 5,000 years in the Holocene (or
5,000 years in the glacial) (Figs. 7 and 8). Even for these shorter evaluation periods, which are dominated by high-frequency,
spatially-incoherent noise, the reanalysis shows overall improvement over the prior ensemble (Figs. S3 and S4).

For all evaluation periods and both variables, the ensemble calibration ratio (ECR) for the prior is skewed towards values
greater than unity (0.66-8.7), which suggests that the prior ensemble tends to have too little spread. Conversely, for the posterior,
380 the ERC is generally less than unity (0.10-1.7) (Figs. 7 and 8), suggesting that the posterior ensemble has more than enough
spread given the error in the ensemble mean (as compared to the proxy records). This result indicates that the reanalysis
ensemble encompasses the climate as recorded by the proxy records for most times and locations over the last 20,000 years.



4.2 Sensitivity evaluation

Proxy networks that change in time, such as ours, can introduce artificial discontinuities into the reanalysis, especially if the
385 number of proxies is low or the proxy uncertainty values are inappropriate. We produce another reconstruction with a fixed
proxy network, in which all assimilated proxy records participate in every time step in the reconstruction (see Sect. 2.3). A
comparison of these results with our main reanalysis shows no apparent discontinuities for the ensemble mean and 5th to 95th
percentiles at Summit (Fig. 9) or other locations around Greenland.

Our results are sensitive to the $\delta^{18}\text{O}$ - T relationship. To test this, we compare the main reanalysis to the four scenarios (S1-
390 S4, as described in Sect. 2.3.1) that use different $\delta^{18}\text{O}$ PSMs, each of which assumes a different slope and spatial pattern of
the $\delta^{18}\text{O}$ - T relationship. We show this comparison for Summit as an example (Fig. 10), but the findings are applicable for
all locations. As discussed previously, scenarios S1-S3 all assume that the $\delta^{18}\text{O}$ - T relationship has a uniform spatial pattern,
but they each assume a different influence of precipitation seasonality. From these scenarios, we find that the temperature
reconstruction is sensitive to the assumed precipitation seasonality, especially in the glacial where a stronger seasonality results
395 in a greater glacial temperature anomaly. At Summit, this difference is nearly 10 °C between S1, which assumes no influence,
and S3, which assumes the most influence of precipitation seasonality (Fig. 10). Similarly, the main reanalysis and S4 scenario
both assume that the $\delta^{18}\text{O}$ - T relationship has a spatially-variable pattern, but S4 assumes a greater influence of precipitation
seasonality. Again we find that the results are sensitive to assumed seasonality, with the greatest impact on the glacial-to-
interglacial change.

400 We also find that the temperature results are sensitive to the spatial pattern of the $\delta^{18}\text{O}$ - T relationship. To test this, we
compare the results from the S1-S3 scenarios that assume a spatially-uniform pattern to results from the main reanalysis and S4
scenario that assume a spatially-variable pattern. The S1-S3 scenarios have a characteristic shape to their time series (Fig. 10),
and, although the main reanalysis and S4 scenario generally fit this characteristic shape in the glacial and middle-late Holocene,
in the early Holocene the main reanalysis and S4 diverge and show slower warming trends than the S1-S3 scenarios. This
405 indicates that there is new information added by using a PSM that accounts for spatial variability in precipitation seasonality.

For precipitation, we find that the results are sensitive to which accumulation record is assimilated at each ice-core site. As
explained in Sect. 2.1 and S1, we use a low, moderate, and high accumulation record for most of the ice-core sites to produce
the low, main, and high precipitation scenarios, respectively (Fig. 11). The largest spread among the scenarios is in the earliest
part of the reconstruction, i.e. the last glacial through the early Holocene, since uncertainties in the ice thinning history have the
410 greatest impact at greater depths (and hence, greater ages). There is also a larger spread among the scenarios at more southern
locations because the accumulation record at Dye3 has both the most influence on southern Greenland (Sect. S3) and the largest
uncertainty in the ice thinning history (Fig. 3, Sect. S1).

5 Discussion

Using paleoclimate data assimilation to combine ice-core records and the TraCE21ka climate-model simulation, we have
415 obtained a Greenland climate reanalysis for the last 20,000 years that both captures the timing and magnitude of major climate



events and shows good evaluation against independent proxy records. Our main reanalysis is one of very few spatially-complete time series of Greenland climate over the last 20,000 years. Here, we compare our reanalysis with other Greenland climate histories, and suggest that, together, they should be treated as an ensemble of climate boundary conditions that can be used to produce ensembles of ice-sheet model simulations. These climate histories can also be further evaluated using a combination
420 ice-sheet models and independent constraints from the glacial-geologic record of past ice-sheet configurations.

We compare our results with the recent reconstruction of Buizert et al. (2018), hereafter referred to as the B18 reconstruction. The B18 temperature reconstruction was produced by adjusting a part of the TraCE21ka temperature field that is affected by changes in the Atlantic meridional overturning circulation such that the full temperature field provides a good match to an average of three temperature records recovered from ice cores. The B18 snow-accumulation reconstruction is simply a
425 reference climatology scaled to accumulation rates from the GISP2 ice core. We treat this as a precipitation reconstruction, but note that accumulation may be less than precipitation at some locations around Greenland, especially near the coast. It is also informative to compare these results with our S4 temperature and high precipitation reconstructions, hereafter referred to as S4 and high P, as well as with the TraCE21ka simulation itself (i.e., the climate model output, unconstrained by data). For brevity, we focus on the area around Kangerlussuaq in southwest Greenland, but the comparisons are generally applicable
430 to any region of Greenland. Southwest Greenland is of interest because the ice-sheet behavior here is primarily a response to changes in surface forcing (i.e., temperature and precipitation) because there are few tidewater glaciers (Cuzzone et al., 2019). Furthermore, the Kangerlussuaq region has a particularly well-documented ice-sheet retreat history through the Holocene (Young and Briner, 2015; Lesnek and Briner, 2018).

In the Kangerlussuaq region, the B18 reconstruction shows more extreme temperature changes than our reconstructions,
435 with late-glacial (20-15 ka) anomalies of about -20°C and peak HTM temperature anomalies of about $+2^{\circ}\text{C}$ at 9 ka (Fig. 12). B18 also shows a faster rate of transition between the glacial and Holocene, reaching temperatures close to modern by 10 ka. In contrast, TraCE21ka shows more moderate temperature anomalies and a slower transition, with late-glacial anomalies of about -8.6°C and near-modern temperatures that first appear around 7 ka. TraCE21ka has no obvious HTM in this location. Our main reanalysis and the S4 version of our temperature reanalysis both lie between B18 and TraCE21ka, with
440 late-glacial anomalies of about -12 and -14°C , respectively, Holocene peak temperature anomalies of $+1^{\circ}\text{C}$ around 5 ka, and temperatures close to modern first appearing around 8 ka.

For precipitation, the B18 reconstruction again tends to show the largest fluctuations and fastest transition, with a late-glacial precipitation fraction of about 0.26 and precipitation rates close to modern first appearing just after 10 ka. TraCE21ka again shows the most moderate fluctuations and a slower transition, with a late-glacial fraction of about 0.38 and rates close to modern
445 not appearing until around 5 ka. Our main reanalysis and high P lie in the middle during the late-glacial, with fractions of about 0.32 and 0.36, respectively; however, our main reanalysis has a slow transition into the Holocene, similar to TraCE21ka, while high P has a fast transition similar to B18. In the Holocene, high P shows the most elevated precipitation out of all the reconstructions, with 10-15% more precipitation than modern occurring around 7-3 ka. B18 shows precipitation values similar to modern for the last 10,000 years of the Holocene, while TraCE21ka and our main reanalysis show lower-than-modern
450 precipitation throughout most of the Holocene.



All of these paleoclimate reconstructions – our main reanalysis, the sensitivity scenarios, and B18 – are plausible histories of temperature and precipitation over Greenland. Given any past change in the ice-sheet, each of these histories has a different implication for ice-sheet sensitivity to climate, the veracity of which could be tested by using them to force an ice-sheet model and comparing this ensemble of results to the geologic record.

455 An important distinction among various different paleoclimate reconstructions for Greenland is in the treatment of elevation changes. Any paleoclimate reconstruction from ice-core records is complicated by ice-sheet elevation changes. In Vinther et al. (2009), it is assumed that the climate history is the same at all locations around Greenland, and that any differences among the ice core paleotemperature records is a result of that elevation change. In B18, past elevation changes are assumed to be negligible. In our reconstruction, the impact of elevation change on the spatial covariances of temperature and precipitation
460 is implicitly accounted for as part of the data assimilation methodology. Formally, our reconstruction is of surface climate, not climate at a fixed elevation, but it depends on the accuracy of the climate-elevation relationships in our prior – i.e. in the TraCE21ka climate model simulation, which probably does not capture such relationships with particularly high fidelity since the model resolution is low and the climate and ice-sheet models are not coupled. Future work could take advantage of the probabilistic relationships among accumulation, temperature, and surface elevation as simulated in fine-scale regional climate
465 models (Edwards et al., 2014).

6 Conclusions

Paleoclimate data assimilation is a novel method for reconstructing climate fields over the Greenland Ice Sheet. Our approach, combining ice-core records with a climate-model simulation, provides complete spatial reanalyses of both temperature and precipitation covering the last 20,000 years. Evaluation against independent proxy records shows that this methodology leads
470 to significant and meaningful improvement over the prior ensemble (drawn from a climate simulation). The results also quantify uncertainty in all aspects, which provides a range of climate scenarios for ice-sheet modeling. Moreover, independently reconstructing both precipitation and temperature allows the assumption of purely thermodynamic control on precipitation to be relaxed, and an examination of the relationship between these quantities over a range of timescales. Specifically, we find that the Clausius-Claypeyron scaling is a good approximation over glacial-interglacial cycles, but not for shorter timescales
475 where precipitation variability partially decouples from temperature.

Our results have potentially important implications for the response of the Greenland ice sheet to climate change. In particular, we find maximum Holocene temperatures were reached around 5 ka, which is between 500 years earlier and 4,000 years later than previous estimates. Moreover, there is little corresponding change in precipitation in our main reanalysis and one of the sensitivity scenarios. If these findings are correct, they imply a relatively rapid response to temperature forcing for sections
480 of the ice sheet margin that retreated less than a century later (Young and Briner, 2015). A caveat is that proxy data remains very sparse, particularly in southern Greenland, where the poorly-resolved Dye3 core is the only long record. Future work to obtain improved measurements on the Dye3 core, or gather other new data from southern Greenland, would help to alleviate this limitation, as would the incorporation of data from off the ice sheet, such as from lake and ocean sediment cores.



485 Finally, we note that our paleoclimate data assimilation approach would benefit from a larger selection of climate-model
simulations at higher resolution. Particularly valuable would be transient simulations that include water isotopes as prog-
nostic variables, which allows for direct assimilation of water isotope ratios (Steiger et al., 2017), rather than the use of an
explicit proxy system model between temperature and $\delta^{18}\text{O}$. Recent work shows significant improvements to the realism of
water-isotope enabled models in the polar regions (Nusbaumer et al., 2017; Dütsch et al., 2019), and longer simulations, once
available, should allow us to further improve upon the results we have presented here. In principle, our method could also be
490 applied to climate-model simulations that include a fully-coupled Greenland Ice Sheet. At present, fully-coupled simulations
of Greenland over thousands of years are prohibitively expensive except at low resolution, and the limited work that has been
done shows significant biases (Vizcaino et al., 2015). Nevertheless, incorporating data assimilation into such models would
provide the groundwork for more-complete data-constrained simulations as computing power becomes less of a limiting factor
in the future.

495 *Code and data availability.* The paleoclimate reconstructions in this paper made use of code from the Last Millennium Reanalysis project,
which is publicly available at <https://github.com/modons/LMR> (Hakim, 2019). The reconstructions, along with the new accumulation histo-
ries for Dye3, GRIP, and NGRIP, will be made publicly available at www.pangaea.de at the time of publication.

Author contributions. JAB and EJS conceived the idea for the study. JAB wrote code improvements necessary for this work, completed the
paleoclimate reconstructions with guidance from EJS and GJH, conducted all analyses of the results, and wrote the first draft of the paper.
500 GJH provided expert advice on data assimilation methodology and code development. TJF made the calculations of ice flow used to model
accumulation for the Dye3, GRIP, and NGRIP cores. All authors contributed to the final version of the manuscript.

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Figures

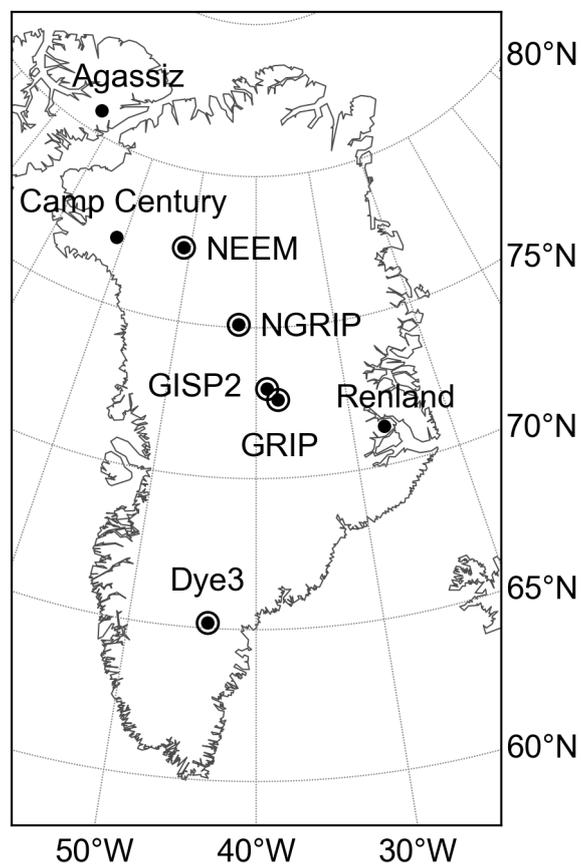


Figure 1. Locations of the ice-core sites referenced in this study. We use oxygen isotope ($\delta^{18}\text{O}$) records from all eight sites and accumulation records from the five circled sites.



Table 1. Metadata for the water isotope ($\delta^{18}\text{O}$) and accumulation (accum) records referenced in this study. "NBI" refers to the Niels Bohr Institute data access site (<http://www.iceandclimate.nbi.ku.dk/data/>) and "Pangaea" refers to the Pangaea data access site (<https://www.pangaea.de/>). Latitude and longitude are in units of decimal degrees (dd) and dates are in thousands of years before 1950 CE (ka).

Ice core name	Latitude (dd)	Longitude (dd)	Variables	Oldest (ka)	Youngest (ka)	Source	Citations
Agassiz	80.7	286.9	$\delta^{18}\text{O}$	11.64	-0.02	NBI	1
Camp Century	77.18	298.88	$\delta^{18}\text{O}$	11.64	-0.02	NBI	1
NEEM	77.45	308.94	$\delta^{18}\text{O}$	>20	-0.0108	NBI	2, 3, 4
			accum	>20	-0.04	NBI	5
NGRIP	75.1	317.7	$\delta^{18}\text{O}$	>20	-0.04	NBI	6
			accum	>20	-0.02	this study	7, 8, 9, 10
GISP2	72.97	321.2	$\delta^{18}\text{O}$	>20	-0.04	NBI	11, 12
			accum	>20	-0.0375	Pangaea	13
GRIP	72.6	322.4	$\delta^{18}\text{O}$	>20	-0.02	NBI	14
			accum	>20	-0.02	this study	7, 8, 9, 10
Renland	71.27	333.27	$\delta^{18}\text{O}$	11.64	-0.02	NBI	1
Dye3	65.18	316.18	$\delta^{18}\text{O}$	>20	-0.02	NBI	1, 15
			accum	11.640	0	this study	16, 17

¹ Vinther et al. (2009), ² Dahl-Jensen et al. (2013), ³ Schüpbach et al. (2018), ⁴ personal comm. Bo Vinther, ⁵ Rasmussen et al. (2013), ⁶ Andersen et al. (2004), ⁷ Vinther et al. (2006), ⁸ Rasmussen et al. (2006), ⁹ Andersen et al. (2006), ¹⁰ Svensson et al. (2006), ¹¹ Grootes and Stuiver (1997), ¹² Stuiver and Grootes (2000), ¹³ Cuffey and Clow (1997), ¹⁴ Johnsen et al. (1997), ¹⁵ Dansgaard et al. (1982), ¹⁶ Vinther et al. (2009), ¹⁷ this study

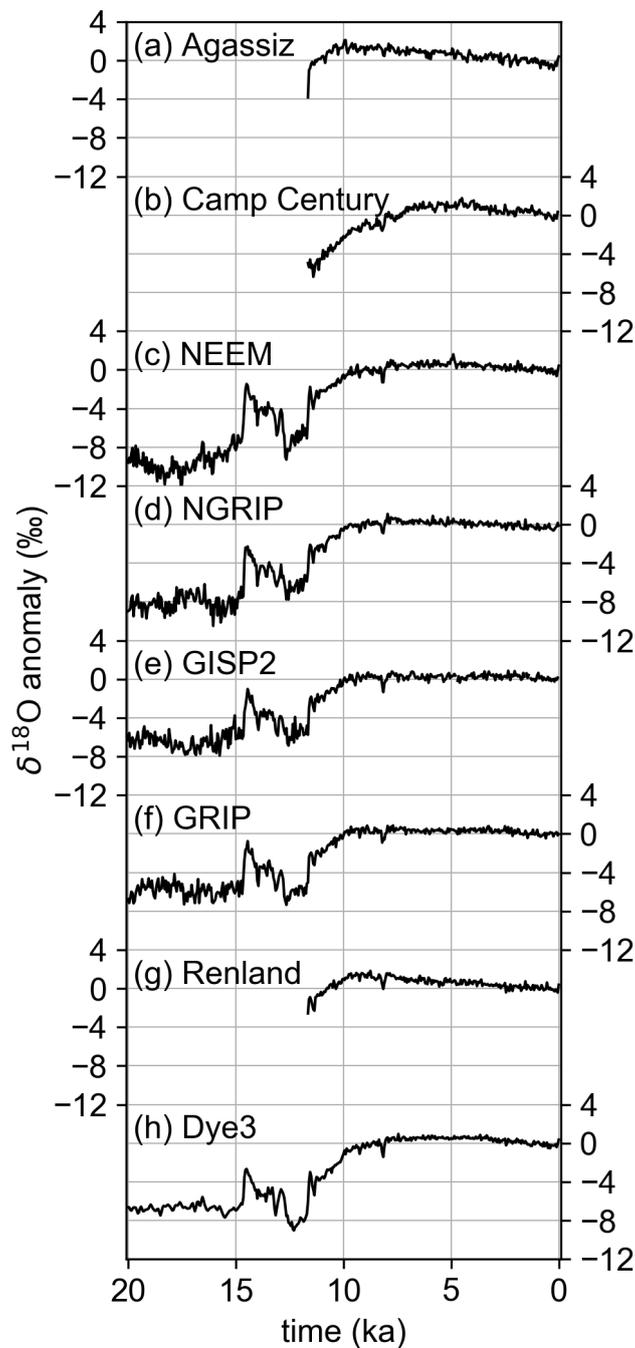


Figure 2. $\delta^{18}\text{O}$ records assimilated into the temperature reconstruction. Records are shown as anomalies relative to the mean of 1850–2000 CE and are ordered top to bottom from northernmost to southernmost. Ice-core site names are given above each record.

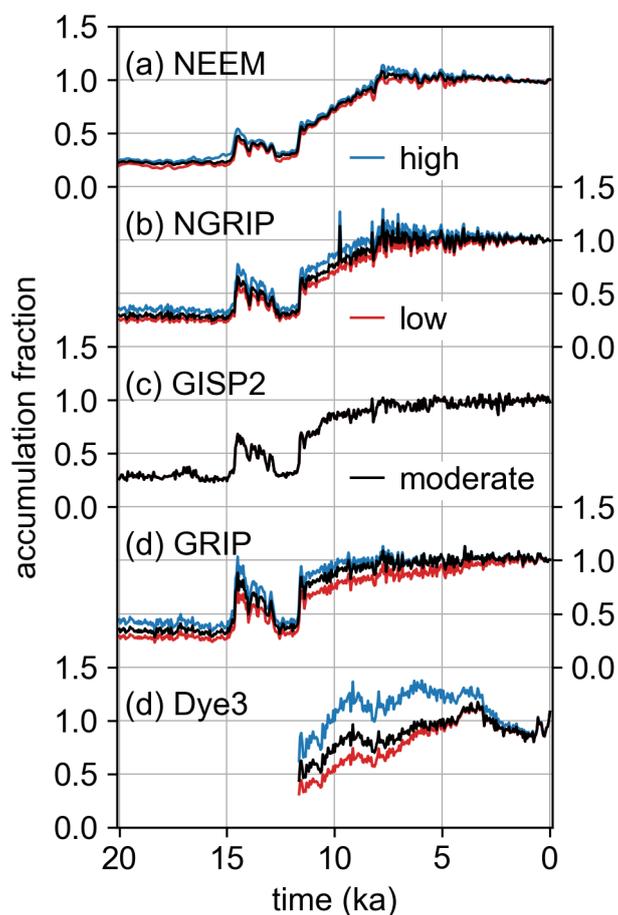


Figure 3. Accumulation records assimilated to reconstruct precipitation for the main reanalysis and two sensitivity scenarios. Records are shown as fractions relative to the mean of 1850-2000 CE and are ordered top to bottom from northernmost to southernmost. Black lines are the moderate records which are included in the main precipitation reanalysis, red lines are the low records which are included in the low sensitivity scenario, and blue lines are the high records which are included in the high sensitivity scenario. Note that we use the same GISP2 accumulation record for the main, high, and low scenarios. Ice-core site names are given above each set of records.

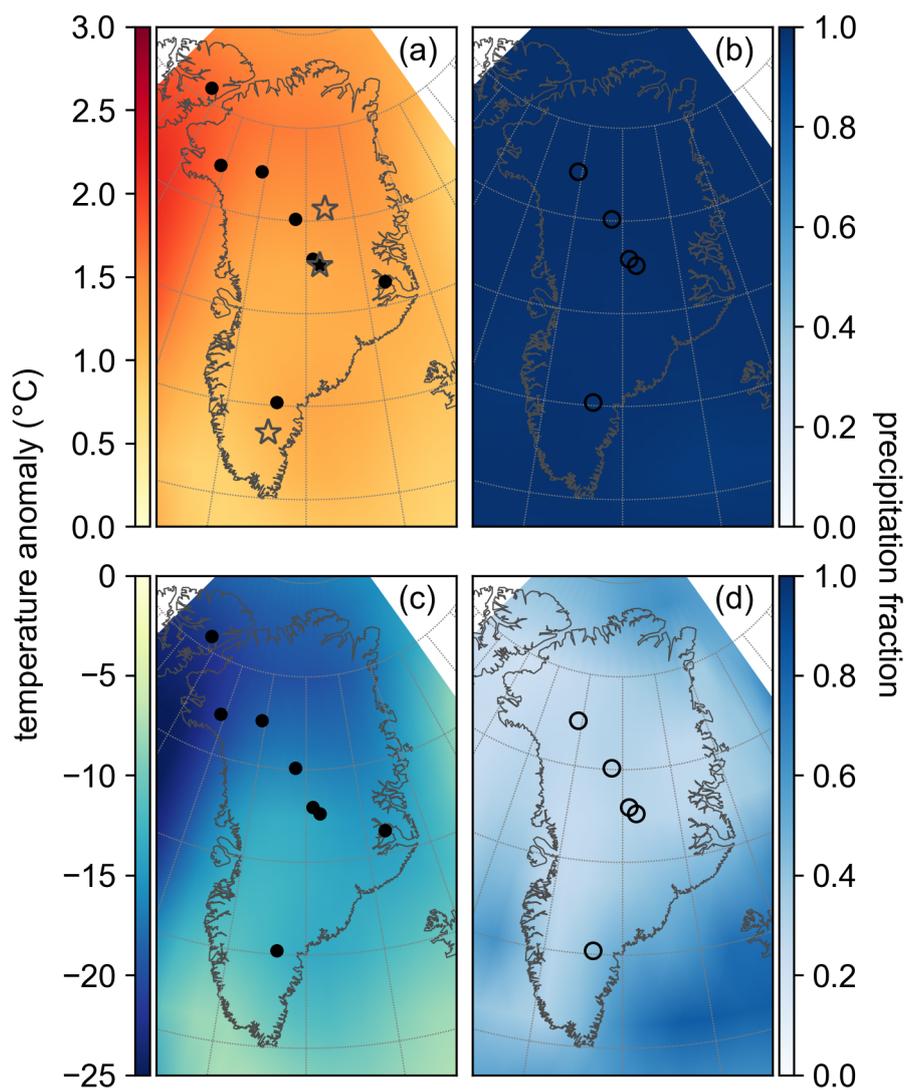


Figure 4. Spatial pattern of the reanalysis mean for temperature (panels (a), (c)) and precipitation (panels (b), (d)). (a) and (b) are averaged over 1,000 years around the peak warmth in the Holocene, 5.5-4.5 ka, while (c) and (d) are averaged over 5,000 years in the late glacial, 20-15 ka. Anomalies and fractions are with respect to the mean of 1850-2000 CE. Points show ice-core locations used for each reanalysis with closed circles indicating $\delta^{18}\text{O}$ records and open circles indicating accumulation records. Grey stars show the locations of the EGRIP ice-core site, Summit, and South Dome, which are referenced in Figs. 5 and 11.

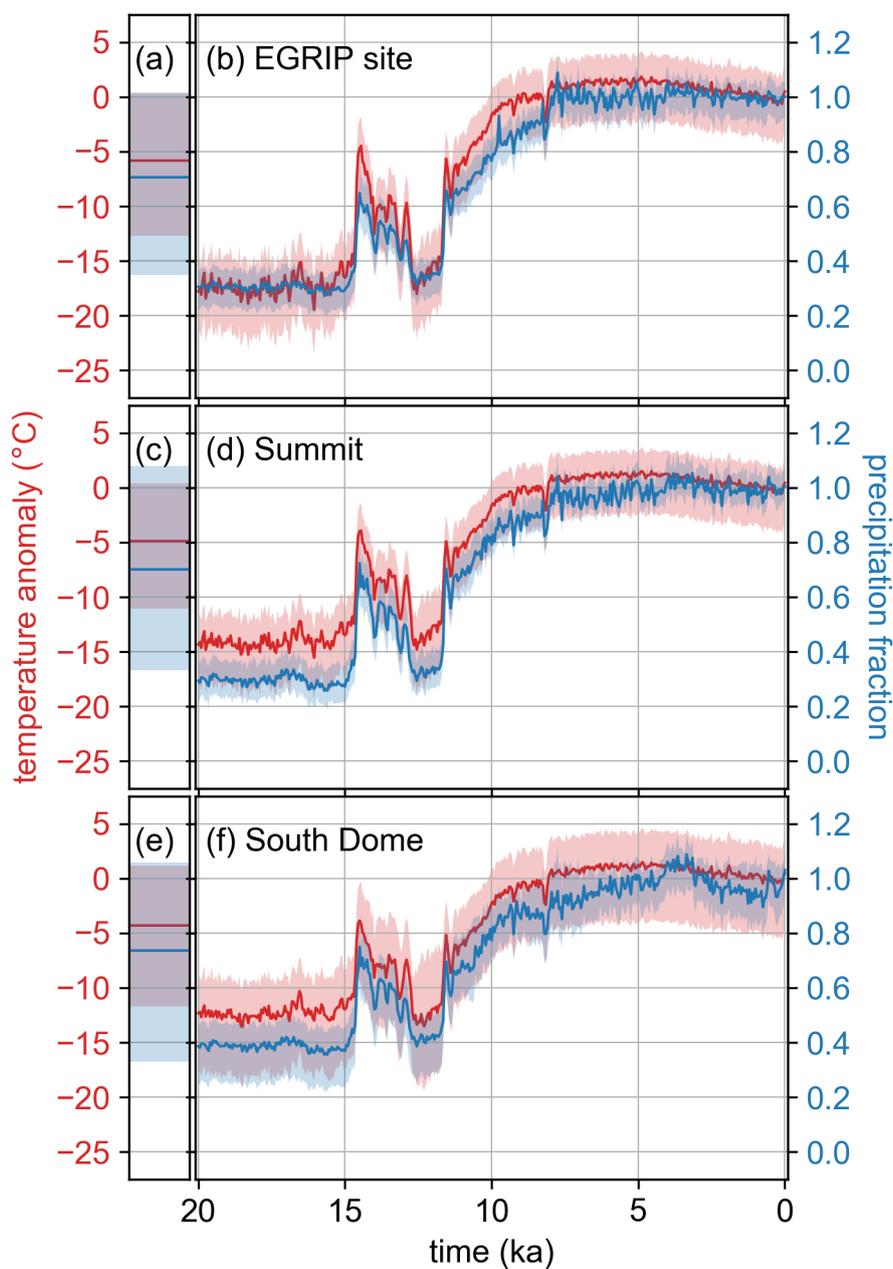


Figure 5. Time series of the prior (panels (a), (c) and (e)) and reanalysis (panels (b), (d) and (f)) ensemble mean and 5th to 95th percentile shading for temperature (red) and precipitation (blue) at three locations. Anomalies and fractions are with respect to the mean of 1850-2000 CE. (a) and (b) show these time series for the location closest to the EGRIP ice-core site, (c) and (d) show for the location closest to Summit, and (e) and (f) show for the location closest to South Dome. These locations are ordered from northernmost (top) to southernmost (bottom) and are shown on a map in Fig. 4.

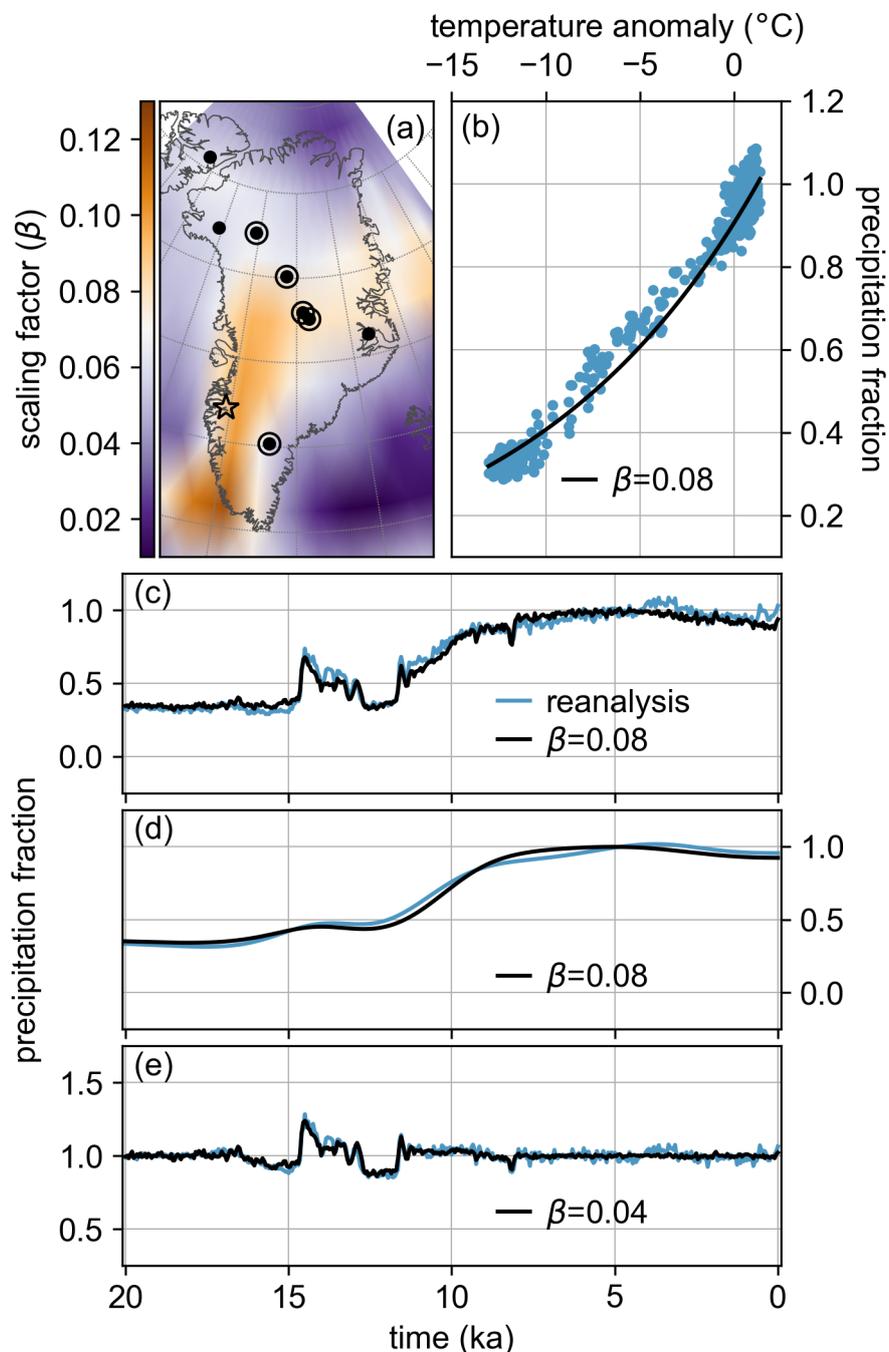


Figure 6. The precipitation-temperature relationship in our reanalysis. (a) shows the spatial pattern of the scaling factor (β) for the best-fit thermodynamic scaling. The colorbar is centered on 0.07, the value used by Greve et al. (2011). Points indicate ice-core locations used for each reanalysis with closed circles indicating $\delta^{18}\text{O}$ records and open circles indicating accumulation records. The star is at the center of the area used in panels (b)-(e) (65°N to 68.7°N and 48.5°W to 52.5°W). (b) is a scatter plot of temperature anomaly vs. precipitation fraction from the reanalysis (blue points). The black line shows the best-fit exponential scaling. (c) shows the time series of the precipitation reanalysis (blue line) and precipitation scaled from temperature using the best-fit scaling (black line). (d) and (e) are the same as (c) except low-pass and high-pass filtered, respectively, with a cutoff frequency of $5,000 \text{ year}^{-1}$.

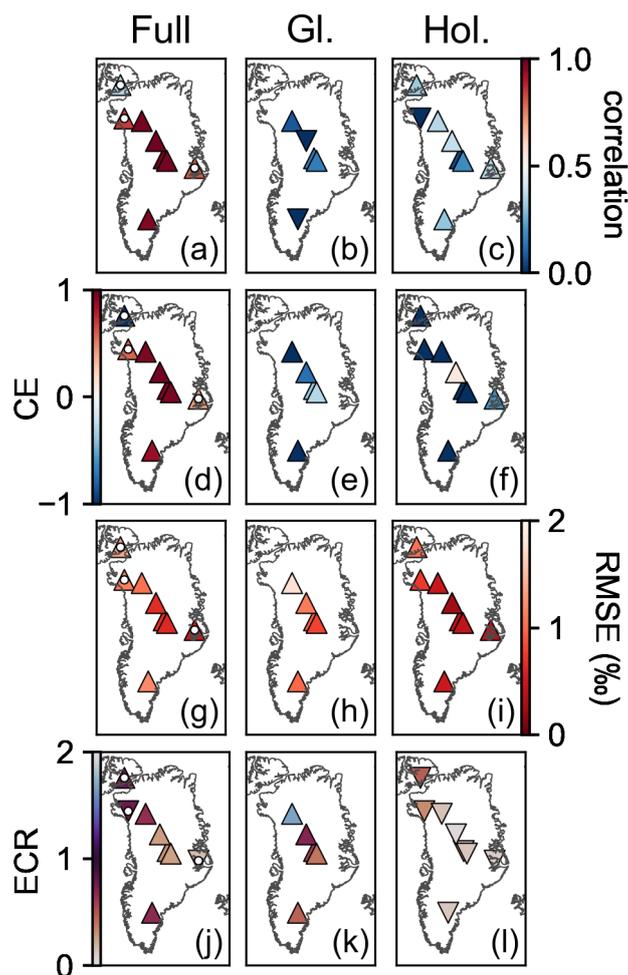


Figure 7. Skill metrics averaged over iterations and time for the temperature reanalysis. The first column (panels (a), (d), (g), and (j)) shows the skill metrics for the full overlap (Full) between the proxy record and reanalysis. A white dot indicates evaluation against proxy records that overlap only the Holocene (11.7-0 ka). The middle column (panels (b), (e), (h), and (k)) shows the skill metrics for a period in the glacial (Gl.) (20-15 ka), while the right column (panels (c), (f), (i), and (l)) is for a period in the Holocene (Hol.) (8-3 ka). The first row (panels (a)-(c)) reports the correlation coefficient, the second row (panels (d)-(f)) the coefficient of efficiency (CE), the third (panels (g)-(i)) the root mean square error (RMSE), and the fourth row (panels (j)-(l)) the ensemble calibration ratio (ECR). Triangle symbols pointing up indicate that the posterior ensemble evaluates better than the prior ensemble for that location and statistic. Triangle symbols pointing down indicate the opposite. We define better evaluation as correlation coefficient closer to 1, CE closer to 1, RMSE closer to 0, and ECR closer to 1.

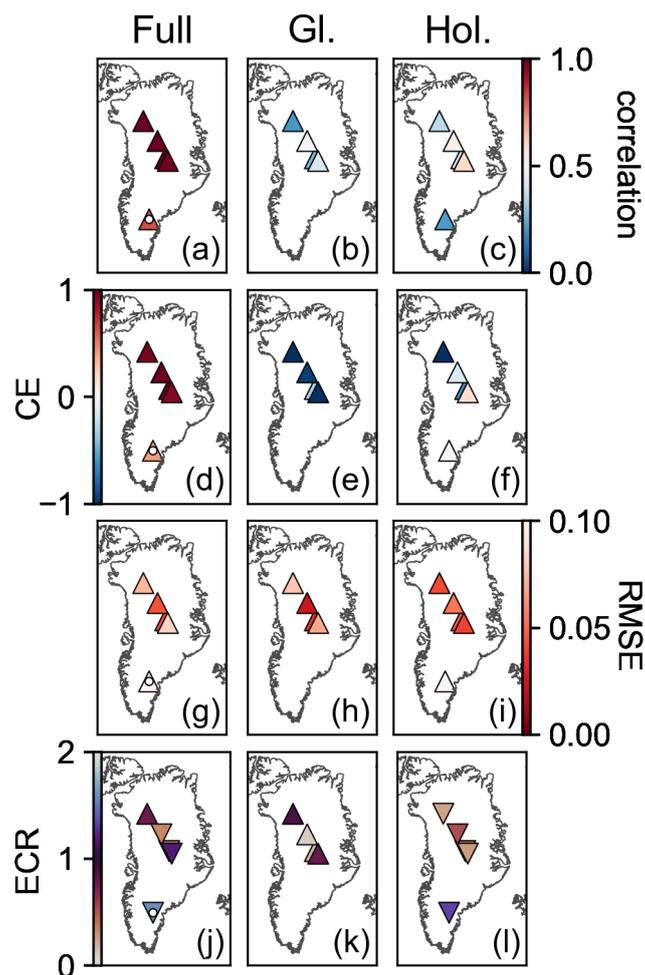


Figure 8. Skill metrics averaged over iterations and time for the precipitation reanalysis. The first column (panels (a), (d), (g), and (j)) shows the skill metrics for the full overlap (Full) between the proxy record and reanalysis. A white dot indicates evaluation against proxy records that overlap only the Holocene (11.7-0 ka). The middle column (panels (b), (e), (h), and (k)) shows the skill metrics for a period in the glacial (Gl.) (20-15 ka), while the right column (panels (c), (f), (i), and (l)) is for a period in the Holocene (Hol.) (8-3 ka). The first row (panels (a)-(c)) reports the correlation coefficient, the second row (panels (d)-(f)) the coefficient of efficiency (CE), the third (panels (g)-(i)) the root mean square error (RMSE), and the fourth row (panels (j)-(l)) the ensemble calibration ratio (ECR). Triangle symbols pointing up indicate that the posterior ensemble evaluates better than the prior ensemble for that location and statistic. Triangle symbols pointing down indicate the opposite. We define better evaluation as correlation coefficient closer to 1, CE closer to 1, RMSE closer to 0, and ECR closer to 1.

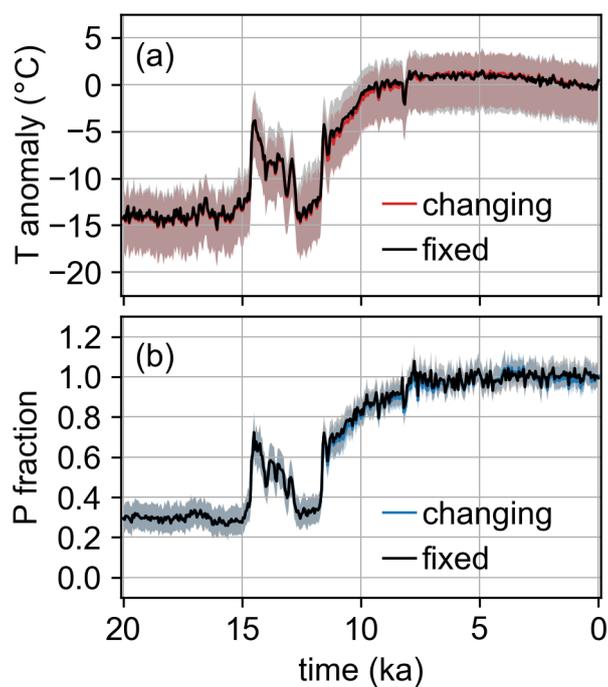


Figure 9. Changing (red and blue) vs. fixed (black) proxy-network for the (a) temperature (T) and (b) precipitation (P) reanalysis mean and 5th to 95th percentile shading. Anomalies and fractions are with respect to the mean of 1850-2000 CE. These time series are for the location closest to Summit, which is representative of the results around Greenland.

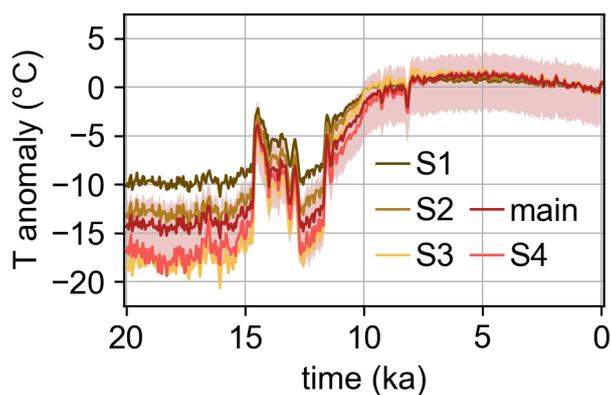


Figure 10. The main temperature (T) reanalysis (ensemble mean and 5th to 95th percentile shading) and ensemble mean for four sensitivity scenarios, S1-S4. Each sensitivity scenario reflects a different assumption about precipitation seasonality, with S1-S3 assuming a spatially-uniform seasonality and S3-S4 assuming stronger seasonality than the main reanalysis. Anomalies are with respect to the mean of 1850-2000 CE. These time series are for the location closest to Summit, which is representative of the results around Greenland.

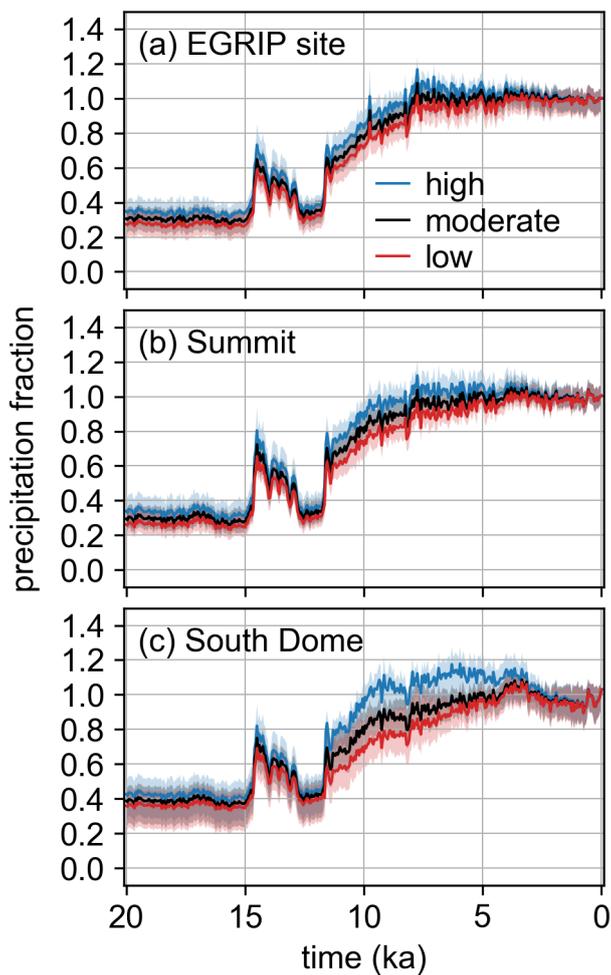


Figure 11. Ensemble mean and 5th to 95th percentile shading for the main precipitation reanalysis (black), high sensitivity scenario (blue), and low sensitivity scenario (red). Fractions are with respect to the mean of 1850-2000 CE. (a) is the time series for the location closest to the EGRIP ice-core site, (b) is closest to Summit, and (c) is closest to South Dome, which are representative of northern, central, and southern Greenland and are shown on a map in Fig. 4.

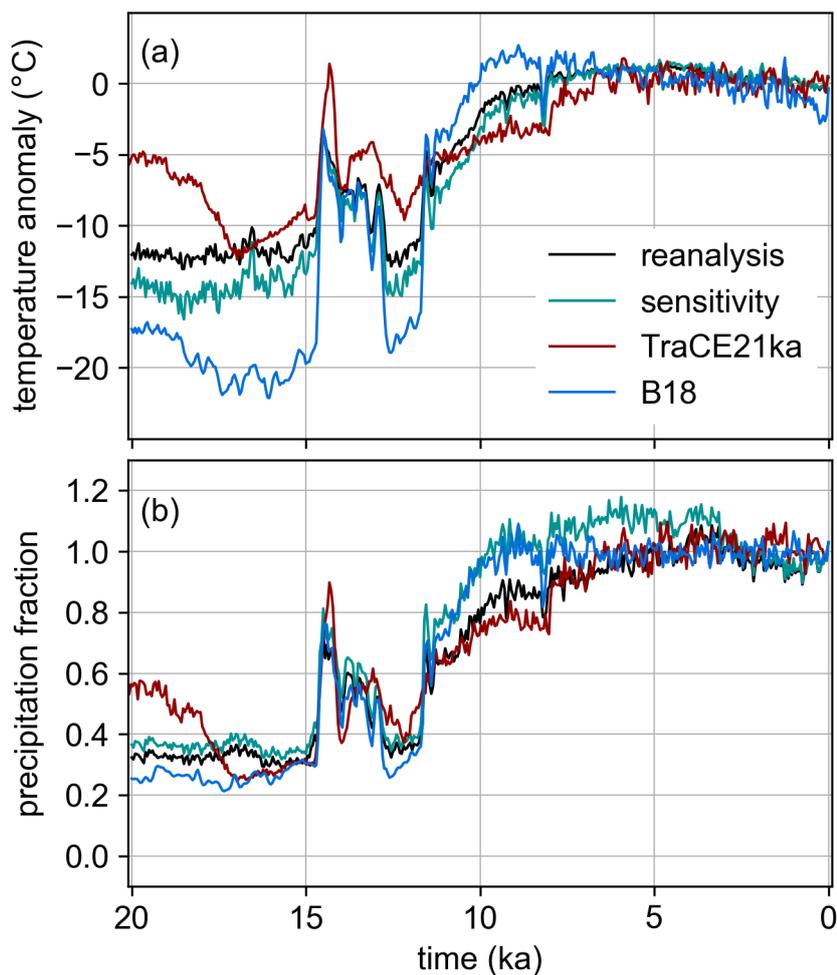


Figure 12. Temperature (a) and precipitation (b) reconstructions from our main reanalysis (black), our sensitivity scenarios S4 and high P (green), TraCE21ka (red), and B18 (blue) (Buizert et al., 2018). Each reconstruction is averaged to a 50-year time resolution and averaged over a spatial domain in the Kangerlussuaq region, defined by the latitude-longitude box 65°N to 68.7°N and 48.5°W to 52.5°W, the center of which is located at the star in Fig. 6a. Temperature anomalies and precipitation fraction are defined with reference to the mean of 1850–2000 CE.