AGU PUBLICATIONS

Geophysical Research Letters

RESEARCH LETTER
10.1002/2015GL066042

Key Points:
- Precipitation seasonality and inversion temperature changes behind YD-OD $\delta^{18}O$ enigma
- Local processes changes accounting up to 65% of the expected YD-OD $\delta^{18}O$ difference
- Moisture transport changes from the Pacific accounting only up to 20% of the expected YD-OD $\delta^{18}O$ difference

Supporting Information:
- Supporting Information S1

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Citation:
Pausata, F. S. R., and M. Löfverström (2015), On the enigmatic similarity in Greenland $\delta^{18}O$ between the Oldest and Younger Dryas, *Geophys. Res. Lett.*, *42*, doi:10.1002/2015GL066042.

Received 2 SEP 2015
Accepted 13 NOV 2015
Accepted article online 24 NOV 2015

On the enigmatic similarity in Greenland $\delta^{18}O$ between the Oldest and Younger Dryas

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Abstract The last deglaciation (20.0–10.0 kyr B.P.) was punctuated by two major cooling events affecting the Northern Hemisphere: the Oldest Dryas (OD; 18.0–14.7 kyr B.P.) and the Younger Dryas (YD; 12.8–11.5 kyr B.P.). Greenland ice core $\delta^{18}O$ temperature reconstructions suggest that the YD was as cold as the OD, despite a 50 ppmv increase in atmospheric CO_2, while modeling studies suggest that the YD was approximately 4–5°C warmer than the OD. This discrepancy has been surmised to result from changes in the origin of the water vapor delivered to Greenland; however, this hypothesis has not been hitherto tested. Here we use an atmospheric circulation model with an embedded moisture-tracing module to investigate atmospheric processes that may have been responsible for the similar $\delta^{18}O$ values during the OD and YD. Our results show that the summer-to-winter precipitation ratio over central Greenland in the OD is twice as high as in the YD experiment, which shifts the $\delta^{18}O$ signal toward warmer (summer) temperatures (enriched $\delta^{18}O$ values and it accounts for ~45% of the expected YD-OD $\delta^{18}O$ difference). A change in the inversion (cloud) temperature relationship between the two climate states further contributes (~20%) to altering the $\delta^{18}O$-temperature-relation model. Our experiments also show a 7% decrease of $\delta^{18}O$-depleted precipitation from distant regions (e.g., the Pacific Ocean) in the OD, hence further contributing (15–20%) in masking the actual temperature difference. All together, these changes provide a physical explanation for the ostensible similarity in the ice core $\delta^{18}O$ temperature reconstructions in Greenland during OD and YD.

1. Introduction

The last deglaciation started about 20,000 years before present (20.0 kyr B.P.) as a result of a gradually increasing atmospheric CO_2 concentration [*Shakun et al.*, 2012] and summer insolation in the Northern Hemisphere high latitudes [*Clark et al.*, 2009]. The overall warming was, however, punctuated by two major cooling events in the Northern Hemisphere, lasting thousands of years each: the Oldest Dryas, OD, also known as Heinrich event 1 (~18.0 kyr–14.7 kyr B.P.) and the Younger Dryas, YD (~12.8–11.5 kyr B.P.).

Isotope records from ice cores provide unique information of past climate change and have been crucial in our understanding of past climate variability. The relative amount of heavy to light oxygen isotopes ($\delta^{18}O$) sequestered in the ice sheets is assumed to reflect the site temperature (T_s), i.e., the local surface temperature where the precipitation occurred [*Dansgaard*, 1964; *Huber et al.*, 2006]. However, the temperature imprinted in the water isotope records is actually the temperature during the precipitation formation, i.e., the cloud temperature (T_{inv}), which is linearly correlated with the surface site temperature in the present climate [*Krinner et al.*, 1997]; this relationship (T_s versus T_{inv}) may however change in a different climate state [*Krinner et al.*, 1997]. Furthermore, the water isotope composition may also be affected by changes in several local and nonlocal processes, e.g., the precipitation seasonality (summer-to-winter ratio: lower atmospheric temperatures yields lighter—more depleted—$\delta^{18}O$) [*Werner et al.*, 2000], the origin of the water vapor delivered to the site, owing to changes in circulation and transport pathways [*Charles et al.*, 1994; *Langen and Vinther*, 2008], and the isotopic composition in the source region (e.g., changes in sea surface conditions) [*Werner et al.*, 2000; *Breitenbach et al.*, 2010].

Ice core $\delta^{18}O$ temperature reconstructions—assuming a stationary $\delta^{18}O$-temperature relationship—suggest that the YD was at least as cold as the OD event in Greenland [*Liu et al.*, 2012; *Buizert et al.*, 2014] (Figures 1a and 1c). The apparent similarity in temperature during OD and YD is striking, given that atmospheric CO_2 concentration increased by about 50 ppmv between the two events [*Monnin et al.*, 2001; *Shakun et al.*, 2012] (Figure 1b).

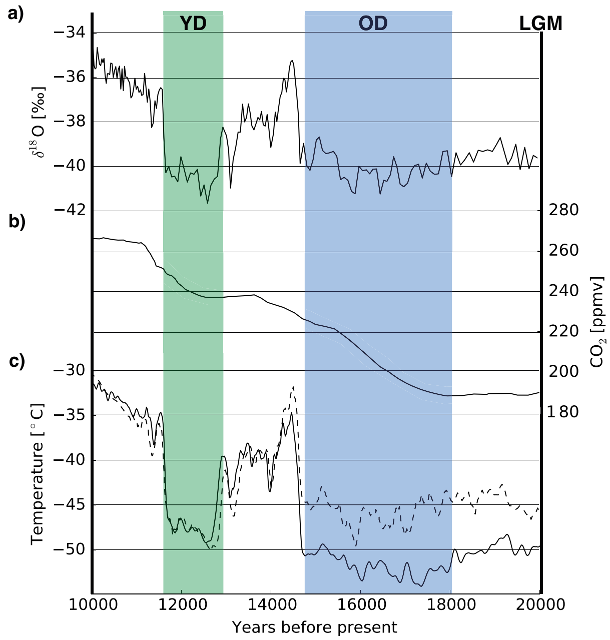

Figure 1. Greenland Ice Sheet Project 2 (GISP2) δ^{18}O values (a), atmospheric CO_2 concentrations (b), and GISP2 surface temperature reconstructions (c) from *Alley* [2000] (dashed line) and a more recent one (solid line) based on δ^{15}N-N2 data from *Buizert et al.* [2014]. Shadings indicate the OD (blue) and YD (green) climate periods.

Recent studies [*Liu et al.,* 2012; *Buizert et al.,* 2014] have addressed this issue, using both model simulations and independent temperature reconstructions, and have shown that the YD was indeed warmer than the OD by approximately 4–5°C. Using an isotope-enabled atmospheric general circulation model, *Liu et al.* [2012] found more depleted δ^{18}O values during the YD than the OD, despite warmer temperatures. They surmised that the lowering of the Laurentide Ice Sheet (by up to 2 km between OD and YD) may have changed the moisture transport from the Pacific Ocean and hence the amount of depleted δ^{18}O delivered to Greenland. A lower and less spatially extensive ice sheet over North America has been found to induce a northward migration of the Atlantic midlatitude jet and storm track [*Carlson et al.,* 2008; *Pausata et al.,* 2011b; *Löfverström et al.,* 2014]; such a shift can potentially increase the moisture delivery to Greenland from the Pacific Ocean, as shown by *LeGrande and Schmidt* [2009] for the mid-Holocene climate and by *Charles et al.* [1994] comparing the Last Glacial Maximum (LGM, ~20.0 kyr B.P.) and the present-day (PD) climate. However, no study has yet investigated the causes of the similar δ^{18}O measurements between YD and OD, despite a temperature difference of 4–5°C.

In this study, we extend the work of *Liu et al.* [2012], examining changes in the Greenland climate and in local and nonlocal processes affecting the isotopic composition of precipitation during **two** key climate periods of the last deglaciation: the OD and the YD. Using a moisture-tracing atmospheric general circulation model, our goal is to test the moisture-delivery-change hypothesis brought about by *Liu et al.* [2012] and to pinpoint key changes in the hydrological cycle between the OD and YD that may provide helpful insights on the apparent paradox in Greenland δ^{18}O temperature reconstructions.

2. Model Description and Experimental Setup

We employ the National Center for Atmospheric Research, Community Atmospheric Model 3 with an embedded moisture-tracing module [*Pausata et al.,* 2011a] to simulate the climate and potential changes in water vapor origin between two different climate states: the Oldest Dryas (OD, ~18.0–14.7 kyr B.P.) and the Younger Dryas (YD, 12.8–11.5 kyr B.P). Each experiment is forced by 55 years history of monthly sea surfacetemperature (SST) and sea ice concentration taken from coupled “Simulation of Transient Climate Evolution over the last 21,000 years” (TraCE-21 ka, [*He,* 2011; *Liu et al.,* 2012]), with the last 50 year used for the analysis (OD: 17,049–17,000 yrs B.P.; YD: 12,149–12,100 yrs B.P.). We adopted the same boundary conditions (solar insolation, atmospheric CO_2, and continental ice sheets) and the same atmospheric model as in TraCE-21 ka and *Liu et al.* [2012], however with increased horizontal resolution: from T31 (~3.75° × 3.75°) to T42 (~2.8° × 2.8°). For reference, we also perform two additional experiments: (1) the Last Glacial Maximum (LGM, ~20.0 kyr B.P.), using prescribed SST and sea ice from the TraCE simulation (20,000–19,951 yrs BP) and (2) the present day (PD), using prescribed SST and sea ice from HadISST (1949–2004) [*Rayner et al.,* 2003]. We trace the water vapor from nine regions (four in the North Atlantic, two in the North Pacific, the Arctic Ocean, the continents and everywhere else; Figure S1 in the supporting information) and follow it until it precipitates, hence allowing us to test *Liu et al.*'s [2012] hypothesis about changes in the transport pathways.

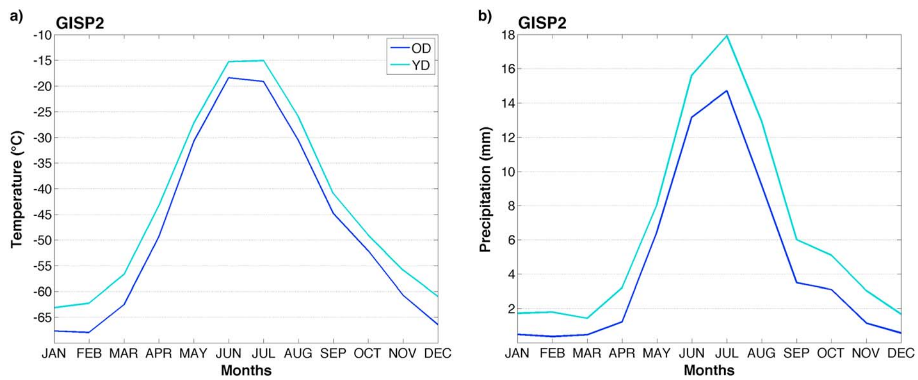

Figure 2. Simulated (a) temperature and (b) precipitation seasonality at GISP2 model location for the OD and YD climate state.

3. Results

We first analyze changes in local processes that can affect the δ^{18}O over Greenland in the OD and YD and then investigate potential variations in the origin of the moisture transported and delivered to Greenland as well as the source temperature changes. Finally, we quantify the relative contribution of local and nonlocal processes for the anomalously enriched δ^{18}O signal recorded in central Greenland in the OD.

3.1. Changes in Local Processes

Some of the most important local processes that can affect the δ^{18}O ratio recorded in ice cores are as follows: seasonal variations in precipitation and in the local temperature at the site (more precisely the cloud temperature where the precipitation forms). A pronounced seasonal cycle of precipitation, with more precipitation in summer than in winter, would shift the δ^{18}O signal toward warmer temperatures than a more homogeneous year-round precipitation distribution. Moreover, changes in the relationship between the site and cloud temperature in different climate state may alter the δ^{18}O-based temperature reconstructions [*Krinner et al.,* 1997], which is based on the observed modern relationship [*Shuman et al.,* 1995].

3.1.1. Changes in Temperature and Precipitation Seasonality

The reference simulations (PD and LGM) agree well with both data reconstructions and other modeling studies (see Figure S1). The annual mean temperature over central Greenland is approximately −30°C in the PD experiment (when accounting for the lower topography due to the model resolution), which is close to observations (~ −31°C). The precipitation shows a year-round snowfall in central Greenland (Table S1 and Figure S2) that amounts to a total of ~400 mm/yr liquid water equivalent (Table S1 and Figure S2), which is higher than modern measurements (~230 mm) [*Steen-Larsen et al.,* 2011]. In the LGM, the simulated annual-mean temperature drops by about 20°C, with a more pronounced cooling in winter than in summer (Figure S2). In central Greenland the annual precipitation amounts to about 60–70 mm and peaks in the summer months (70% of the total snowfall) in agreement with other modeling studies [*Krinner et al.,* 1997; *Werner et al.,* 2000].

The YD and OD experiments show lower annual mean temperatures than the LGM simulation (Table S1). However, the cooling is mostly occurring in winter (~6° in the YD and ~10°C in the OD with respect to the LGM values over central Greenland); summer temperatures do not change appreciably in the OD, whereas the YD summers are almost 5°C warmer than in the LGM. These results confirm the previous finding that abrupt climate changes in the deglaciation were predominantly a winter phenomena [*Denton et al.,* 2005; *Li et al.,* 2005; *Flückiger et al.,* 2008; *Bromley et al.,* 2014; *Buizert et al.,* 2014]. In agreement with *Liu et al.* [2012], both the OD and YD experiments show seasonal temperature variations of about 45°C, but the OD climate is around 4–5°C colder in all seasons (Figure 2). Therefore, the comparable δ^{18}O minima in the OD and YD cannot be directly attributed to differences in the temperature seasonality. On the other hand, the seasonal cycle of precipitation differs considerably in the two experiments: the OD summer-to-winter ratio is twice as large as in the YD and is thus giving more weight to the δ^{18}O-enriched summer signal (Table 1). While the annual-mean site temperature, calculated as a standard arithmetic mean, shows that the OD was 4–5°C colder than the YD, the precipitation-weighted temperature shows no difference over central

Table 1. Summer-to-Winter Precipitation Ratio, Site Temperature (T_s), Site Temperature Weighted by Monthly Mean Precipitation, Inversion Temperature (T_{inv}), Inversion Temperature Weighted by Monthly Mean Precipitation ($T_{inv,pr}$), Source Temperature Accounting for the Percentage of Precipitation Coming From Each Tagged Region (T_{source}) and Precipitation Origin for the YD Climate and the OD Changes Relative to YD at GISP2 Model Location[a]

		Site Temperature (°C)		Cloud Temperature (°C)		Inversion Temperature Relation (°C)		Source Temperature (°C)	Precipitation Origin	
GISP2	Precipitation Summer/Winter Ratio	T_s	$T_{s,pr}$	T_{inv}	$T_{inv,pr}$	$T_{inv} - T_s$	$T_{inv,pr} - T_{s,pr}$	T_{source}	Close	Far
YD	4.3	−42.7	−29.0	−33.0	−24.1	+9.7	+4.9	+4.6	43%	57%
ΔOD	× 1.9	−4.8	0.0	−1.7	−0.8	+3.1	−0.8	−2.0	+7%	−7%

[a]The precipitation ratio changes in the OD are expressed as *times* the YD ratio. The summer and the winter seasons are defined as the warmest (April to September) and the coldest 6 months (October to March), respectively. Close origin comprises the following regions: the northern North Atlantic Ocean, the Arctic Ocean, and Continental recycling; far precipitation origin comprises the Central and Tropical North Atlantic Ocean, the Pacific Ocean, and the rest of the oceans.

Greenland (Table 1). Therefore, changes in the seasonal cycle of precipitation may indeed help explain the comparable δ^{18}O values recorded in Greenland ice cores in the two abrupt transitions.

3.1.2. Changes in Site Versus Cloud Temperature

The δ^{18}O signal recorded in the ice cores is strongly related to the temperature where the precipitation droplets are formed, i.e., the cloud temperature, rather than the surface temperature. The modern cloud-surface temperature relationship is typically adopted to reconstruct surface site temperatures; however, it has been shown that this relationship may change in glacial climates [*Krinner et al.,* 1997; *Werner et al.,* 2000].

We assume that most of the precipitation is formed in the warmest tropospheric layer [*Krinner et al.,* 1997; *Werner et al.,* 2000], which we refer to as the inversion temperature (T_{inv}). The precipitation-weighted inversion temperature ($T_{inv,pr}$) thus represents the local component of the water isotope signal that is the best estimate for changes in δ^{18}O associated with local factors [*Krinner et al.,* 1997; *Werner et al.,* 2000]. Changes in the inversion relationship ($T_{inv} - T_s$) and the seasonality of precipitation can affect the recorded $T_{inv,pr}$ and hence the temperature reconstructions. Our results show a 6–7°C strengthening of the inversion relationship during the LGM compared to the PD (Table S1), in agreement with a previous modeling study using a different atmospheric model [*Werner et al.,* 2000]. The temperature inversion is further strengthened by about 2.5°C during the OD, whereas the YD shows similar values as the LGM (Table S1). On the other hand, the precipitation-weighted inversion relationship ($T_{inv,pr} - T_{s,pr}$) is similar between OD and YD (~1°C difference), and it is also relatively constant across the different climate states, with at most 2–2.5°C difference from the PD values (Table S1). Furthermore, the $T_{inv,pr}$ difference is less than 1°C between the OD and YD climates, pointing to local changes in precipitation seasonality and in the inversion temperature relationship as key components when solving the apparent contradiction in the δ^{18}O temperature reconstructions.

3.2. Changes in Nonlocal Processes

The isotopic composition of Greenland precipitation can also be influenced by changes in the large-scale atmospheric circulation. For example, a long transport pathway generally leads to a more δ^{18}O-depleted precipitation compared to that originating from closer source regions. The precipitation δ^{18}O can also be influenced by the fractionation of oxygen isotopes in the source region (e.g., due to variations in SST, relative humidity and/or wind speed) and changes in cloud microphysics and other atmospheric processes en route to the site [*Jouzel et al.,* 1997]. The moisture-tracing module does not allow us to investigate the latter processes explicitly, but it enables us to evaluate potential changes in the origin of the water vapor delivered to Greenland (see section 2 and Figure S1) and in the temperature at the source region.

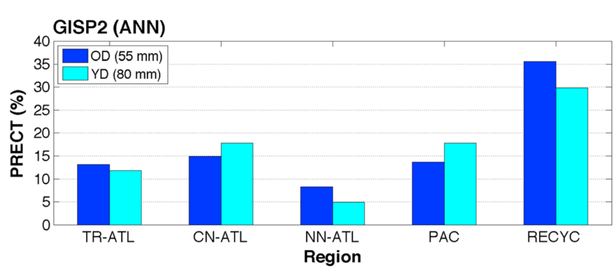

Figure 3. Relative amount of annual precipitation coming from the five most important regions, collectively accounting for approximately 85% of the total annual mean precipitation (Tropical Atlantic, central North Atlantic, northern North Atlantic, Pacific Ocean, and Continental recycling, regions shown in Figure 4) for the OD and YD climate states at GISP2 model location.

Our results show that precipitation from distant locations is about 7% lower over central Greenland in the OD relative to the YD experiment (Table 1). More specifically, the precipitation originating over the Pacific Ocean is ~5% lower in the OD relative to the YD experiment, while the moisture transport from the northern and central North Atlantic Ocean increases of about the same amount (Figure 3).

Therefore, our results confirm the hypothesis proposed by *Liu et al.* [2012] that the amount of strongly depleted Pacific water vapor is lower in the OD compared to the YD simulation, which contributes to increasing the recorded δ^{18}O and hence masking the colder OD temperature.

The source temperatures are about 2°C lower in the OD compared to the YD (Table 1), which yields heavier δ^{18}O and hence decreasing the expected δ^{18}O difference between the two climate states [*Masson-Delmotte et al.*, 2005].

3.3. Effects of Changes in Local and Nonlocal Processes on the δ^{18}O Values

Using a Rayleigh-type distillation model, *Masson-Delmotte et al.* [2005] estimated that the temporal water isotope sensitivity to site temperature changes alone is 0.88 ‰/°C (equations (1) and (2) in section 4.1). Therefore, accounting for changes in sea-water composition between OD and YD (~0.7‰ [*Liu et al.*, 2012]), the 4–5°C temperature difference yields about 3.5–4‰ lighter δ^{18}O in the OD than in the YD (equation (1) in section 4.1). However, the recorded δ^{18}O is about 0.5‰ higher in the OD relative to YD at GISP2 (Figure 1a). Therefore, changes in local and non-local processes must account for the 4–4.5‰ higher δ^{18}O values in the OD than what is expected from the temperature difference.

Shuman et al. [1995] presented an empirical formula to determine the modern relationship between the δ^{18}O and the site temperature (equation (3) in section 4.1). Using the simulated T_s, it is possible to estimate the seasonal cycle of δ^{18}O for each climate state and hence the precipitation-weighted annual mean δ^{18}O ($\delta^{18}O_{wgt}$), which is what is recorded in the ice cores (equation (4) in section 4.1). The simulated precipitation-weighted δ^{18}O over central Greenland is almost identical in both the OD and YD: $\delta^{18}O_{wgt}^{OD} = -34.8‰$ and $\delta^{18}O_{wgt}^{YD} = -34.7‰$, without accounting for the altitude effect associated with the lower model topography (~5.6‰) [see *Liu et al.*, 2012] and differences in the ocean surface composition ($\delta^{18}O_{sw}$).

The OD simulation shows a remarkably different precipitation seasonality compared to the YD, with a much higher summer-to-winter precipitation ratio—giving more weight to the δ^{18}O-enriched summer signal—as well as a variation in the inversion temperature relationship (Table 1). In order to account for the precipitation shift, we assume that the seasonal variations in precipitation in the OD were identical to the YD. The resulting δ^{18}O in the OD is about 1.8‰ lower than the $\delta^{18}O_{wgt}^{OD}$ (approximately $-36.6‰$, see equation (5) in section 4.1). To estimate the changes in the inversion temperature, we use the model-based isotope sensitivity to inversion (cloud) temperature changes of 0.96‰/°C as suggested by *Johnsen et al.* [2001] (equation (6) in section 4.1). When imposing the same inversion temperature relationship (i.e., the same temperature change between the surface and inversion (cloud) layer) as in the YD (equation (7) in section 4.1), the $\delta^{18}O_{wgt}^{OD}$ decreases by ~1.0‰. These calculations suggest that changes in local processes account for approximately 2.8‰ of the δ^{18}O discrepancy in the OD.

Regarding changes in nonlocal processes, *Charles et al.* [1994] suggested that changes in moisture transport from the Atlantic to the Pacific Ocean may yield as much as ~15‰ lower δ^{18}O over central Greenland. Our model simulation suggests a 5% lower moisture delivery from the Pacific Ocean to central Greenland during the OD compared to the YD (Figure 3). Hence, the contribution of the moisture source change may account for about 0.75‰.

Finally, the temporal water isotope sensitivity to source temperature changes is -0.58 ‰/°C (equation (1) in section 4.1). Therefore, the ~2°C lower source temperature in the OD reduces the expected δ^{18}O difference between OD and YD of roughly 1‰. Therefore, the non-local processes may account for about 1.8‰.

In conclusion, our analysis suggests that changes in local processes between OD and YD may explain 60–65% of the difference between the expected and recorded δ^{18}O in the OD. The decreased moisture transport from the Pacific Ocean contributes to about 15–20% of the total difference. The remaining 20–25% may be associated with changes in the source region.

4. Discussion and Conclusions

In summary, our results illustrate the importance of changes in the inversion (cloud) temperature, the seasonality and the source of precipitation when accounting for the higher δ^{18}O values recorded in central and

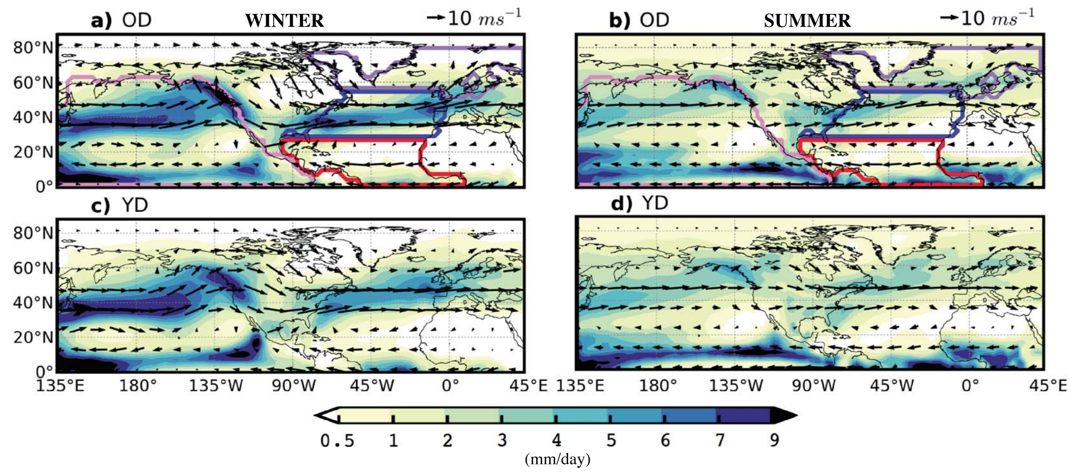

Figure 4. (a and c) Winter (October to March) and (b and d) summer (April to September) precipitation (shading, millimeter per day) and 850 hPa wind (arrows) for the OD (a, b) and YD (c, d) experiments. Bold contours in the upper panels show, respectively, the tagged regions in the Pacific Ocean (pink), the tropical Atlantic Ocean (red), the central North Atlantic Ocean (blue), and the northern North Atlantic Ocean (purple).

southern Greenland during the OD compared to the warmer YD climate. While showing no major changes in the temperature seasonality, our results stress the importance of differences in the temperature inversion relationship ($T_{inv} - T_s$) and in the summer-to-winter precipitation ratio; the latter is about twice as large in the OD as in the YD simulation, thus giving more weight to the δ^{18}O-enriched summer signal in the OD climate (Table 1). The precipitation-weighted inversion temperature, which is what is recorded in the ice cores, shows comparable annual mean values in the OD and YD simulations (Table 1). In addition, our results indicate a 5% lower amount of δ^{18}O-depleted precipitation originated from the Pacific Ocean during the OD, which increases the recorded δ^{18}O and hence further masking the colder temperature relative to the YD climate. However, the impact of the moisture origin changes on the recorded d18O is small compared to the variations in local processes.

The cooler OD temperature compared to YD suggests that the expected δ^{18}O in the OD should have been roughly 4–4.5‰ lower than that measured in the ice cores. A back-of-the-envelope calculation suggests that the impact of the simulated changes in precipitation seasonality, and the inversion temperature relationship between OD and YD increases the δ^{18}O in the OD by about 1.8‰ and 1.0‰, respectively. Therefore, the combined effect of changes in local processes may explain the largest part (60–65%) of this discrepancy. The residual is likely associated with changes in the moisture transport from the Pacific Ocean (~0.75‰ or 15–20% of the discrepancy) and changes in the source region (~1‰ or 20–25% of the discrepancy).

Previous studies [*Charles et al.,* 1994; *LeGrande and Schmidt*, 2009] have shown that a lower Laurentide Ice Sheet yields an increased moisture transport from the Pacific Ocean to Greenland owing to a northward migration of the North Atlantic storm track [*Eisenman et al.,* 2009]. A lower Laurentide Ice Sheet also favors a meridionally tilted Atlantic storm track, both in winter and summer, whereas a higher ice sheet yields stronger seasonal differences with a more zonal and narrow winter storm track (cf. Figures 4a–4d) [see also *Carlson et al.,* 2008; *Pausata et al.,* 2011b; *Löfverström et al.,* 2014], hence largely depriving Greenland of winter precipitation (Figure 4) [see *Liu et al.,* 2012, Figures 3c and 3d]. The comparatively large seasonal variations of the OD storm track relative to YD imply that the winter precipitation makes up an overall smaller fraction of the annual precipitation rate over Greenland, hence providing a larger imprint of the warm summer season in the signal recorded in the ice cores during the OD (Figures 2 and 4).

Modeling studies have suggested that similar variations in precipitation seasonality occurred between the LGM and PD climates [e.g., *Charles et al.,* 1994], and accounting for such variations have been deemed crucial in order to correctly estimate the amplitude of the mean temperature changes between glacial and interglacial climate states [*Masson-Delmotte et al.,* 2005]. Here we show that the precipitation seasonality also varied substantially at different stages in the deglaciation phase.

All together, our results provide a physical explanation for the ostensible similarity in the ice core $\delta^{18}O$ temperature reconstructions and how the colder OD temperature may have been masked by changes in local and nonlocal atmospheric processes. Further studies using climate models with isotope tracing modules, which are able to faithfully reproduce the $\delta^{18}O$ seasonal cycle and the LGM-PD $\delta^{18}O$ shift over Greenland, should be employed to better quantify the relative importance of local and nonlocal changes. Finally, our study confirms that the climate sensitivity, as gauged from ice core $\delta^{18}O$ temperature reconstructions, does not properly capture the amplitude of past abrupt climatic shifts and may therefore underestimate future rapid responses of the Greenland ice sheet to global warming.

4.1. Methods

Using a Rayleigh-type distillation model, *Masson-Delmotte et al.* [2005] presented the following equation to calculate changes in $\delta^{18}O$ relative to the present day:

$$\Delta\delta^{18}O = 0.88 \cdot \Delta T_s - 0.58 \cdot \Delta T_{source} + 0.9 \cdot \Delta\delta^{18}O_{sw} \ , \quad (1)$$

where ΔT_s is the change in site temperature, ΔT_{source} is the change in source temperature, and $\Delta\delta^{18}O_{sw}$ is the change in the sea-water isotopic composition between the present day and a given climate state. Hence, it follows that $\delta^{18}O$ variations, owing to changes in site temperature alone, can be expressed as follows:

$$\Delta\delta^{18}O = 0.88 \cdot \Delta T_s + 0.9 \cdot \Delta\delta^{18}O_{sw} \ . \quad (2)$$

Given a 4–5°C temperature difference between OD and YD (ΔT_s) and $\Delta\delta^{18}O_{sw} = 0.7$ ‰, the expected $\delta^{18}O$ in the OD should be ~3.5–4‰ lower than in the YD (i.e., 4–4.5‰ lower than recorded in the OD over central Greenland).

Shuman et al. [1995] determined the modern relationship between the $\delta^{18}O$ and the site temperature (T_s) to be as follows:

$$\delta^{18}O = -148.04 + 0.46403 \cdot T_s \ . \quad (3)$$

Using the simulated T_s (in Kelvin), it is possible to estimate the seasonal cycle of $\delta^{18}O$ for each climate state (*) and hence the precipitation-weighted $\delta^{18}O$:

$$\delta^{18}O^*_{wgt} = \frac{\sum_{j=1}^{12} \delta^{18}O^*_j \cdot PRECT^*_j}{\sum_{j=1}^{12} PRECT^*_j}, \quad (4)$$

where the $PRECT^*_j$ is the monthly (index j) precipitation for a given climate state.

To isolate the impact of changes in precipitation seasonality between the OD and the YD, we weight the $\delta^{18}O^{OD}_j$ by the monthly precipitation from the YD simulation $\left(PRECT^{YD}_j\right)$:

$$\delta^{18}O^{OD-YDprt}_{wgt} = \frac{\sum_{j=1}^{12} \delta^{18}O^{OD}_j \cdot PRECT^{YD}_j}{\sum_{j=1}^{12} PRECT^{YD}_j} \ , \quad (5)$$

where $\delta^{18}O^{OD-YDprt}_{wgt}$ is the weighted $\delta^{18}O^{OD}$ value assuming that the precipitation seasonality in the OD was equal to that in the YD.

Johnsen et al. [2001] show that the inversion (cloud) temperature has a higher water isotope sensitivity compared to the surface temperature (0.46‰/°C, equation (3)) and suggested a value of 0.96 ‰/°C.

$$\delta^{18}O = 0.96 \cdot T_{inv} - 10. \quad (6)$$

To isolate the impact of variations in the inversion temperature relationship, we substitute the OD inversion $\left(T^{OD}_{inv} - T^{OD}_s\right)$ with the one from the YD $\left(T^{YD}_{inv} - T^{YD}_s\right)$ and recalculate the inversion temperature for the OD:

$$T^{OD-YDinv}_{inv} = T^{OD}_{inv} + \left(\Delta T^{YD-OD}_{inv}\right) \ , \quad (7)$$

where ΔT^{YD-OD}_{inv} is the change in the inversion temperature relationship $\left(T^*_{inv} - T^*_s\right)$ between YD and OD (the annual mean $\Delta T^{YD-OD}_{inv} = -3.1$°C, see Table 1). In other words, the $T^{OD-YDinv}_{inv}$ is assuming that the OD and the YD had the same temperature difference between the surface and the inversion layer. Using $T^{OD-YDinv}_{inv}$ in equations (6) and (4), the resulting $\delta^{18}O_{wgt}$ is about 1‰ lower than that calculated using T^{OD}_{inv}, suggesting that changes in the temperature inversion relationship accounts for around 1‰ of the total signal.

Acknowledgments
The authors would like to thank Qiong Zhang and two anonymous reviewers for insightful comments on the manuscript, David Noone and James White for helpful discussions, and Feng He for providing the TraCE data set. The authors also acknowledge GISP2 for making the ice cores temperatures and $\delta^{18}O$ available and EPICA for providing CO_2 measurements. The simulations were performed on resources provided by the Swedish National Infrastructure for Computing (SNIC) at National Supercomputing Centre (NSC), Linköping. Data for this paper can be requested to F.S.R.P. (francesco.pausata@misu.su.se). F.S.R.P. conceived and designed the research; M.L. performed the model simulations; F.S.R.P. and M.L. analyzed the results; F.S.R.P. led the writing, and M.L. contributed to the writing of the manuscript.

References

Alley, R. B. (2000), The Younger Dryas cold interval as viewed from central Greenland, *Quat. Sci. Rev.*, *19*(1–5), 213–226, doi:10.1016/S0277-3791(99)00062-1.

Breitenbach, S. F. M., J. F. Adkins, H. Meyer, N. Marwan, K. K. Kumar, and G. H. Haug (2010), Strong influence of water vapor source dynamics on stable isotopes in precipitation observed in Southern Meghalaya, NE India, *Earth Planet. Sci. Lett.*, *292*(1–2), 212–220, doi:10.1016/j.epsl.2010.01.038.

Bromley, G. R. M., A. E. Putnam, K. M. Rademaker, T. V. Lowell, J. M. Schaefer, B. Hall, G. Winckler, S. D. Birkel, and H. W. Borns (2014), Younger Dryas deglaciation of Scotland driven by warming summers, *Proc. Natl. Acad. Sci. U.S.A.*, *111*(17), 6215–6219, doi:10.1073/pnas.1321122111.

Buizert, C., et al. (2014), Greenland temperature response to climate forcing during the last deglaciation, *Science*, *345*(6201), 1177–1180, doi:10.1126/science.1254961.

Carlson, A. E., A. N. LeGrande, D. W. Oppo, R. E. Came, G. A. Schmidt, F. S. Anslow, J. M. Licciardi, and E. A. Obbink (2008), Rapid early Holocene deglaciation of the Laurentide ice sheet, *Nat. Geosci.*, *1*(9), 620–624, doi:10.1038/ngeo285.

Charles, C. D., D. Rind, J. Jouzel, R. D. Koster, and R. G. Fairbanks (1994), Glacial-interglacial changes in moisture sources for greenland: Influences on the ice core record of climate, *Science*, *263*(5146), 508–511, doi:10.1126/science.263.5146.508.

Clark, P. U., A. S. Dyke, J. D. Shakun, A. E. Carlson, J. Clark, B. Wohlfarth, J. X. Mitrovica, S. W. Hostetler, and A. M. McCabe (2009), The Last Glacial Maximum, *Science*, *325*(5941), 710–714, doi:10.1126/science.1172873.

Dansgaard, W. (1964), Stable isotopes in precipitation, *Tellus*, *16*(4), 436–468, doi:10.1111/j.2153-3490.1964.tb00181.x.

Denton, G., R. Alley, G. Comer, and W. Broecker (2005), The role of seasonality in abrupt climate change, *Quat. Sci. Rev.*, *24*(10–11), 1159–1182, doi:10.1016/j.quascirev.2004.12.002.

Eisenman, I., C. M. Bitz, and E. Tziperman (2009), Rain driven by receding ice sheets as a cause of past climate change, *Paleoceanography*, *24*, PA4209, doi:10.1029/2009PA001778.

Flückiger, J., R. Knutti, J. W. C. White, and H. Renssen (2008), Modeled seasonality of glacial abrupt climate events, *Clim. Dyn.*, *31*(6), 633–645, doi:10.1007/s00382-008-0373-y.

He, F. (2011), Simulating transient climate evolution of the last deglaciation with CCSM3, PhD thesis, Dep. of Atmos. and Oceanic Sci., Univ. of Wisconsin-Madison.

Huber, C., M. Leuenberger, R. Spahni, J. Flückiger, J. Schwander, T. F. Stocker, S. Johnsen, A. Landais, and J. Jouzel (2006), Isotope calibrated Greenland temperature record over Marine Isotope Stage 3 and its relation to CH_4, *Earth Planet. Sci. Lett.*, *243*(3–4), 504–519, doi:10.1016/j.epsl.2006.01.002.

Johnsen, S. J., D. Dahl-Jensen, N. Gundestrup, J. P. Steffensen, H. B. Clausen, H. Miller, V. Masson-Delmotte, A. E. Sveinbjörnsdottir, and J. White (2001), Oxygen isotope and palaeotemperature records from six Greenland ice-core stations: Camp Century, Dye-3, GRIP, GISP2, Renland and NorthGRIP, *J. Quat. Sci.*, *16*(4), 299–307, doi:10.1002/jqs.622.

Jouzel, J., et al. (1997), Validity of the temperature reconstruction from water isotopes in ice cores, *J. Geophys. Res.*, *102*, 26,471–26,487, doi:10.1029/97JC01283.

Krinner, G., C. Genthon, and J. Jouzel (1997), GCM analysis of local influences on ice core δ signals, *Geophys. Res. Lett.*, *24*, 2825–2828, doi:10.1029/97GL52891.

Langen, P. L., and B. M. Vinther (2008), Response in atmospheric circulation and sources of Greenland precipitation to glacial boundary conditions, *Clim. Dyn.*, *32*(7–8), 1035–1054, doi:10.1007/s00382-008-0438-y.

LeGrande, A. N., and G. A. Schmidt (2009), Sources of Holocene variability of oxygen isotopes in paleoclimate archives, *Clim. Past*, *5*(3), 441–455, doi:10.5194/cp-5-441-2009.

Li, C., D. S. Battisti, D. P. Schrag, and E. Tziperman (2005), Abrupt climate shifts in Greenland due to displacements of the sea ice edge, *Geophys. Res. Lett.*, *32*, L19702, doi:10.1029/2005GL023492.

Liu, Z., et al. (2012), Younger Dryas cooling and the Greenland climate response to CO_2, *Proc. Natl. Acad. Sci. U.S.A.*, *109*(28), 11,101–11,104, doi:10.1073/pnas.1202183109.

Löfverström, M., R. Caballero, J. Nilsson, and J. Kleman (2014), Evolution of the large-scale atmospheric circulation in response to changing ice sheets over the last glacial cycle, *Clim. Past*, *10*(4), 1453–1471, doi:10.5194/cp-10-1453-2014.

Masson-Delmotte, V., J. Jouzel, A. Landais, M. Stievenard, S. J. Johnsen, J. W. C. White, M. Werner, A. Sveinbjornsdottir, and K. Fuhrer (2005), GRIP deuterium excess reveals rapid and orbital-scale changes in Greenland moisture origin, *Science*, *309*(5731), 118–121, doi:10.1126/science.1108575.

Monnin, E., A. Indermühle, A. Dällenbach, J. Flückiger, B. Stauffer, T. F. Stocker, D. Raynaud, and J. M. Barnola (2001), Atmospheric CO_2 concentrations over the last glacial termination, *Science*, *291*(5501), 112–114, doi:10.1126/science.291.5501.112.

Pausata, F. S. R., D. S. Battisti, K. H. Nisancioglu, and C. M. Bitz (2011a), Chinese stalagmite $\delta^{18}O$ controlled by changes in the Indian monsoon during a simulated Heinrich event, *Nat. Geosci.*, *4*(7), 474–480, doi:10.1038/ngeo1169.

Pausata, F. S. R., C. Li, J. J. Wettstein, M. Kageyama, and K. H. Nisancioglu (2011b), The key role of topography in altering North Atlantic atmospheric circulation during the last glacial period, *Clim. Past*, *7*(4), 1089–1101, doi:10.5194/cp-7-1089-2011.

Rayner, N. A., D. E. Parker, E. B. Horton, C. K. Folland, L. V. Alexander, D. P. Rowell, E. C. Kent, and A. Kaplan (2003), Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century, *J. Geophys. Res.*, *108*(D14), 4407, doi:10.1029/2002JD002670.

Shakun, J. D., P. U. Clark, F. He, S. A. Marcott, A. C. Mix, Z. Liu, B. Otto-Bliesner, A. Schmittner, and E. Bard (2012), Global warming preceded by increasing carbon dioxide concentrations during the last deglaciation, *Nature*, *484*(7392), 49–54, doi:10.1038/nature10915.

Shuman, C. A., R. B. Alley, S. Anandakrishnan, J. W. C. White, P. M. Grootes, and C. R. Stearns (1995), Temperature and accumulation at the Greenland Summit: Comparison of high-resolution isotope profiles and satellite passive microwave brightness temperature trends, *J. Geophys. Res.*, *100*, 9165–9177, doi:10.1029/95JD00560.

Steen-Larsen, H. C., et al. (2011), Understanding the climatic signal in the water stable isotope records from the NEEM shallow firn/ice cores in northwest Greenland, *J. Geophys. Res.*, *116*, D06108, doi:10.1029/2010JD014311.

Werner, M., U. Mikolajewicz, M. Heimann, and G. Hoffmann (2000), Borehole versus isotope temperatures on Greenland: Seasonality does matter, *Geophys. Res. Lett.*, *27*, 723–726, doi:10.1029/1999GL006075.