Moisture Source Tagging Confirming the Polar Amplification Effect in Amplifying the Temperature-$\delta^{18}$O Temporal Slope Since the LGM

Jian Guan 1,*, Zhengyu Liu 1,2 and Guangshan Chen 3

1 Laboratory for Climate, Ocean and Atmosphere Studies, Department of Atmospheric and Oceanic Sciences, School of Physics, Peking University, Beijing 100871, China; liu.7022@osu.edu
2 Atmospheric Science Program, Department of Geography, Ohio State University, Columbus, OH 43210, USA
3 Institute of Earth Environment, Chinese Academy of Sciences, Xi’an 710061, China; gchen9@gmail.com

* Correspondence: lotusescy@gmail.com

Received: 8 May 2020; Accepted: 7 June 2020; Published: 9 June 2020

Abstract: Stable water isotopologues in paleoclimate archives ($\delta^{18}$O) have been widely used as an indicator to derive past climate variations. The modern observed spatial $\delta^{18}$O-temperature relation in the middle and high latitudes has been used to infer the paleotemperatures changes from ice core data. However, various studies have shown that the spatial slope is larger than the temporal slope at the drill site by a factor of 2. Physically, the different spatial and temporal slope has been suggested to result from the amplified local surface air temperature cooling in the polar region at Last Glacial Maximum (LGM), according to the slope ratio equation derived in our previous study. To explicitly confirm the “polar amplification” effect in understanding the differences between temporal and spatial isotope–temperature relations, here we use the same isotope-enabled atmospheric general circulation model with a moisture-tracing module embedded to quantitatively estimate the contributions of different sources to the precipitated heavy oxygen isotopes in the middle and high latitudes. Our results show that the major sources of $\delta^{18}$O in precipitation over middle and high latitudes are from oceans where the sea surface temperature cooling at Last Glacial Maximum (LGM) is less than $-2^\circ$C, while the local moisture sources with a higher cooling can be also relevant for polar regions, such as north Greenland. Additionally, the neglect of the strengthened local inversion layer strength at LGM could be the main cause for the overestimated source temperature cooling by the slope ratio equation, especially for the polar regions in the Northern Hemisphere.

Keywords: water isotope–temperature relation; temporal slope; spatial slope; polar amplification; moisture tagging

1. Introduction

The ratio of stable oxygen isotopes $^{18}$O/$^{16}$O (hereafter $\delta^{18}$O) in paleoclimate archives (i.e., ice cores, speleothems) has been established as a well-recognized, most useful tool in deriving essential information of past climate changes during the last 22,000 years. On its way typically from low latitudes to high latitudes, an air parcel undergoes continuous depletion of $^{18}$O in the remaining water vapor via condensation processes. Various theoretical studies have shown that the precipitated $\delta^{18}$O can reflect the local temperature variations [1–3] at high latitudinal regions (the so-called temperature effect).

The observed modern (spatial) relations between $\delta^{18}$O and surface air temperature (SAT) of polar sampling sites [2,4,5] are taken as transfer functions (spatial slope, $\sim 0.65\% \times ^\circ$C$^{-1}$) to interpret temporal changes of $\delta^{18}$O at the drill site. Later studies, however, show that the temporal relation between $\delta^{18}$O in precipitation and SAT (temporal slope) is about half of the spatial slopes [6–8].
The smaller temporal slope exists not only in the polar region where the ice cores are located but also in the middle and high latitudes according to the model evidence [9,10]. By implementing stable water isotopes in atmospheric general circulation models (AGCMs), the reduced temporal slope relative to spatial slope in the local region can be qualitatively interpreted by the changes of precipitation in seasonality [8,11], moisture sources [12–14], atmospheric circulation and meteorological processes [15–17] and evaporation recharge processes over ocean [8]. To quantitatively interpret the difference between temporal and spatial slope over the middle and high latitudes, a semi-empirical and semi-analytical slope ratio equation \( \frac{\alpha_t}{\alpha_s} = 1 - \left( \frac{\Delta T_0 + \Delta T_d}{\Delta T} \right) \) was promoted [9] based on Boyle’s (1997) [18] conceptual model under two major assumptions: (1) the spatial slope only has a slight change between Last Glacial Maximum (LGM) and present and is fixed as 0.6 % × °C⁻¹, and (2) according to the Rayleigh distillation theory, the local precipitated \( \delta^{18}O \) will change ~0.55% while the source temperature changes 1 °C. Here, \( \alpha_t \) and \( \alpha_s \) represent the temporal and spatial isotope-temperature slope of some local region, respectively; \( \Delta T_0 \) is for the source moisture temperature change; \( \Delta T \) is for the local surface air temperature change and \( \Delta T_d \) is the equivalent source temperature rescaled based on the additional effects on the slope ratio, such as the ice volume effect which equals about ~2 °C cooling at LGM, and \( r \) is a semiempirical parameter, which is close to 1. According to the slope ratio equation, the temporal slope is usually smaller than the spatial slope as long as the local temperature change is greater than the source temperature change. Within a backward approach test, by assuming the slope ratio equation predicts the model temporal slope/spatial slope ratio perfectly, a major moisture source of the ocean where the sea surface temperature (SST) having ~2 °C cooling at LGM for the precipitation at high latitudes was implied, and it was further indirectly suggested that the “polar amplification” feature in global climate change is the main cause of a smaller temporal slope compared with the spatial slope over the middle and high latitudes. It should be noted that here, the polar amplification refers primarily to the amplified local SAT change in the polar area comparing with the corresponded moisture sources temperature change.

However, several issues still remain. Firstly, the contributions of different moisture source changes to the local precipitated isotopic content at middle and high latitudes in the LGM and present days were not quantified in our previous study. Hence, it remains unclear that the degree of the polar amplification of SAT from LGM to present can affect the difference between the temporal and spatial temperature–oxygen isotope slope.

In this study, we conduct a suite of source water tagging experiments with an isotope-enabled AGCM to quantify the isotopic contributions from different pre-defined moisture sources to the precipitated isotopic content over middle and high latitudes and further investigate how polar amplification of the SAT correlated to the temperature–oxygen relation based on slope ratio equation theory. The study area focused in this paper are the middle and high latitudes sub-regions, which were the same as in the previous study [9] (shown in Figure 1c).

This paper consists of the following sections: firstly, Section 2 will describe the model and how the tagging experiments are conducted. Section 3 will introduce the climate of isotopic signature from each predefined tagging region to the sub-regions over middle and high latitudes. In Section 4, we will examine the estimated source temperature changes using a backward method by comparing with the results of tagging experiments. Finally, we will summarize the results and further discuss the validity of the slope ratio equation in Section 5.
Figure 1. Selected tagging regions in isoCAM3. (a) Tagging regions over the continents; (b) tagging regions over the ocean. The tagging regions index over the ocean is listed on the right side of the color bar according to corresponding sea surface temperature (SST) cooling between Last Glacial Maximum (LGM) and pre-industrial (PI). The prefix “tro” means tagging regions over the ocean; “n” stands for the tagging regions over the Northern Hemisphere, which are marked by circles; “s” stands for the tagging regions over the Southern Hemisphere, which are marked by squares, and “eq” stands for the tagging regions over equatorial oceans, which are marked by diamonds. All the markers shown in (a, b) represent the model grid cell locations; (c) surface air temperature difference between LGM and PI simulations. The boxes show the locations of the sub-regions (about 30 degrees in longitude, 12–15 degrees in latitude) in the middle and high latitudes in each hemisphere. In addition, the 0-degree contour line is shown in a solid black line.
2. Model Description and Experiments

The results reported in this study are mainly based on an isotope-enabled atmospheric component model of the Nation Center for Atmospheric Research Community Climate System Model version 3, isoCAM3 [9,19]. The experiments were performed in T31 resolution (about 3.75° × 3.75° in horizontal, 26 levels in vertical) running for 10 years under LGM and pre-industrial (PI) conditions with the same external forcing as for the 22 ka and 00 ka snapshot run in Guan et al. (2016) [9], and in addition, by a 10-year history of interannually varying monthly sea surface temperature (SST) and sea ice cover from the same period in the Transient simulation of Climate Evolution (TraCE) experiment in Liu et al. (2009) [20].

In total, we defined 11 different tagging areas over land (Figure 1, panel a) and 12 different tagging areas over the ocean (Figure 1, panel b). For land surfaces, each continent was defined as a different evaporation source except Greenland and Eurasia, which take more consideration of the amplitude of SAT difference between LGM and PI. Greenland was separated into southern (S. Greenland) and northern part (N. Greenland), and Eurasia was separated into three parts, from south to north (Mid-East–SE Asia, Mid-Eurasia and N. Eurasia). For ocean surfaces, the amplitude of annual mean sea surface temperature difference of each grid box between LGM and PI was chosen as the criterion to define different evaporation source areas, as shown in Figure 1b: (1) \(-9^\circ C \geq \Delta SST\), tro n07; (2) \(-7^\circ C \geq \Delta SST \geq -9^\circ C\), tro n06; (3) \(-5^\circ C \geq \Delta SST \geq -7^\circ C\), tro s04 and tro n05; (4) \(-3^\circ C \geq \Delta SST \geq -5^\circ C\), tro s03 and tro n04; (5) \(-1^\circ C \geq \Delta SST \geq -3^\circ C\), tro s02, tro eq01 and tro n03; (6) \(1^\circ C \geq \Delta SST \geq -1^\circ C\), tro s01 and tro n02; (7) \(\Delta SST \geq 1^\circ C\): tro n01. Here, the prefix “tro” means tagging regions over ocean, “n” is for the Northern Hemisphere, “s” is for the Southern Hemisphere and “eq” is for tagging regions close to equatorial oceans. The extratropics over the middle and high latitudes in each hemisphere are divided into 24 subregions (Figure 1c). Each subregion is about 30° in longitude and 12–15° in latitude.

In addition, since the seasonality of precipitation does not affect the differences between temporal and spatial slope too much over a broader area in isoCAM3 [9], we will not discuss the seasonality effect for a certain local region. But for some local regions such as the Greenland summit, the seasonality of precipitation could be relevant to the discrepancy between spatial and temporal isotope–temperature relations. To interpret the universal spatial and temporal slope difference and exam the slope ratio equation, we only discuss the annual mean isotope–temperature relationships in this study. We also noticed that the great warming occurs in the North Pacific and Alaska at LGM, which is mainly caused by the topography effect of ICE-5G ice sheet in isoCAM3 [21–23].

3. Climate of Isotopic Signature in Precipitation

We first analyzed the isotopic signature in precipitation from different predefined tagging regions for the sub-regions at the middle and high latitudes (Figure 2). In Figure 2, each line shows the spatial pattern of the isotopic source contribution from its corresponding tagging source region to the sub-regions in the Northern Hemisphere (Figure 2a, LGM; 2c, PI) and Southern Hemisphere (Figure 2b, LGM; 2d, PI) by using the percentage of precipitated O^{18} from the sources. The solid and dashed lines are for tagging regions over ocean and land, respectively. For a certain sub-region, the values over different lines show how many percentages of the area-weighted O^{18} in precipitation evaporates from the corresponding tagging regions. It should be noticed that the percentage of the area-weighted O^{18} in precipitation from tagging regions is also related to the area of the tagging regions.

For the Northern Hemisphere, the majority of the O^{18} in precipitation of sub-regions in the Northern Hemisphere and Southern Hemisphere both evaporated from the tropical and subtropical ocean at PI (Figure 2c), solid black line, tro eq01. For one thing, the largest rate of evaporation is in subtropics, which are almost covered by tro eq01. In addition, the total contribution of a single tagging region is also correlated with its coverage over the ocean. Generally, the major moisture source of precipitation at middle and high latitudes is located at the tropical and subtropical ocean, which is consistent with previous studies. However, for some local sub-regions (i.e., N04, N05, N06, northern Greenland; N18, Greenland–Iceland–Nordic Sea), the isotopic contribution from local moisture sources...
(i.e., tro n02, tro n03) is comparable with that from the remote sources at low latitude, or even more. In addition, the local moisture sources are also much more important for some inland sub-regions (i.e., N22, Siberia). Compared with the PI stage, the isotopic signal from different tagging regions does not change too much at LGM, but for northern Greenland, the local sources become more important because of the southward displacement of storm track forced by the Laurentide ice sheet. Most of the O\(^{18}\) in precipitation over middle and high latitudes is evaporated from the ocean surface in the tropical and subtropical ocean. Nevertheless, the changes of the contributions from local moisture sources are complicated for different sub-regions. Some of the source changes are related to the changes of moisture path (i.e., N01, N02, North Alaska) because of the ice sheet effect at LGM [23].

For the Southern Hemisphere, the major moisture source of the isotopic signal in precipitation is from the oceans. The moisture source from southern oceans (tro s02 and tro s03) are comparable with the sources at the tropical and subtropical ocean, which account for 25% on average. The remote sources over land (South America, South Africa and Australia) are less important when comparing with the sources over ocean and are also dependent on the distance off the land. In addition, it is interesting that the local moisture source over Antarctica mostly contributes to the isotopic signal in precipitation over western Antarctica in isoCAM3, which might imply the topography effect on the local precipitation over Antarctica.

4. Equivalent Source Temperature

To explain the differences between spatial and temporal \(\delta^{18}O\)–temperature slopes, a semi-empirical and semi-theoretical slope ratio equation was promoted in Guan et al. (2016) [9] under two important assumptions by following Boyle’s mechanism (1997) [18]: (1) the \(\delta^{18}O\) in precipitation is mainly dominated by a Rayleigh distillation process, and (2) the spatial slope does not change substantially between LGM (Northern Hemisphere, 0.49%o × °C\(^{-1}\); Southern Hemisphere, 0.50%o × °C\(^{-1}\)) and present (Northern Hemisphere, 0.55%o × °C\(^{-1}\); Southern Hemisphere, 0.63%o × °C\(^{-1}\)). However, the slope
ratio equation is only tested through a theoretical backward method. In the backward method test, the source temperature change ($\Delta T_0$) can be estimated according to the slope ratio equation by assuming the temporal slope derived in the model is perfectly predicted. Here, we derived the source temperature and its difference between LGM and PI based on tagging experiments, and further explicitly discussed the validity of the slope ratio equation in understanding the temporal and spatial slope differences.

For a local region, $\delta^{18}\text{O}(x,t)$ in precipitation of a specific location $x$ at time $t$ can be written as a function of local SAT $T(x,t)$ and source temperature $T_0(t)$ as follows [9]. Here, $\alpha_s$ represents the spatial slope between the local SAT and $\delta^{18}\text{O}$ in precipitation, and $b$ is constant (~$0.55\%_\text{o} \times ^\circ\text{C}^{-1}$) according to the Rayleigh distillation process, which represents the relation between the moisture source temperature and $\delta^{18}\text{O}$ in precipitation. $d(t)$ represents the processes other than Rayleigh distillation that can affect the $\delta^{18}\text{O}(x,t)$, such as local evaporation recharge [8].

$$\delta^{18}\text{O}(x,t) = \alpha_s T(x,t) - bT_0(t) + d(t) \quad (1)$$

Since $\delta^{18}\text{O}$ is the ratio of mixing ratio ($R^{18}$),

$$\delta^{18}\text{O}(x,t) = \frac{R^{18}(x,t)}{R_{\text{standard}}^{18}} = \frac{O^{18}(x,t)}{O_{\text{standard}}^{18}} \quad (2)$$

Here, the $O^{18}(x,t)$ and $O^{16}(x,t)$ consist of the local precipitated $O^{18}$ and $O^{16}$ from different tagging sources $i$, namely $O^{18}(x,t,i)$ and $O^{16}(x,t,i)$ ($i = 1,2,\ldots, N-1,N$).

$$O^{18}(x,t) = \sum_{i=1}^{N} O^{18}(x,t,i) \quad (3)$$

$$O^{16}(x,t) = \sum_{i=1}^{N} O^{16}(x,t,i) \quad (4)$$

So, we have

$$\sum_{i=1}^{N} O^{18}(x,t,i) = (\alpha_s T(x,t) - bT_0(t) + d(t)) \times R_{\text{standard}}^{18} \times \sum_{i=1}^{N} O^{16}(x,t,i) \quad (5)$$

We assumed that, the isotopic signal in precipitation at local $x$ from a single source $i$, namely $\delta^{18}(x,t,i)$, is related to local SAT $T(x,t)$ and the corresponding source temperature $T_0(t,i)$, which can be written as follows. Here, the $\alpha_s(i)$ represents the spatial slope between the $\delta^{18}(x,t,i)$ and $T(x,t)$. The parameter $b$ remains the same value in Equation (1), since it is a theoretical parameter estimated according to the Rayleigh distillation process. Therefore,

$$\delta^{18}\text{O}(x,t,i) = \alpha_s(i) T(x,t) - bT_0(t,i) + d(t,i) \quad (6)$$

then,

$$O^{18}(x,t,i) = \alpha_s(i) T(x,t) - bT_0(t,i) + d(t,i) \times R_{\text{standard}}^{18} \times O^{16}(x,t,i) \quad (7)$$

Considering the terms of source temperature,

$$\sum_{i=1}^{N} T_0(t,i) \times O^{16}(x,t,i) = T_0(t) \times \sum_{i=1}^{N} O^{16}(x,t,i) \quad (8)$$

namely,

$$T_0(t) = \frac{\sum_{i=1}^{N} T_0(t,i) \times O^{16}(x,t,i)}{\sum_{i=1}^{N} O^{16}(x,t,i)} \equiv T_{x0}(t) \quad (9)$$

Here, we assumed that $\alpha_s(i)$ remains the same in a specific climate stage according to Figure 3, which shows the distribution of spatial slopes calculated from different tagging source regions where the moisture source contributions are larger than 10%.
We assumed that the isotopic signal in precipitation at local \( x \) from a single source \( i \), namely \( \delta^{18}O_{x,t,i} \), is related to local SAT \( T(x,t) \) and the corresponding source temperature \( T_0(t,i) \), which can be written as follows. Here, the \( \alpha_0(i) \) represents the spatial slope between the \( \delta^{18}O_{x,t,i} \) and \( T(x,t) \).

\[
\delta^{18}O_{x,t,i} = \alpha_0(i) T(x,t) - b T_0(t,i) + d(t,i)
\]

(6)

Then,

\[
O^{18}O_{x,t,i} = (\alpha_0(i) T(x,t) - b T_0(t,i) + d(t,i)) \times R_{012}^{\alpha_0}
\]

(7)

Considering the terms of source temperature,

\[
T_0(t) = \frac{\sum T_0(t,i) \times O^{18}O_{x,t,i}}{\sum O^{18}O_{x,t,i}}
\]

(8)

namely,

\[
T_0(t) = \sum T_0(t,i) \times O^{18}O_{x,t,i}
\]

(9)

Here, we assumed that \( \alpha_0(i) \) remains the same in a specific climate stage according to Figure 3, which shows the distribution of spatial slopes calculated from different tagging source regions where the moisture source contributions are larger than 10%.

Figure 3. The spatial slopes over middle and high latitudes in each hemisphere at LGM and PI, respectively. The spatial slopes were calculated using \( \delta^{18}O \) of tagging sources and local surface air temperature, namely, \( \alpha_0(i) \) in Equation (6). The box shows the range from the 25th percentile to the 75th percentile. The bottom line stands for the 10th percentile and the top line stands for the 90th percentile. The red line represents the mean value of spatial slope.

With the assumption of a stable \( \alpha_0(i) \), the source temperature of a local sub-region \( T_0(t) \) can be written as the temperature weighted by the contributions from different moisture sources (hereafter \( T_{e0} \)). For a specific location \( x \) at time \( t \), a higher \( T_{e0} \) represents that a greater proportion of precipitation comes from the remote sources at low latitude, where the source temperature is warmer than the local source. On the contrary, a lower \( T_{e0} \) is mainly caused by more contribution from local and cooler sources. Figure 4a,b shows \( T_{e0} \) at LGM and PI, respectively. It is obvious that \( T_{e0} \) is higher than the local SAT (Figure 1c) or sea surface temperature (not shown) at middle and high latitudes, as a consequence of the main source of precipitation being from the tropical and subtropical ocean (tro eq01). In addition, \( T_{e0} \) decreases poleward in both the Northern Hemisphere and the Southern Hemisphere because more water vapor is evaporated from the local source. This can also be inferred somehow from the increasing difference of zonal mean \( T_{e0} \) between LGM and PI in the poleward direction (supplement materials, Figure S2).
Atmosphere 2020, 11, x of 12

Precipitation comes from the remote sources at low latitude, where the source temperature is warmer than the local source. On the contrary, a lower $T_{e0}$ is mainly caused by more contribution from local and cooler sources. Figure 4a,b shows $T_{e0}$ at LGM and PI, respectively. It is obvious that $T_{e0}$ is higher than the local SAT (Figure 1c) or sea surface temperature (not shown) at middle and high latitudes, as a consequence of the main source of precipitation being from the tropical and subtropical ocean. In addition, $T_{e0}$ decreases poleward in both the Northern Hemisphere and the Southern Hemisphere because more water vapor is evaporated from the local source. This can also be inferred somehow from the increasing difference of zonal mean $T_{e0}$ between LGM and PI in the poleward direction (supplement materials, Figure S2).

**Figure 4.** The equivalent source temperature ($T_{e0}$) derived based on Equation (9) in isoCAM3. (a) LGM, (b) PI, (c) LGM-PI.

We compared the area-weighted $T_{e0}$ changes with the required source temperature cooling derived from the slope ratio equation using a backward method for all sub-regions (Figure 5, blue dots). The tagging experiments show that most of $T_{e0}$ cooling is about $-2$ °C and $-4$ °C (Figure 6a), in which $-2$ °C cooling is more dominant, which is consistent with the backward method test. However, quantitatively, the slope ratio equation implies an overestimated source temperature cooling by ~5 °C and ~3.5 °C systematically in the polar sub-regions of the Northern (N01, N02, …, N12) and Southern (S01, S02, …, S12) Hemispheres. This can be partly caused by a fixed spatial slope ($0.6\% \times ^\circ C^{-1}$) used in the backward method when calculating the required source temperature cooling for all sub-regions. According to the slope ratio equation, more source temperature cooling will be derived if the local spatial slope is fixed to a larger value. After using the zonal mean spatial slope, the source temperature cooling for polar sub-regions (purple line, Figure S1) is close to the results of
the tagging experiment. In addition, the inaccuracy of the slope ratio equation, by neglecting the local inversion layer strength changes, could also induce an inaccurate deduction of the source temperature cooling. After considering the local inversion layer strength difference between the LGM and PI (Equation (S3)), the estimated source temperature cooling derived by the backward method (Figure 5, red dots) is modified closer to the results of tagging experiments, especially over the polar regions in the Northern Hemisphere. However, it should be mentioned as well that some other causes still remain unknown that could be relevant over eastern Antarctica. Overall, the tagging experiments explicitly confirmed that the polar amplification of the SAT is the main cause of the reduced temporal slopes over middle and high latitudes both in the Northern Hemisphere and Southern Hemisphere, which is implied by the slope ratio equation.

![Figure 5](image-url)

**Figure 5.** The comparison between the area-weighted equivalent source temperature cooling ($\Delta T_{eq}$) in the tagging experiment and the estimated source temperature cooling based on the backward method for all the sub-regions in the Northern Hemisphere ((a), polar; (c), subpolar) and Southern Hemisphere ((b), polar; (d), subpolar). The blue dots represent the results of the backward method in Guan et al., (2016) [9], and the red dots represent the results of the improved backward method after considering the local inversion layer strength effect.
Figure 6. The histogram of estimated source temperature changes. (a) gives the results of tagging experiments in this study; (b) cites the results of the backward method (Figure 9c in [9]) using the slope ratio equation; (c) gives the results of the backward method after considering local inversion layer effect.

5. Conclusions

In this study, we investigated different moisture sources’ contribution to the isotopic signal in precipitation over the middle and high latitudes at LGM and PI, respectively. By comparing the results of the tagging experiments and that of the backward method using slope ratio equation, we further confirmed the “polar amplification” effect in understanding the difference between temporal and spatial $\delta^{18}O$-temperature relations.

The main conclusions are listed below:

First, the tagging experiments show that the dominant moisture source region of $\delta^{18}O$ lies in the ocean, where the sea surface temperature has a $\sim2^\circ C$ cooling at LGM. Meanwhile, the local moisture sources can also be relevant, especially for precipitation over the polar regions. It is consistent with the results suggested from the backward method test using the slope ratio equation [9]. As shown in Figure S2, the zonal mean equivalent source temperature has a cooling of 3–4 $^\circ C$ over middle and high latitudes at LGM, which is much smaller than local SAT cooling. Figure 6 further shows the source temperature changes calculated using tagging experiments in isoCAM3 (a) and backward method (b). The tagging results and backward method both show a dominant 2 $^\circ C$ source temperature cooling at LGM, except that the backward method overestimates source temperature cooling for polar regions. Generally, the two methods demonstrate that the local SAT cooling is amplified compared with the moisture source temperature cooling; subsequently, this “polar amplification” effect induced the reduced local temporal slope according to the slope ratio equation.
Second, the equivalent source temperature changes derived in tagging experiments show a smaller amplitude but a similar latitudinal pattern in a majority of sub-regions when comparing with the backward method results, especially over the polar regions at both hemispheres. The overestimation of source temperature cooling by slope ratio equation implies that the difference of temporal-spatial δ18O-temperature can be affected systematically by some other mechanisms in the polar regions beyond polar amplification, such as local inversion layer strength changes. In other words, a steeper temporal slope will be derived based on the slope ratio equation if we only take “polar amplification” effect between local SAT difference and source temperature change into consideration at polar regions.

Overall, our modeling experiments identified the remote (less cooling) and the local (more cooling except for the area covered mainly by sea ice) moisture source contributions to the precipitation over the middle and high latitudes and further proved that the “polar amplification” effect in global climate change is the main reason for a reduced temporal slope between temperature and oxygen isotopes in precipitation at middle and high latitudes. Moreover, for the local precipitation in the polar region, there are still some factors that are important as well (e.g., local inversion layer strength changes) and some remain unknown which can change the temporal slope as well, such as topography changes and local weathering processes. It should also be noticed and cautioned that the averaged local spatial slope has a slight increase at present when comparing LGM in isoCAM3 for both Northern Hemisphere and Southern Hemisphere (Figure 3 and also in [9]). The slight change of the spatial slope could also induce the difference between the temporal and spatial slopes not only in the model but also in the real world.

Supplementary Materials: The following are available online at http://www.mdpi.com/2073-4433/11/6/610/s1, Figure S1: The estimated source temperature changes of the sub-regions in the Northern Hemisphere (left) and Southern Hemisphere (right), Figure S2: The difference between LGM and PI for zonal mean of annual mean sea surface temperature in TraCE (blue), annual mean surface air temperature (red) and T_0 (yellow) in isoCAM3.

Author Contributions: The manuscript was prepared by J.G. The manuscript and experiments were designed by J.G. and Z.L. The experiments were performed by J.G., and G.C. provided the technical support. Z.L. All the authors reviewed the manuscript and contributed to improving the manuscript. All authors have read and agreed to the published version of the manuscript.

Funding: This research received no external funding.

Acknowledgments: This work is supported by Chinese NSF 41630527 and by the US National Science Foundation (NSF P2C2). We are grateful to Xinyu Wen and previous reviewers for helpful discussions. The model output of the tagging experiments using isoCAM3 can be obtained by sending a written request to the corresponding author (Jian Guan, lotusecy@gmail.com).

Conflicts of Interest: The authors declare no conflict of interest.

References
1. Aristarain, A.J.; Jouzel, J.; Pourchet, M. Past Antarctic Peninsula climate (1850–1980) from an ice isotope record. *Clim. Chang.* 1986, 8, 69–86. [CrossRef]
2. Dansgaard, W. Stable isotopes in precipitation. *Tellus* 1964, 16, 436–468. [CrossRef]
3. Jouzel, J.; Alley, R.B.; Cuffey, K.M.; Dansgaard, W.; Grootes, P.; Hoffmann, G.; White, J. Validity of the temperature reconstruction from water isotopes in ice cores. *J. Geophys. Res.* 1997, 102, 26471–26487. [CrossRef]
4. Johnsen, S.J.; Dansgaard, W.; White, J.W.C. The origin of Arctic precipitation under present and glacial conditions. *Tellus* 1989, 41, 452–468. [CrossRef]
5. Lorius, C.; Merlivat, L.; Jouzel, J.; Pourchet, M. A 30,000-yr isotope climatic record from Antarctic ice. *Nature* 1979, 280, 644–648. [CrossRef]
6. Cuffey, K.M.; Clow, G.D.; Alley, R.B.; Stuiver, M.; Waddington, E.D.; Saltus, R.W. Large Arctic temperature change at the Wisconsin-holocene glacial transition. *Science* 1995, 270, 455–458. [CrossRef]
7. Jouzel, J. Calibrating the isotopic paleothermometer. *Science* 1999, 286, 910–911. [CrossRef]
8. Lee, J.E.; Fung, I.; DePaolo, D.J.; Otto-Bliesner, B. Water isotopes during the Last Glacial Maximum: New general circulation model calculations. *J. Geophys. Res. Atmos.* 2008, 113, D19109. [CrossRef]
9. Guan, J.; Liu, Z.; Wen, X.; Brady, E.; Noone, D.; Zhu, J.; Han, J. Understanding the temporal slope of the temperature-water isotope relation during the deglaciation using isoCAM3: The slope equation. *J. Geophys. Res. Atmos.* 2016, 121, 10-342. [CrossRef]

10. Werner, M.; Haese, B.; Zhang, X.; Butzin, M.; Lohmann, G.; Xu, X. Glacial-interglacial changes in H218O, HDO and deuterium excess-results from the fully coupled ECHAM5/MPI-OM Earth system model. *Geosci. Model Dev.* 2016, 9, 647–670. [CrossRef]

11. Werner, M.; Mikolajewicz, U.; Heimann, M.; Hoffmann, G. Borehole versus isotope temperatures on Greenland: Seasonality does matter. *Geophys. Res. Lett.* 2000, 27, 723–726. [CrossRef]

12. Charles, C.D.; Rind, D.H.; Jouzel, J.; Koster, R.D.; Fairbanks, R.G. Glacial-interglacial changes in moisture sources for Greenland: Influences on the ice core record of climate. *Science* 1994, 263, 508–511. [CrossRef] [PubMed]

13. Liu, Z.; Carlson, A.E.; He, F.; Brady, E.C.; Otto-Bliesner, B.L.; Briegleb, B.P.; Wehrenberg, M.; Clark, P.U.; Wu, S.; Cheng, J.; et al. Younger Dryas cooling and the Greenland climate response to CO2. *Proc. Natl. Acad. Sci. USA* 2012, 109, 11101–11104. [CrossRef] [PubMed]

14. Werner, M.; Heimann, M.; Hoffmann, G. Isotopic composition and origin of polar precipitation in present and glacial climate simulations. *Tellus Ser. B* 2001, 53, 53–71. [CrossRef]

15. Hendricks, M.B.; DePaolo, D.J.; Cohen, R.C. Space and time variation of δ18O and δD in precipitation: Can paleotemperature be estimated from ice cores? *Glob. Biogeochem. Cycles* 2000, 14, 851–861. [CrossRef]

16. Lee, J.E.; Fung, I.; DePaolo, D.J.; Henning, C.C. Analysis of the global distribution of water isotopes using the NCAR atmospheric general circulation model. *J. Geophys. Res.* 2007, 112, D16306. [CrossRef]

17. Noone, D. The influence of midlatitude and tropical overturning circulation on the isotopic composition of atmospheric water vapor and Antarctic precipitation. *J. Geophys. Res.* 2008, 113, D04102. [CrossRef]

18. Boyle, E.A. Cool tropical temperatures shift the global δ18O-T relationship: An explanation for the ice core δ18O-borehole thermometry conflict? *Geophys. Res. Lett.* 1997, 24, 273–276. [CrossRef]

19. Noone, D.; Sturm, C. Comprehensive dynamical models of global and regional water isotope distributions. In *Isoscapes: Understanding Movement, Patterns, and Process on Earth Through Isotope Mapping*; West, J.B., Bowen, G.J., Dawson, T.E., Tu, K.P., Eds.; Springer: Dordrecht, The Netherlands, 2010; pp. 195–219.

20. Liu, Z.; Otto-Bliesner, B.L.; He, F.; Brady, E.C.; Tomas, R.; Clark, P.U.; Carlson, A.E.; Lynch-Stieglitz, J.; Curry, W.; Brook, E.; et al. Transient simulation of last deglaciation with a new mechanism for Bølling-Allerød warming. *Science* 2009, 325, 310–314. [CrossRef]

21. Bromwich, H.D.; Toraricnta, E.R.; Wei, H.; Oglesby, R.J.; Fastook, J.L.; Hughes, T.H. Polar MM5 simulations of the winter climate of the Laurentide Ice Sheet. *J. Clim.* 2004, 17, 3415–3433. [CrossRef]

22. CLIMAP Project Members. *Seasonal Reconstructions of the Earth’s Surface at the Last Glacial Maximum*; Geological Society of America: Boulder, CO, USA, 1981; 18p.

23. Otto-Bliesner, B.L.; Brady, E.C.; Ciauzet, G.; Tomas, R.; Levis, S.; Kothavala, Z. Last Glacial Maximum and Holocene Climate in CCSM3. *J. Clim.* 2006, 19, 2526–2544. [CrossRef]

© 2020 by the authors. Licensee MDPI, Basel, Switzerland. This article is an open access article distributed under the terms and conditions of the Creative Commons Attribution (CC BY) license (http://creativecommons.org/licenses/by/4.0/).