Reply to Referee Comment #3

We thank the referee for their time and insights. In the following, we address the referee's comments, which are in black. Our replies are inline in blue. Note that figures and tables in this reply are labeled with "R" preceding the number.

Review of Badgeley et al. 2020 on Greenland paleo data assimilation

Badgeley et al. present temperature and precipitation fields for the last 20,000 years over Greenland generated using a paleo data-assimilation technique. This is an interesting and potentially very valuable new approach to investigating past climates. The paper is well written and clearly illustrated, and I am generally enthusiastic about the work.

Thank you.

While the methodology represents a big step forward, the paper is also a step backwards in other regards as it assumes a constant linear scaling of d18O to site temperature for all sites and periods based on the spatial d18O-T relationship. This assumption has been disproven in the last 2 decades through careful work in the ice core community (including some of the papers cited here). This assumption will dominate all the spatial and temporal patterns in the temperature reconstructions, and deserves more careful consideration than it is given here. The authors suggest that this problem is alleviated by using the precipitation weighted temperature, but they do not demonstrate this. Below I recommend some comparisons that should be done before the paper is suitable for publication.

My main concern is the use of a single linear d18O-T scaling based on the spatial d18O-T pattern at all sites and locations. While water isotopes are a valuable proxy, its temperature interpretation has proved very difficult. Borehole thermometry and d15N gas thermometry are the most reliable methods to get absolute (calibrated) temperature changes, and both suggest a d18O slope that is around half of the spatial relationship (0.67 permil/K) used here (as the authors acknowledge).

We agree with the referee that it would be overly-simplistic to assume "a constant linear scaling of δ^{18}O to site temperature for all sites and periods based on the spatial δ^{18}O-T relationship". This is not, however, what we do; we allow the δ^{18}O-temperature relationship to vary spatially by relying on precipitation-weighted temperature (T*) from TraCE-21ka. As we write in our paper (lines 229-232), "Numerous studies have suggested that precipitation seasonality is the largest source of nonlinearity in the δ^{18}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 δ^{18}O-T site relationship for the glacial-interglacial transition has such a low slope (Werner et al., 2000)." We convert T* to δ^{18}O using the equation $0.67\mathrm{T}^* = \delta^{18}\mathrm{O}$, and then compute the best-fit slope between δ^{18}O and temperature to find their relationship. This effectively assumes a constant linear scaling between the *instantaneous* δ^{18}O and site temperature, while allowing for changes in precipitation to affect the time-averaged relationship. These TraCE-21ka-derived slopes vary between 0.42 and 0.66 ‰ °C^{-1} at the core sites (Table R1), and are less than the modern spatial relationship of 0.67 ‰ °C^{-1} at most locations around Greenland (e.g., Fig. R1).

The TraCE-21ka-derived slopes vary both in space and across prior ensembles (Fig. R1 and Table R1). By using ten different prior ensembles, we capture the uncertainty in the δ^{18}O-temperature relationship, as determined by TraCE-21ka. We also examine a wider range of slope estimates in our sensitivity experiments. The slopes for S4 are shown in Table R1, and the slopes for S1, S2, and S3 are 0.67, 0.5, and 0.335 ‰ °C^{-1}, respectively. As described in our paper (lines 389 to 405), the magnitude of the slope affects the magnitude of the anomalies (Fig. 10 in the paper). The spatial pattern of the slope also has an effect. For example, in the early Holocene, the spatially-variable slopes result in a reconstruction that warms more slowly. In addition, the reconstructions with spatially-varying slopes show stronger north-south gradients than those with spatially-constant slopes. This north-south gradient shows up especially in the abrupt transitions, with larger changes in the north relative to the south (Table R2). In the paper, we had not discussed the impact of the spatially-varying slopes on the spatial pattern of the reconstructions; we will include this discussion in the revised paper by making the following revisions to Sect. 4.2, lines 400 to 405

New text is in *italics*.

The temperature results are also sensitive to the spatial pattern of the δ^{18}O-T relationship. *We find this by comparing* the results from the S1-S3 scenarios that assume a spatially-uniform *relationship* to results from the main reanalysis and S4 scenario that assume a spatially-variable *relationship*. 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. *In addition, the reconstructions with spatially-varying δ^{18}O-T relationships show stronger north-south gradients during times of abrupt temperature change than those with spatially-constant relationships (e.g., Table R2). These findings indicate* that there is new information added by using a PSM that accounts for spatial variability in precipitation seasonality.

The reviewer is correct that we effectively assume that the δ^{18}O-temperature relationship is constant in time. We could in principle account for temporal changes in the δ^{18}O-temperature relationship by using a time-varying prior; however, this method is complicated by discontinuities and/or extra assimilation parameters. We use the same prior ensemble for all time steps of the reconstruction, which avoids discontinuities in the reconstruction and does not require us to constrain extra assimilation parameters with so few proxy records. A consequence of this method is that the δ^{18}O-temperature relationship, which is derived from the prior ensemble, is constant in time. Unfortunately, we are restricted in our ability to both use a time-varying prior and avoid the complications stated above until more long, fully-coupled climate simulations become available (see our reply to Reviewer #1, lines 22-105, for more details).

As the referee mentioned, borehole thermometry and δ^{15}N gas thermometry are reliable methods to getting at the δ^{18}O-temperature relationship; however, as we say on lines 233-237 in the paper, we do not rely on these methods because borehole thermometry and δ^{15}N gas thermometry are not available at all sites. Instead, we rely on the TraCE-21ka-derived relationships described above, which we compare to relationships found previously using borehole and δ^{15}N gas thermometry (Table R1). It is known that the δ^{18}O-temperature relationship varies temporally depending on the length of time considered and the date (Jouzel et al., 1997). Table R1 shows that indeed, different investigations have estimated a variety of slopes for a variety of time periods. Differences in the estimated slopes are likely to result from the method used in a particular investigation and the time period considered. Some estimates, such as Guillevic et al. (2013) and Buizert et al. (2014) are for abrupt transitions, such as Dansgaard-Oeschger events, while others find mean slopes over longer periods of time, such as Kindler et al. (2014) and our own paper. The slopes that we derive from TraCE-21ka mostly fall within the ranges found by previous studies, even though our slopes are estimated for a time period that is not addressed in the other investigations, the last 20,000 years.

We will include this discussion, along with the associated figures and tables, in the revised paper and supplementary information. One of these revisions will be edits to lines 242 to 244 in Sect. 2.3.1 of the paper. New text is in *italics*.

With T^*_{site} in our PSM, we find that the δ^{18}O-T_{site} slope is spatially variable *(e.g., Fig. R1), ranging from 0.42 and 0.66 ‰ °C^{-1} at the ice-core sites (Table R1), and tending to be less than the modern spatial relationship of 0.67 ‰ °C^{-1} at most locations around Greenland. These slopes vary both in space and across prior ensembles. By using ten different prior ensembles, we capture the uncertainty in the δ^{18}O-temperature relationship resulting from variations in precipitation seasonality. These TraCE-21ka-derived estimates lie within the range of slopes estimated for sites around Greenland for a variety of time periods (Table R1). Differences seen in Table R1 reflect both the different methods used and the time period considered. Some estimates, such as Guillevic et al. (2013) and Buizert et al. (2014) are for abrupt transitions, such as Dansgaard-Oeschger events, while others find mean slopes over longer periods of time, such as Kindler et al. (2014) and this investigation.*

I suspect this assumption will lead to underestimated temperature variability in the posterior. The authors should check this for the abrupt transitions at the three sites (GISP2, NEEM, NGRIP) where d15N-based temperature changes are known (Buizert et al. 2014).

We thank the referee for this suggestion. In Table R3 we compare the magnitudes of three abrupt transitions (warming into the Bølling-Allerød, cooling into the Younger Dryas, and warming into the Holocene) against the δ^{15}N-derived temperature estimates from Buizert et al. (2014) at three locations. We note that these three sites are all in central and northern Greenland and are not necessarily representative of southern Greenland or the coastal ice caps (e.g., Agassiz and Renland). The comparison shows that our approach does not underestimate temperature variability as compared to Buizert et al. (2014); for the NEEM and NGRIP sites, the mean values of each study are within one standard deviation, while for the GISP2 site, the one standard deviation uncertainty bounds overlap for two of the three transitions. The specifics of the comparison are dependent on the location and the time period. For example, our reconstruction shows greater variability than the δ^{15}N-derived reconstruction at NEEM, but less variability at NGRIP and GISP2. In addition, the reconstructions agree best during the Younger Dryas at the NGRIP and GISP2 sites, but are in better agreement during the Holocene and Bølling-Allerød transitions at NEEM.

However, there is also a clear spatial gradient, as first noted by [Guillevic et al., 2013], a paper that should be cited and discussed. Guillevic observes that d18O changes are largest towards the north (i.e. NEEM), and smaller towards the south (i.e. Summit). However, the actual temperature changes have the opposite gradient – smallest in the north and largest in the south. This means that the d18O-T relationship has an enormous spatial gradient, from $\sim$0.6 at NEEM to $\sim$0.3 at Summit. The Guillevic temperature gradient is seen in many (all?) climate model simulations and should thus be considered very robust.

Thank you for the suggestion. We agree that the Guillevic et al. (2013) paper should be cited. We are aware of the discrepancy between the spatial gradient in the δ^{18}O and in the δ^{15}N-derived temperature; however, we do not agree that this discrepancy has been resolved. The Guillevic temperature gradient (larger temperature changes in the south than in the north) is based on four central to northern ice core records (GISP2 and GRIP, however, are at essentially the same site near Summit). Guillevic et al. (2013) find that the temperature changes at NGRIP and the Summit cores are statistically indistinguishable; thus most of their spatial pattern is driven by the difference between NEEM and NGRIP/GISP2/GRIP. We cannot know without more constraints whether this north-south pattern that appears between NEEM and these three other cores holds for other parts of Greenland, such as southern Greenland. As we show in Sect. S3 of our supplementary information, a southern data point is key to reconstructing southern Greenland climate.

The referee notes the reproduction of Guillevic et al.'s north-south pattern by climate models as evidence for the interpretation that this pattern extends to southern Greenland. We acknowledge that this has been found in some climate simulations, for example, Buizert et al. (2014) show that the TraCE-21ka simulation has this pattern during the Bølling-Allerød transition. Our prior ensembles are selected randomly from all time-steps in TraCE-21ka, such that the covariance patterns reflect the dominant patterns of climate change in TraCE-21ka. These dominant patterns primarily show that the highest-magnitude temperature changes are in nortnern Greenland and, as a weaker signal, that there are higher-magnitude changes in southern than in central Greenland (e.g., Fig. R2). Thus, the TraCE-21ka model suggests that the pattern during the Bølling-Allerød transition is not the dominant spatial pattern of Greenland climate for the last 20,000 years. When combined with the covariance pattern of our proxy records, the result is our reconstruction, which shows larger changes to the north. We acknowledge that the spatial patterns of our results may be different with a time-varying prior ensemble. We will add this discussion to our revised paper (see lines 188 to 217 in this document).

These patterns are such that when using a single constant slope (as the authors do), the larger temperature changes would appear to be in the north, as is indeed the case in their reconstructions (Fig 4a, 4c). However, the Guillevic result would actually suggest the opposite pattern in temperature. The authors need to plot the magnitude of abrupt climate warming in their reanalysis (either the 14.7 ka or 11.6 ka transition), and compare it to the d15N-based values. My hunch is that they will find the opposite pattern from the Guillevic result.

As noted above, the reviewer is not correct that we use a single slope for all ice-core sites. We use a spatially-varying slope (for an example, see Fig. R1). We will be sure to clarify this in the revised paper.

Figs. R3, R4, and R5 show the spatial pattern of three abrupt transitions in our main reconstruction: warming into the Bølling-Allerød, cooling into the Younger Dryas, and warming into the Holocene, respectively. These same results are shown in Table R3 for the eight ice core locations and are compared to results from Buizert et al. (2014). Our results show increasing variability to the north accross seven of the eight cores, while the Guillevic et al. (2013) and Buizert et al. (2014) results show increasing variability to the south across just three cores. So yes, the referee is correct that our results show the opposite pattern from the Guillevic et al. (2013) results; however, the reason for this difference is not due to a spatially-constant δ^{18}O-tempreature relationship. Instead, our spatial pattern is a result of how the spatial patterns in the proxy records are spread throughout the domain via the spatial covariance pattern between temperature and precipitation-weighted temperature in the prior ensemble.

Despite the opposite gradient in north-south variability during abrupt transitions, there are some similarities between our reconstruction and the Buizert et al. (2014) reconstructions. Using the same time definitions, we computed the difference between the Younger Dryas and Older Dryas temperatures (Fig. R6). We find a similar north-south pattern, with a greater difference in the north than in central and southern Greenland. We find even better correspondance between the Buizert et al. (2014) results and our S3 sensitivity experiment (Fig. R7), which has a spatially-constant δ^{18}O-temperature relationship. We will add a brief discussion about these comparisons to the revised version of the manuscript.

It has also been documented that the d18O-T slope is strongly variable in time, changing by almost a factor of 2 [Kindler et al., 2014].

Please refer to lines 60 to 67 of this document for an explanation of why we use a method that has a constant δ^{18}O-T relationship in time in the prior ensemble. We note, however, at the δ^{18}O-T relationship is free to vary in time in the posterior ensemble.

It would be unreasonable to ask the authors to redo all the work abandoning a key assumption; rather I think they should do a careful comparison to d15N-based estimates of abrupt climate change to assess how well their method captures both the magnitude and spatial pattern of abrupt temperature changes – and the implications this may have for the LGM and Holocene optimum patterns shown in Fig 4a and 4c. Perhaps they can provide some suggestions for future work on ways to assimilate the d15N-based climate constraints directly.

We agree with the referee that it would be ideal to assimilate more than just δ^{18}O records. We thank the referee for their suggestion of including this in a discussion of future work, and we will do so in the revised paper.

In the preceding replies, we have done a careful comparison between our reconstructions and δ^{15}N-based reconstructions. We will include these comparisons in the revised paper. Our comparisons show that our main reconstruction is within error of the δ^{15}N-derived estimates for the three abrupt temperature transitions that occured in the last 20,000 years (Table R3). At the three locations where we can compare to δ^{15}N-derived estimates – NEEM, NGRIP, and Summit – we find some discrepency in the spatial pattern of these abrupt transitions; however, because there are so few δ^{15}N-derived estimates, we cannot say whether this disparity extends to southern Greenland or the coastal ice caps.

We will add the following revisions to lines 306 to 316 in Sect. 3 of the paper. New text is in *italics*. We are not certain yet where we will put the figures mentioned in this revised text; we may merge them into one figure to include in the paper or put them into the Supplementary Information.

> 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 three periods of abrupt temperature*

change, and 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 δ^{18}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 covers three periods of abrupt temperature change: the Bølling-Allerød waming, cooling into the Younger Dryas, and warming into the Holocene. Figs. R3, R4, and R5, respectively, show the spatial patterns of these abrupt changes for our main reanalysis. Our results consistently show the largest-magnitude temperature changes in northern Greenland for each of these events. This finding is in contrast with results from δ^{15}N-derived temperature reconstructions that show larger changes in central Greenland cores than in north-central Greenland cores (e.g., Guillevic et al., 2013; Buizert et al., 2014). It has been argued that this δ^{15}N-derived pattern may be extended to southern Greenland because some climate simulations replicate this pattern (e.g., Buizert et al., 2014). For example, TraCE-21ka shows the largest temperature changes in southern Greenland during the Bølling-Allerød transition (Liu et al., 2009).

Our prior ensembles are selected randomly from all time steps in TraCE-21ka, such that the covariance patterns reflect the dominant patterns of climate change in TraCE-21ka. These dominant patterns primarily show that the highest-magnitude temperature changes are in nortnern Greenland and, as a weaker signal, that there are higher-magnitude changes in southern than in central Greenland. Thus, the TraCE-21ka model suggests that the pattern during the Bølling-Allerød transition is not the dominant spatial pattern of Greenland climate for the last 20,000 years. When combined with the covariance pattern of our proxy records, the result is our reconstruction, which shows larger changes to the north. We acknowledge that the spatial patterns of our results may be different with a time-varying prior ensemble. Estimates of the abrupt temperature events in our current reanalysis, however, are within error of the estimates from Buizert et al. (2014) for most of the three events at each of the three sites, GISP2, NGRIP, and NEEM (Table R3).

If the reconstructed N-S temperature gradient during abrupt change is indeed opposite to the Guillevic gradient, this should be clearly stated in the abstract.

We will edit the relevant part of our abstract to say the following. Note that new text is in *italics* and that some of the edits are in response to later comments (see lines 384 to 386).

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 *oxygen-isotope ratios of ice and accumulation rates* from long ice-core records and extends *this information* to all locations across Greenland using spatial relationships derived from a transient climate-model simulation. *Standard evaluation metrics for this method show that our results capture climate at locations without ice-core records. A comparison to other paleoclimate proxy records shows that our results generally agree with previous findings. There are some differences, however, especially relating to the spatial pattern of abrupt climate transitions, for which our results show greater temperature changes in the north, while temperature reconstructions from $\delta^{15}N$ of N_2 show the opposite. We additionally investigate the relationship between precipitation and temperature, finding that it* 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.

The authors suggest that using precipitation-weighted temperatures alleviates the problems associated with using a linear d18O-T scaling. To validate this claim, at the very least they should show a comparison of the 21ka histories of TraCE 2m temperature and TraCE precipitation-weighted temperature at a key site (e.g. Summit), to show how different these two really are. Ideally, they would show more clearly how this impacts the reconstructed magnitude of the abrupt climate change events (that are most strongly constrained by the d15N data).

We have provided a figure as the referee suggests (Fig. R8), which shows that there is a difference between the temperature and precipitation-weighted temperature. The figure is only for the grid cell closest to Summit, but this is the case for all locations around Greenland, though the magnitude of the difference varies by location. We will include this figure in the revised paper or Supplementary Information.

Through our temperature sensitivity experiments, we have tested the impact that precipitation-weighted temperature has on our reconstructions (see lines 389-405 in our paper for a discussion of the results of our sensitivity experiments). We thank the reviewer for their suggestion that we discuss this in more detail. Previously in this reply (lines 39 to 58 of this document), we discussed the impact that a spatially-varying δ^{18}O-temperature relationship has on our results, and provided a revised section of the paper. This is relevant because the spatial variations in the δ^{18}O-temperature relationship are a result of using precipitation-weighted temperature.

General comments:

Please describe the data assimilation method in more general terms understandable to the non-initiated, so the reader won't have to track down the Hakim reference. Can we think of the posterior as a cleverly weighted sum of the randomly selected model timesteps put into the prior? Is there some relationship between the posterior and the 21ka climate simulation – for example, is the posterior solution for the LGM very similar to the TraCE simulation of the LGM? Is the posterior LGM solution strongly weighted towards LGM model years randomly selelcted in the prior?

The referee is correct that this data assimilation method can be thought of as "a cleverly weighted sum of the randomly selected model timesteps put into the prior" because the ensemble mean is indeed a weighted sum of the ensemble members *if* there is no covariance localization, as is the case in our paper. The referee is also correct that it can be determined whether the posterior mean for a time step in the glacial is more strongly weighted towards ensemble members that were selected from the glacial period in TraCE-21ka; however, these weights are not a direct output of our method.

Thank you for the suggestion to describe data assimilation in more general terms. We agree that it's a good idea to provide a summary of the method. We will make the following edit to the first paragraph in Sect. 2.3, which describes the data assimilation method. The edit is in *italics*.

> 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). *This method can be summed up as a linear combination of randomly-selected model states that are weighted according to new information provided by the proxy records (if no covariance localization is used).*

The TraCE simulation has quite a coarse grid I imagine? Please specify the exact resolution. I imagine it may even put multiple of the ice core sites in a single grid box. Perhaps the grid box resolution could be drawn onto figure 1? It seems that the spatial fields in Fig 4 are much smoother than the model would be. Did you apply smoothing or some other technique?

Yes, we agree that we should specify the exact resolution of TraCE-21ka. As we replied to Referee #2, we will add this to Sect. 2.2 of the paper (lines 132 to 147). The spatial resolution is T-31, or about 3.75 degrees, and the temporal resolution is monthly, but we average it to 50-year resolution (resulting in about 440 time steps).

We thank the reviewer for calling our attention to the fact that we forgot to say in the figure captions that the spatial fields are smoothed. We agree that it could be helpful to see the original resolution, so we will convert our plots back to the original resolution.

How meaningful is it to use global climate simulations and constrain them only in Greenland? From a global perspective, Greenland is essentially a single location and the global climate field is not at all constrained. How well-behaved is the far-field response in the reanalysis? And does this somehow impact the reconstruction? I think doing this with global proxy databases (such as [Shakun et al., 2012] would be a great next step (beyond the scope of this paper of course).

Our paper is on reconstructing Greenland climate using proxy records from Greenland. We agree that a next step in applying paleoclimate data assimilation to galcial-interglacial timescales is to use a global proxy database to reconstruct global climate variables. We will menton this in a paragraph about future work in the conclusion of the revised paper.

Seasonality is very briefly addressed, but it deserves more attention as it is an important climate parameter. Please specifically address seasonality in both the prior and posteriors. Will the reconstructions made available online have T and/or P seasonality in them, and if so, describe how this seasonality is derived. I imagine the seasonality of the posterior can be derived via the assimilation method?

We agree that seasonality deserves more attention. As we note in our reply to Referee #2, we plan to include a description of TraCE-21ka's seasonality in Sect. 2.2 of the paper (lines 132 to 147). Seasonality is only used to compute T^*; it is in no other way included in the data assimilation process or results. Our results are mean-annual 50-year averages, which we will be sure to state early on in the paper. The referee is correct that climate variables for specific seasons or the seasonal cycle itself can theoretically be reconstructed using this data assimilation method.

The authors find an unusually late timing of the Holocene optimum around 5ka – much later than other ice-core based estimate from both d18O and melt layers. Looking at Fig 2, it appears that Camp Century (and perhaps Dye 3) are the only cores that suggest such timing, and since the temperature reanalysis is fully determined by ice core d18O, it follows that those two cores must be responsible for this timing (do you agree with this assessment?). However, as pointed out by [Vinther et al., 2009], these sites experience strong thinning in the first half of the Holocene, which will shift their apparent climatic optimum towards a later age (as early Holocene climatic warmth is masked by a cooler site temperature at higher elevation). Could the late (5ka) timing of the climatic optimum in your reanalysis be an artifact of the thinning history of the Greenland ice sheet? Please discuss briefly in the text.

We agree with the referee that the timing of the Holocene thermal maximum (HTM) in our reconstructions tends to be on the later end of the range of previous findings (see lines 317 to 325 of the paper). We also agree that this signal appears to result from the Camp Century, Dye3, and perhaps the NEEM δ^{18}O records. The referee mentions how thinning in the early Holocene (Vinther et al., 2009) would affect our results. As we state in lines 127 and 460 of the paper, our reconstructions are for climate at the ice-sheet surface, meaning that they include the lapse-rate effect of changing surface elevation. This is opposed to climate reconstructions at a reference elevation. Thus, our reconstruction of a later HTM is consistent with previous findings of an earlier HTM (given a fixed reference elevation) (e.g., McFarlin et al., 2018) and early Holocene thinning (Vinther et al., 2009). This is something we have thought about in great detail, but decided not to include in the paper; however, as suggested by the referee, we will make the following edits to lines 455 to 465 in Sect. 5 of the paper. New text is in *italics*.

> 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 is implicitly accounted for as part of the data assimilation methodology. Formally, our reconstruction is of surface climate, not climate at a fixed elevation. *Consequently, our reanalysis may not be directly comparable to other paleoclimate reconstructions. For example, the HTM is commonly reconstructed as an early Holocene event in records that are at a fixed or nearly-fixed elevation. In our reanalysis, the HTM occurs later, in the mid-Holocene. This is likely a result of thinning in the early Holocene (Vinther et al., 2009), which is captured in the ice-core records and acts to dampen early-Holocene warming signals in our reanalysis.* Our method depends on the accuracy of the climate-elevation relationships in our prior – i.e. in the TraCE-21ka 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 models (Edwards et al., 2014).

The data assimilation is fully dependent upon the accuracy of the TraCE-21 climate model simulation in capturing Greenland climate. Therefore, the paper needs a short evaluation of how well this model actually simulates Greenland T and P in the modern day. The TraCE T and P fields should be compared to modern-day Greenland reconstructions thereof; I would recommend the works by Box et al. on this topic [Box, 2013; Box and Colgan, 2013; Box et al., 2009; Box et al., 2013], but general reanalysis products such as NCEP or ERA5 are suitable also.

The statement that "data assimilation is fully dependent upon the accuracy of the TraCE-21ka climate model simulation in capturing Greenland climate" is incorrect. Our results are dependent only on the spatial covariance patterns of the temperature anomalies and fractional precipitation in TraCE-21ka (referenced to 1850-2000 CE) and the covariance pattern of the proxy records. For our use of TraCE-21ka, what matters is how well TraCE-21ka captures the spatial pattern and variance of past climate anomalies, which is unknown except by comparison with the ice-core data (which our method does implicitly). The fidelity of the modern-day climatology is not particularly relevant.

All the figures show relative temperature changes and accumulation changes (relative to the reference period, which is not defined as far as I can tell). But when forcing ice sheet models absolute values are needed. Are these absolute values taken from the last time-slice of the TraCE simulations, or is something better used?

Before we use TraCE-21ka in our assimilation, we subtract (for temperature) or divide (for precipitation) by the mean of TraCE-21ka over the period 1850-2000 CE. Before we assimilate each proxy record, we subtract (for δ^{18}O) or divide (for accumulation rate) by the mean of that record over the same period, 1850-2000 CE. Thus, the reconstructions are of temperature anomalies and fractional precipitation, which are referenced to the 1850-2000 CE mean climate as recorded by the ice-core records. We thank the referee for pointing out that this was not entirely clear. We will state this more clearly and earlier on in the revised paper.

Now that our reconstructions are complete, they may be turned into absolute values for applications such as ice-sheet modeling. This can be done by adding a 1850-2000 CE temperature climatology to the temperature fields and multiplying the precipitation fields by a 1850-2000 CE precipitation climatology (e.g., from Box, 2013). We do not do this in our paper.

Minor comments

L8: What are “independent ice core records”? d18O? Again, I think the reconstructions should be compared during the abrupt temperature transitions at NEEM NGRIP and GISP2, which is where d15N-N2 provides a very robust estimate of the magnitude of change. Those are the truly independent ice core records to compare to.

Our use of the term "independent ice-core records" refers to the comparison of our results against $\delta^{18}O$ records that are excluded from that reconstruction iteration. This is explained later in the text around line 164, so we will edit the abstract to clarify what is meant. See lines 225 to 240 of this document for these edits.

In our results and discussion we compare to findings from previous work; however, the referee is correct that we do not directly compare to temperature reconstructions derived from $\delta^{15}N$ of N_2. We have now done this comparison, discussed it previously in this reply, and provided a revision to include this discussion in Sect. 3 of the paper (see lines 188 to 217 of this document for the revisions).

L24: This is somewhat misleading, because you'll always need to do such precip corrections unless you are doing a fully coupled ice-climate simulation. As the ice elevation in the ice sheet simulation evolves, it differs from the reference elevation at which the climate field is defined; this needs to be corrected for via clausius-clapeyron or similar. So also with your forcing the ice sheet models will need to apply thermodynamic precip corrections.

We agree that ice-sheet models need to correct precipitation fields for changing surface elevation using some assumption about the relationship between precipitation and elevation. Our point, however, is about ice-sheet simulations that base their precipitation histories entirely on their temperature histories using a thermodynamic relation. Thus, the time series and spatial pattern the precipitation anomalies perfectly match those from the temperature fields, which we know to be false from ice-core records. We will clarify this distinction in the revised paper.

L28: Many more d15N studies to cite here: [Guillevic et al., 2013; Kindler et al., 2014; Severinghaus and Brook, 1999; Severinghaus et al., 1998]

Thank you. We appreciate the recommended citations and will include them in the revised paper.

L38: "restricted to a single climate model realization"; wouldn't this critique apply to your study as well? It appears that both use the exact same climate model run.

Yes, the referee is correct, and we will remove the quoted language from the revised paper. Our current results rely on a single climate simulation; however, our method is easily generalizable to any climate simulation of the past 20,000 years. The method in Buizert et al. (2018) is also generalizable to other climate simulations as long as these simulations are accompanied by single-fociring experients.

L70: "measured layer thickness" is not really true. For several cores you use volcanic ties, in which case the layer thicknesses are not measured but inferred

We will change the phrase to "the layer thickness".

L136: "captures the. . .." This is in the eye of the beholder. With the exception of the Bolling warming itself the TraCE run matches the abrupt transitions poorly – there is no YD to speak of.

We will reword this sentence as follows: "By design, TraCE-21ka captures the major glacial-to-modern temperature change, as well as some of the short-term, rapid climate changes, such as the Bølling-Allerød transition (Liu et al., 2009)."

L145: why not use P-E? is evaporation negligible?

As we write later in the paper (lines 257 to 262), "accumulation is closely related to total precipitation at our ice-core sites". We demonstrate this with high-resolution model results (Langen et al., 2015, 2017) that show at most of the ice-core sites in Greenland, precipitation and accumulation are nearly equivalent, suggesting that evaporation is negligible. Dye3 is the only site where there is a difference between accumulation and precipitation, and this is due primarily to melt, not evaporation. This is why we extract P from TraCE-21ka (as mentioned on line 145 of the paper), and use it, rather than P-E, for our proxy system model (described in Sect. 2.3.2 of the paper, lines 256-269).

L217: “highly correlated” is a strong statement. Do you have a reference? Normally d18O and site temperature are not highly correlated at most sites on observational time scales (< 0.5).

The referee is correct that the correlations are low on interannual timescales, but it is well established to be high on longer timescales (Jouzel et al., 1997).

L224: Based on the recent literature, I think that post-depositional alternation may be the largest complication in interpreting the d18O record. Please mention.

We disagree. As we note in the paper, diffusion in the firn column is irrelevant for our timescales of 50 years (Cuffey and Steig, 1998). The reviewer may be thinking of post-depositional processes involving water exchange between the snowpack and the atmosphere (e.g., Steen-Larsen et al., 2011), but it is not established that this has any significant effect on long-term relationships. Indeed, if anything, such processes improve the relationship between temperature and δ^{18}O as they tend to reduce the bias caused by the fact that snow does not accumulated continuously, but as discrete events.

L241: Can you plot T_{site} and T^*_{site} together for the last 21ka at a key site (e.g. Summit). That will let the reader judge the impact of using T^* instead of T.

Thank you for the suggestion. We have done this in Fig. R8, which we will include in a section of the supplementary information.

How is the seasonality of the posterior linked to the seasonality of the prior?

As we stated previously, we only use seasonality to compute T^*; it is in no other way included in the data assimilation process or results. Both our prior ensemble and our reconstructions are made up of mean-annual 50-year averages. We will clarify this in the revised paper.

L261: “grid-cell closest to site” is this also done for T, or do you use 2D linear interpolation or similar? Are there cases where multiple cores share a closest grid cell?

Yes, this is also done for selecting which T^* value to use in the δ^{18}O PSM. We will clarify that in the revised paper. At the resolution of TraCE-21ka, only the GISP2 and GRIP ice cores have the same closest grid cell.

L295: maybe a sentence on how this was estimated?

Thank you for bringing our attention to this point of confusion. We will clarify this in the revised paper. We use the same method as was explained for δ^{18}O.

L313-314: But [Dahl-Jensen et al., 1998] estimates it a lot colder at GRIP, more like -22K cooling at the LGM (25ka). This should be mentioned.

Dahl-Jensen et al. (1998) found a colder temperature at GRIP for the LGM, but the time period we discuss is 20-15ka, which is five to ten thousand years later, when the Dahl-Jensen et al. (1998) estimate shows that it has significantly warmed.

L340: Maybe reference [Buchardt et al., 2012] who did very detailed analyses of this.

Thank you. We are aware of this study, and will refer to it in our discussion of the temperature-precipitation relationship on shorter timescales.

L416: are other d18O records really independent? They suffer the same biases from seasonality, source effects, etc. For true independence, compare to d15N-N2.

We agree with the referee that it is important to compare our results to other types of proxy records. In this reply we have compared our temperature reconstructions to those from Buizert et al. (2014) and provided revisions that we will make to the paper.

L438: TraCE has no HTM anywhere! (one of its many problems. . .)

We will change the wording to say, "TraCE-21ka has no obvious HTM in this location or any location around Greenland."

L476: This is more of a discussion than a conclusion item. Consider moving it. Also, see my comment above, the 5ka timing could be an artifact of ice sheet elevation changes.

We will move this paragraph to the discussion, and we will include a discussion of elevation effects (see lines 323 to 348 of this document).

Figure 4: please add panels (e) and (f) with the T and P change over an abrupt transition (e.g. the Bolling onset). In panel (c), only show the cores that actually constrain the LGM (so not Agassiz, camp century and Renland). Why are the field so much smoother than the TraCE CCSM3 model resolution? Baffin bay has a large temp response with no cores to constrain it – can we trust this?

Thank you for these suggestions. We will add spatial plots of at least one of the abrupt transitions. We will also be sure to only include the ice core locations that contribute to the reconstruction of each time period. As we agreed previously in this reply, we will restore the plots to their original resolution because we agree with the reviewer that this is helpful. We had originally smoothed the plots due to rendering issues, but we have fixed the previous problem.

The goal in using a method like data assimilation is to spread the information from point proxy locations to locations without proxy records. This allows us to reconstruct spatially-complete climate fields. With few proxies and many locations, the problem is underconstrained; however, we have shown that the method is skillful for some locations where proxies are not assimilated (see Sect. 4.1, lines 364 to 382 in the paper).

Fig 5: the “noise” in T (i.e. high frequency signals) at all core sites seem strongly correlated. How come? Could it be that the posterior is more or less reflecting the mean d18O of the various sites?

Each core has some influence on the reconstruction at every location. Thus, the time series at each location is a weighted sum of the time series of each core. This would indeed make the higher-frequency noise correlated at different locations. We will say this in the revised paper.

Fig 6: The largest features in the plot are not directly constrained by any cores. Do you trust these?

For our reply, please see lines 513 to 516 in this document.

Figs 7 and 8 are very technical and could be moved to the supplement.

These figures demonstrate how well our reconstruction performs at locations without any assimilated information. We think this is an important aspect of our paper, and we wish to keep these figures in the main text.

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575 ## Figures

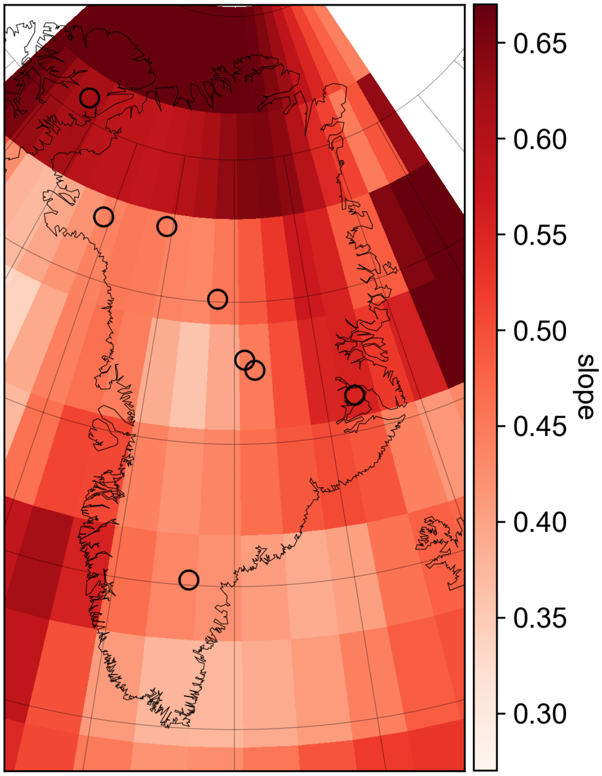

Figure R1. Slope (°C ‰$^{-1}$) of the linear δ^{18}O-temperature relationship for each grid cell. The δ^{18}O-temperature relationship between grid cells is not shown. This is an example from one of the prior ensembles. For reference, open circles show the locations of ice-core records used in this study.

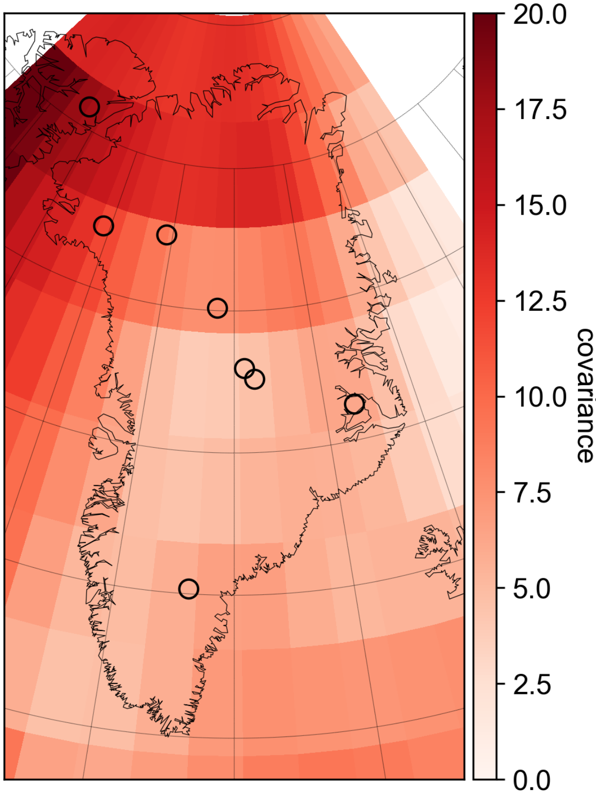

Figure R2. Covariance (°C ‰) of δ^{18}O and temperature for each grid cell. The δ^{18}O-temperature relationship between grid cells is not shown. This is an example from one of the prior ensembles. For reference, open circles show the locations of ice-core records used in this study.

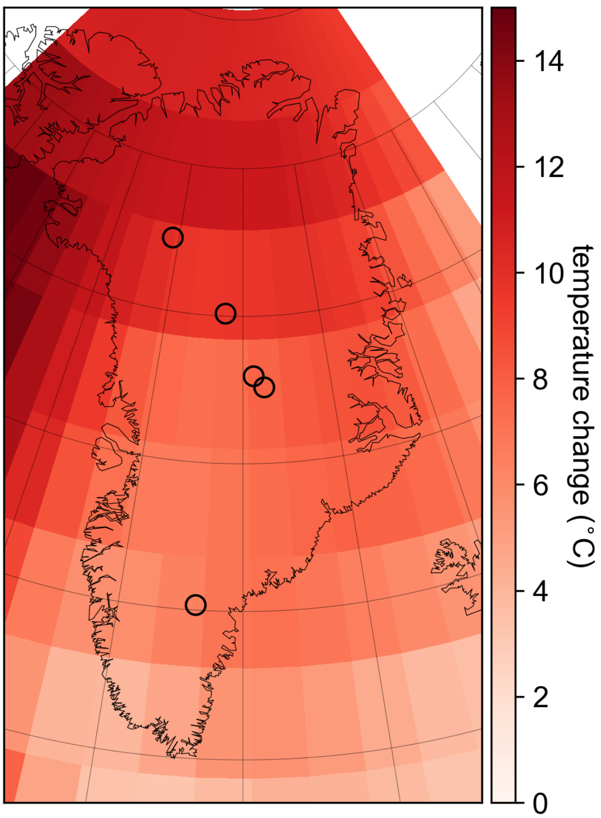

Figure R3. The magnitude of the Bølling-Allerød warming in our main reconstruction. The time definition of this event is the mean of 14.55 to 14.35 ka minus the mean of 14.9 to 14.7 ka, which is the same as in Buizert et al. (2014). For reference, open circles show the locations of ice-core records used in this reconstruction.

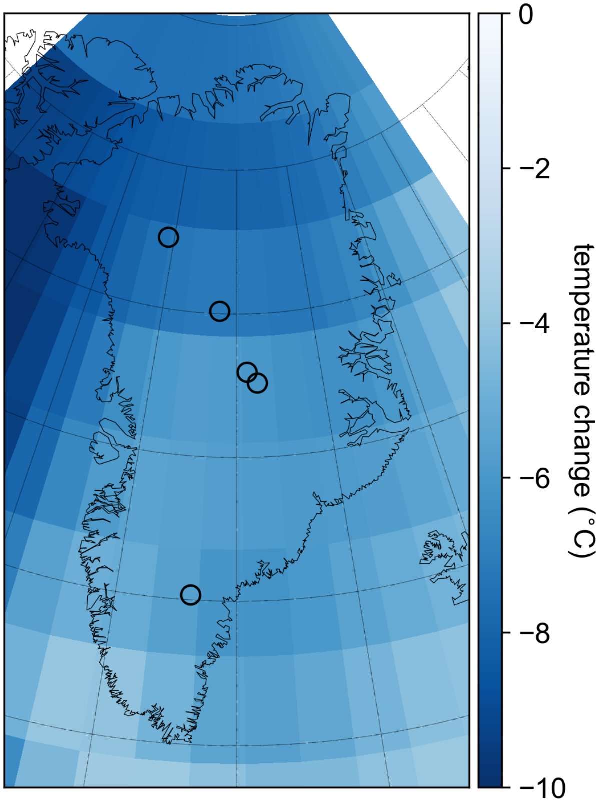

Figure R4. The magnitude of the Younger Dryas cooling in our main reconstruction. The time definition of this event is the mean of 12.5 to 12.3 ka minus the mean of 13.5 to 13.3 ka, which is the same as in Buizert et al. (2014). For reference, open circles show the locations of ice-core records used in this reconstruction.

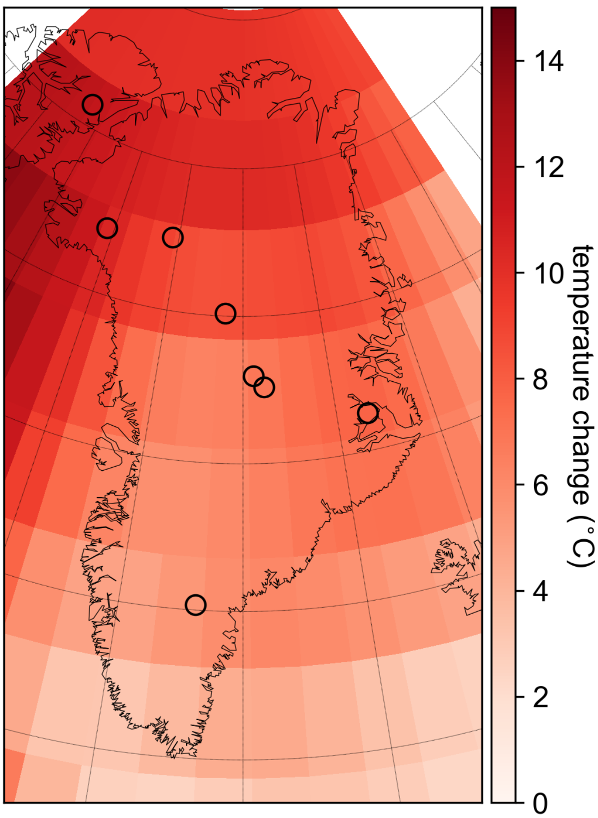

Figure R5. Magnitude of the Holocene warming in our main reconstruction. The time definition of this event is the mean of 11.2 to 11.0 ka minus the mean of 11.9 to 11.7 ka, which is the same as in Buizert et al. (2014). For reference, open circles show the locations of ice-core records used in this reconstruction.

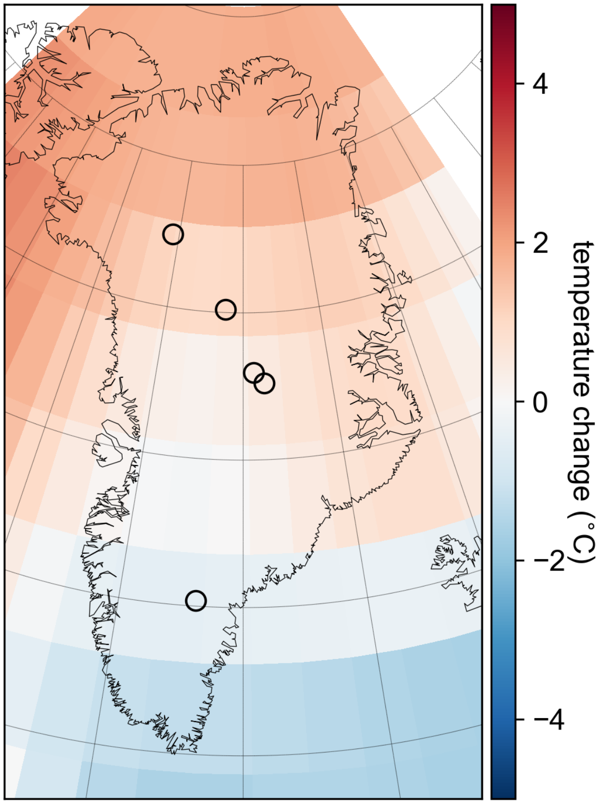

Figure R6. Difference between the Younger Dryas and the Older Dryas in our main reconstruction. The time definition of this event is the mean of 12.5 to 12.0 ka minus the mean of 16.5 to 15.5 ka, which is the same as in Buizert et al. (2014). For reference, open circles show the locations of ice-core records used in this reconstruction.

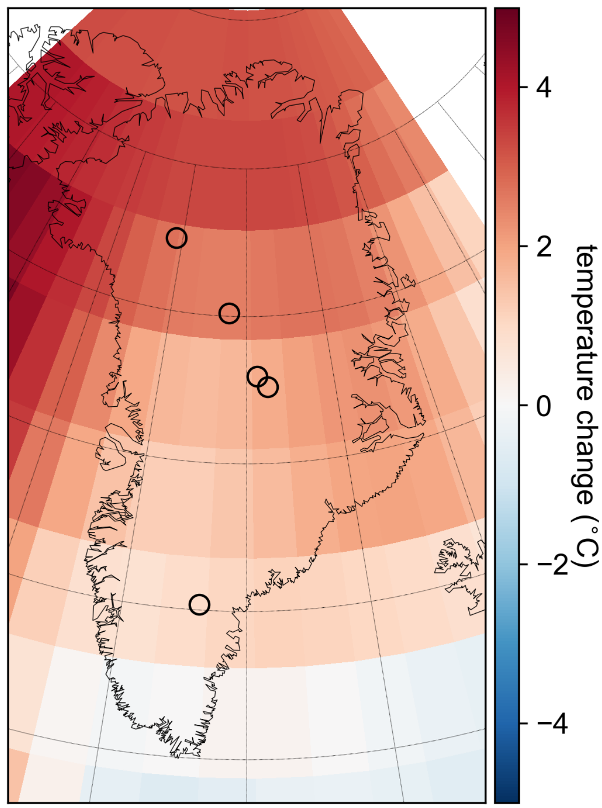

Figure R7. Difference between the Younger Dryas and the Older Dryas in our S3 sensitivity reconstruction, which uses the definition δ^{18}O$=0.335$T. The time definition of this event is the mean of 12.5 to 12.0 ka minus the mean of 16.5 to 15.5 ka, which is the same as in Buizert et al. (2014). For reference, open circles show the locations of ice-core records used in this reconstruction.

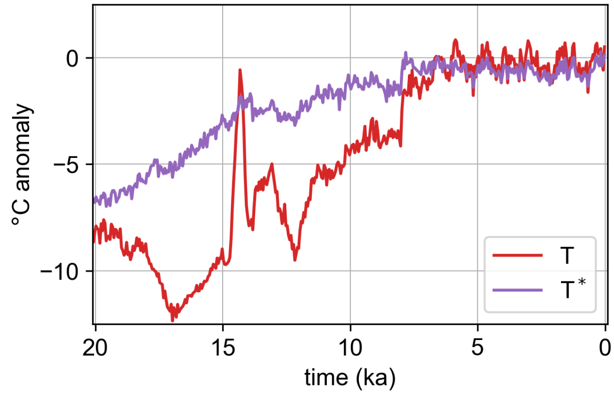

Figure R8. Temperature (T) and temperature weighted by monthly precipitation (T^*) from TraCE-21ka at Summit, Greenland. Both variables are shown as anomalies with respect to 1850-2000 CE and have been averaged to 50-year resolution. T^* was computed before the anomaly was taken.

Table R1. The mean slope for the linear δ^{18}O-temperature relationship used for the main reconstruction in this study (black) and the mean slope for the relationship used in the S4 sensitivity experiment in this study (red). We also include estimates from previous work, inlcuding, slopes found by Buizert et al. (2014) (purple) as estimated from their Fig. 3 for time periods between 20 and 10 ka; slopes found by Guillevic et al. (2013) (green) from their Table 3 for Dansgaard-Oeschger events 8, 9, and 10; slopes found by Kindler et al. (2014) (blue) from their Fig. 5 and estimated from their Fig. 6 for 120 to 10 ka; and slopes found by Cuffey and Clow (1997) (orange) for time periods between 50 and 0.5 ka. The cores are arraged from North (top) to South (bottom).

Core Name	**Slope Range (°C ‰$^{-1}$)**	**Slope Average (°C ‰$^{-1}$)**
Agassiz	0.618 - 0.656 0.412 - 0.437	0.640 0.425
Camp Century	0.439 - 0.468 0.293 - 0.312	0.456 0.304
NEEM	0.450 - 0.480 0.300 - 0.320 ~0.25 - ~0.75 0.51 - 0.63	0.465 0.310 ~ 0.51 0.57
NGRIP	0.454 - 0.489 0.303 - 0.326 ~0.3 - ~0.4 0.34 - 0.47 ~0.3 - ~0.57	0.470 0.313 ~0.37 0.42 0.52
GISP2	0.442 - 0.493 0.294 - 0.329 ~0.1 - ~0.3 0.38 0.251 - 0.465	0.467 0.311 ~0.26 0.324
GRIP	0.442 - 0.493 0.294 - 0.329 0.49	0.467 0.311
Renland	0.546 - 0.595 0.364 - 0.397	0.571 0.381
Dye3	0.424 - 0.475 0.283 - 0.317	0.444 0.296

Table R2. Comparison of abrupt climate transitions in our main reconstruction and sensitivity reconstructions, S1, S2, S3, and S4. We use the same time definitions as in Buizert et al. (2014). Note that the main reconstruction and S4 do not warm as rapidly into the Holocene as S1, S2, and S3. Thus, our use of a single time definition may not allow us capture the full transition for all of these reconstructions.

Core Name	Reconstruction Name	Bølling-Allerød Transition	Younger Dryas	Holocene Transition
Agassiz	Main (δ^{18}O$= 0.67T^{*}$)	12.83	-9.10	11.64
	S1 (δ^{18}O$= 0.67T$)	8.67	-5.16	7.78
	S2 (δ^{18}O$= 0.5T$)	11.20	-6.96	10.50
	S3 (δ^{18}O$= 0.335T$)	15.41	-9.99	14.84
	S4 (δ^{18}O$= 0.67T^{*}$, stronger P seasonality)	15.49	-11.05	13.06
Camp Century	Main	11.71	-8.51	10.53
	S1	8.04	-4.94	7.13
	S2	10.40	-6.60	9.65
	S3	14.34	-9.41	13.71
	S4	14.07	-10.36	11.46
NEEM	Main	10.17	-7.53	9.10
	S1	7.11	-4.51	6.34
	S2	9.22	-5.99	8.33
	S3	12.74	-8.47	12.16
	S4	12.18	-9.20	9.71
NGRIP	Main	9.62	-7.17	8.57
	S1	6.76	-4.33	6.03
	S2	8.77	-5.73	8.14
	S3	12.13	-8.09	11.57
	S4	11.52	-8.77	9.07
GISP2	Main	7.78	-6.03	6.88
	S1	5.66	-3.88	5.18
	S2	7.39	-5.06	6.92
	S3	10.26	-7.03	9.81
	S4	9.27	-7.41	7.03
GRIP	Main	7.78	-6.03	6.88
	S1	5.66	-3.88	5.18
	S2	7.39	-5.06	6.92
	S3	10.26	-7.03	9.81
	S4	9.27	-7.41	7.03
Renland	Main	8.51	-6.27	7.77
	S1	5.93	-3.78	5.39
	S2	7.70	-5.02	7.22
	S3	10.64	-7.08	10.22
	S4	10.15	-7.62	8.53
Dye3	Main	6.45	-5.64	5.41
	S1	5.33	-4.34	5.26
	S2	7.05	-5.45	6.84
	S3	9.92	-7.29	9.60
	S4	7.62	-7.10	4.79

Table R3. Comparison of abrupt climate transitions in our main reconstruction (black) and from Buizert et al. (2014) (purple). Uncertainties are given as standard deviations, and were computed using summation in quadrature. We use the same time definitions as in Buizert et al. (2014).

Core Name	Bølling-Allerød Transition	Younger Dryas	Holocene Transition
Agassiz	12.83 ± 3.49	-9.10 ± 3.65	11.64 ± 3.46
Camp Century	11.71 ± 3.23	-8.51 ± 3.42	10.53 ± 3.24
NEEM	10.17 ± 3.01 8.89 ± 3.70	-7.53 ± 3.2 -4.92 ± 3.45	9.10 ± 3.05 8.41 ± 2.97
NGRIP	9.62 ± 2.90 11.18 ± 5.08	-7.17 ± 3.09 -8.10 ± 4.89	8.57 ± 2.95 10.89 ± 4.59
GISP2	7.78 ± 2.77 14.37 ± 3.55	-6.03 ± 2.96 -9.23 ± 3.59	6.88 ± 2.84 12.46 ± 3.00
GRIP	7.78 ± 2.77	-6.03 ± 2.96	6.88 ± 2.84
Renland	8.51 ± 2.55	-6.27 ± 2.70	7.77 ± 2.57
Dye3	6.45 ± 3.84	-5.64 ± 4.10	5.41 ± 3.96