# Effects of LGM sea surface temperature and sea ice extent on the isotope-temperature slope at polar ice core sites

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**Abstract.** Stable water isotopes in polar ice cores are widely used to reconstruct past temperature variations over several orbital climatic cycles. One way to calibrate the isotope-temperature relationship is to apply the present-day spatial relationship as a surrogate for the temporal one. However, this method leads to large uncertainties because several factors like the sea surface conditions or the origin and the transport of water vapor influence the isotope-temperature temporal slope. In this study, we investigate how the sea surface temperature (SST), the sea ice extent and the strength of the Atlantic Meridional Overturning Circulation (AMOC) affect these temporal slopes in Greenland and Antarctica for Last Glacial Maximum (LGM, ~21 000 years ago) to preindustrial climate change. For that, we use the isotope-enabled atmosphere climate model ECHAM6-wiso, forced with a set of sea surface boundary condition datasets based on reconstructions (e.g., GLOMAP) or MIROC 4m simulation outputs. We found that the isotope-temperature temporal slopes in East Antarctic coastal areas are mainly controlled by the sea ice extent, while the sea surface temperature cooling affects more the temporal slope values inland. ~~Mixed effects on~~On the other hand, ECHAM6-wiso simulates the impact of sea ice extent on EDC and Vostok sites through the contribution of water vapor from lower latitudes. Effects of sea surface boundary condition changes on modeled isotope-temperature temporal slopes are ~~simulated~~various in West Antarctica ~~with sea surface boundary conditions changes, because~~. This is due partly to the transport of water vapor from the Southern Ocean to this area that can dampen the influence of local temperature on the changes ~~of~~in the isotopic composition of precipitation and snow. In the Greenland area, the isotope-temperature temporal slopes are influenced by the sea surface temperatures ~~very~~ near the coasts of the continent. The greater the LGM cooling off the coast of southeast Greenland, the greater the transport of water vapor from the North Atlantic, and the larger the temporal slopes. The presence or absence of sea ice very near the coast has a large influence in Baffin Bay and the Greenland Sea and influences the slopes at some inland ice ~~cores~~core stations. ~~We emphasize that the~~The extent ~~far south~~ of the sea ice ~~is not so important. On~~far south slightly influences the temporal slopes in Greenland through the ~~other hand, the~~transport of more depleted water vapor from lower latitudes to this area. The seasonal variations of sea ice distribution, especially its retreat in

summer, influence the isotopic composition of the water vapor transport in this region and the modeled isotope-temperature

35  temporal slopes in the eastern part of Greenland. A stronger LGM AMOC decreases LGM to preindustrial isotopic anomalies in precipitation in Greenland, degrading the isotopic model-data agreement. The AMOC strength does not modifymodifies the temporal slopes over inner Greenland slightly, and onlyby a little on the coasts along the Greenland Sea where the changes in surface temperature and sea ice distribution due to the AMOC strength mainly occur.

## 1 Introduction

40  Stable isotopologues of water ($H_2^{16}O$, $H_2^{18}O$ and $HD^{16}O$, called hereafter stable water isotopes) are integrated tracers of climate processes occurring in diverse parts of the hydrological cycle (Craig and Gordon, 1965; Dansgaard, 1964). Because of their differences in mass and symmetries, an isotopic fractionation happens at each phase change of water. This process is reflected by a change in the water isotope ratio values, expressed hereafter in the usual δ notation (as $\delta^{18}O$ and $\delta^{2}H$ with respect to the Vienna Standard Mean Ocean Water V-SMOW if not stated otherwise). As a result, water isotopes have been widely used to

45  describe past variations of the Earth's climate. For example, their measurements in polar ice cores made it possible to reconstruct the temperature variations over several glacial-interglacial cycles (Jouzel et al., 2007; Jouzel, 2013, and references therein; NEEM Community Members, 2013).

For such a reconstruction, the present-day isotope-temperature spatial slope can be taken as a surrogate for the temporal gradient at a given site. For example, a spatial slope of 0.80 ‰ °C$^{-1}$ for $\delta^{18}O$ in Antarctica was calculated based on a compilation

50  of measured surface temperatures and $\delta^{18}O$ of snow at various locations in the continent (Masson-Delmotte et al., 2008). However, this method often leads to a large error in the temperature reconstructions because the temporal isotope-temperature slope depends on many factors like the sea surface temperature (SST) (Risi et al., 2010), the sea ice extent (Noone and Simmonds, 2004), the ice sheet elevation (Werner et al., 2018), the origin and the transport of water vapor (Casado et al., 2018). For example, it has been suggested that the relationship between temperature and the isotopic signature for warmer

55  interglacial periods in East Antarctica can vary among ice core sites, with an error in the temperature reconstruction that can reach up to 100 % (Sime et al., 2009; Cauquoin et al., 2015). In Greenland, the use of the spatial relationship between the $\delta^{18}O$ in Greenland ice core records and surface temperature to evaluate the local temperature variations during the last deglaciation leads to a large uncertainty of a factor of 2 (Jouzel, 1999; Buizert et al., 2014). Recently, Buizert et al. (2021) proposed a reconstruction of surface cooling in Antarctica during the Last Glacial Maximum (LGM, ~21 000 years ago) using borehole

60  thermometry and firn properties of different ice cores. Based on these results, they proposed new estimates of temporal $\delta^{18}O$-temperature slopes at these ice core stations, varying from 0.8 to 1.45 ‰ °C$^{-1}$.

The LGM is a period with full glacial conditions and represents the beginning of the last deglaciation. It is one of the key climate periods chosen by the Paleoclimate Modeling Intercomparison Project (PMIP, Kageyama et al., 2018, 2021) because it allows to evaluate how well state-of-the-art models are able to simulate climate changes as large as those expected in the

65  future. In addition to being very different from the preindustrial climate (PI), the LGM period also offers a wealth of isotope

proxy data, including stable water isotopes in polar ice cores for an in-depth comparison with outputs from isotope-enabled models (Lee et al., 2008; Risi et al., 2010; Werner et al., 2016, 2018).

One way to capture the physical processes influencing the temporal isotope-temperature slope in polar regions is the use of Atmospheric General Circulation Models (AGCMs) equipped with prognostic stable water isotopes. Such models can simulate different climate conditions, like LGM and PI periods. Moreover, the use of isotope-enabled AGCMs in combination with isotopic observations allows us to investigate the physical processes controlling the variations of isotopic delta values at a given site. This method makes it possible to estimate the temporal isotope-temperature slope for LGM to preindustrial climate change (Lee et al., 2008; Risi et al., 2010; Werner et al., 2018). Even if such models simulate various temporal isotope-temperature slopes, implying that processes like water vapor transport, post-depositional effects, or polar atmospheric boundary layer are poorly or not represented (Krinner et al., 1997; Werner et al., 2000; Casado et al., 2018), these models are very useful for evaluating the sensitivity of the temporal slopes to parameters like the change of elevation (Werner et al., 2018). Ocean surface conditions are one of the factors that influences LGM-PI isotope changes (Risi et al., 2010; Noone and Simmonds, 2004). Two reconstructions of SST and one of sea ice extent during the LGM period have been released recently. Paul et al. (2021) reconstructed both the SST and the sea ice extent fields, based on faunal and floral assemblage data of the Multiproxy Approach for the Reconstruction of the Glacial Ocean Surface (MARGO) project and several recent estimates of the LGM sea ice extent. The Data-Interpolation Variational Analysis (DIVA) software was used to optimally interpolate sparse SST reconstruction data. The resulting reconstruction was called GLOMAP (Glacial Ocean Map). Tierney et al. (2020) reconstructed the LGM SST field with a different method, by combining a large collection of geochemical proxies for sea surface temperature with simulations outputs from the isotope-enabled model iCESM1.2 (Brady et al., 2019) using an offline data assimilation technique to produce a field reconstruction of LGM temperatures. Tierney et al. (2020) LGM cooling is globally larger than in GLOMAP (3.6°C and 1.7°C, respectively), with possible impacts on LGM to PI isotope changes and their temporal relationship with near surface air temperature. In addition, other SST and sea ice fields, with different characteristics compared to the reconstructions of LGM sea surface conditions described above, can be extracted from atmosphere-ocean coupled model simulations like MIROC 4m (Obase and Abe-Ouchi, 2019).

In the present study, we investigate the impacts of SST and sea ice boundary conditions on the isotope-temperature temporal slope at polar ice core sites for LGM to PI changes. For that Are air temperatures near the surface and the isotopic composition of precipitation in the polar regions influenced by LGM to PI changes in SST and sea ice distribution in the same way? What are the underlying dynamics, for example, in terms of changes in concentrations and transport of water vapor? To answer to these questions, we performed multiple simulations with the isotope-enabled AGCM ECHAM6-wiso driven by different LGM SST and sea ice boundary conditions. We evaluate the modeled LGM-PI $\delta^{18}O$ anomalies with available observations and we investigate how the SST and the sea ice extent patterns influence the model-data agreement on a global scale and at polar ice core stations. The influence of ocean circulation, particularly the strength of the Atlantic Meridional Overturning Circulation (AMOC), on sea surface conditions and by extension on our modeled $\delta^{18}O$ of meteoric water is also investigated. Finally, the

impacts of the sea surface boundary conditions on the $\delta^{18}$O-temperature slopes for LGM-to-preindustrial climate change are evaluated and discussed for Greenland and Antarctic ice core stations.

## 2 Methodology

### 2.1 ECHAM6-wiso

ECHAM6 (Stevens et al., 2013) is the sixth generation of the atmospheric general circulation model ECHAM, developed at the Max Planck Institute for Meteorology. It consists of a dry spectral-transform dynamical core, a transport model for scalar quantities other than temperature and surface pressure, a suite of physical parameterizations for the representation of diabatic processes, and boundary datasets for externalized parameters (trace gas and aerosol distributions, land surface properties, etc.). ECHAM6 forms the atmospheric component of the fully coupled Earth system model MPI-ESM (Giorgetta et al., 2013; Mauritsen et al., 2019). The implementation of the water isotopes in ECHAM6 as part of MPI-ESM has been described and evaluated in detail by Cauquoin et al. (2019b), and this model version has been labeled ECHAM6-wiso. At a later stage, Cauquoin and Werner (2021) updated the water isotope module of ECHAM6-wiso in several aspects. The supersaturation has been slightly re-tuned, the kinetic fractionation factors for the evaporation over the ocean are now assumed as independent of wind speed, and the isotopic content of snow on sea ice is taken into account for sublimation processes in sea ice covered regions. The latter leads to a stronger depletion of surface water vapor over such sea ice covered areas (while the surface temperature remains the same). As a consequence, this change is expected to contribute to a steeper temporal isotope-temperature slope over sea ice covered areas.

### 2.2 Sea surface temperature and sea ice extent boundary conditions for LGM conditions

### 2.2.1 SST

Tierney et al. (2020) SST reconstruction has a larger and more homogeneous cooling than GLOMAP, except for the high southern latitudes at which the Pacific sector cools more than the Atlantic sector (Figure 1). On the other hand, the LGM cooling in the Northern North Atlantic Ocean is stronger in GLOMAP than in Tierney et al. reconstruction (-5.4°C and -4.8°C, respectively, see Table 4 in Paul et al., 2021). These differences between the two SST reconstructions are due to the use of different proxy datasets for the reconstructions (geochemical proxies only for Tierney et al., MARGO dataset for GLOMAP) and to the methods applied to produce SST gridded maps from scattered observations (see Section 1). For their offline data assimilation technique, Tierney et al. (2020) used results from the coupled climate model iCESM1.2, which shows one of the largest cooling among the PMIP4 models (Figure 1b of Kageyama et al., 2021). In addition to these two reconstructions, we used SST and sea ice extent outputs from a MIROC 4m LGM simulation (Obase and Abe-Ouchi, 2019) with oscillating AMOC strength. The global LGM cooling is between -2.3 and -2.7°C according to the considered simulations (Figure 1) i.e., higher than GLOMAP and lower than the Tierney et al. reconstruction. The main specificity of MIROC 4m LGM SST is a very strong

cooling in the North Atlantic (more than 10°C, Figure 1) and more uniform temperature anomalies between -2 and -4°C in the other areas, including off the coast of Greenland. We extracted the MIROC 4m SST outputs, averaged over a 100-year period, at two different times of the LGM simulation depending on the AMOC strength: during a weak AMOC phase (average AMOC index was equal to 8.44 Sv) and a strong AMOC phase (19.95 Sv). A weaker AMOC during LGM implies larger cooling in the North Atlantic (Figure 1) and more extended sea ice (Figure 2), while it does less cooling in the Southern Ocean. The strong AMOC phase period in MIROC 4m simulation was selected in the middle of the AMOC peak (Figure S1). Therefore, the values of MIROC 4m average near surface air temperature in Antarctica are very similar regardless the selected AMOC phase. For example, MIROC 4m simulates LGM temperature of -41.87 and -41.75°C in WDC station for strong and weak AMOC phase, respectively. A similar pattern is found for the eastern part of the continent (-56.80 and -56.50°C in Dome Fuji for strong and weak AMOC phase, respectively).
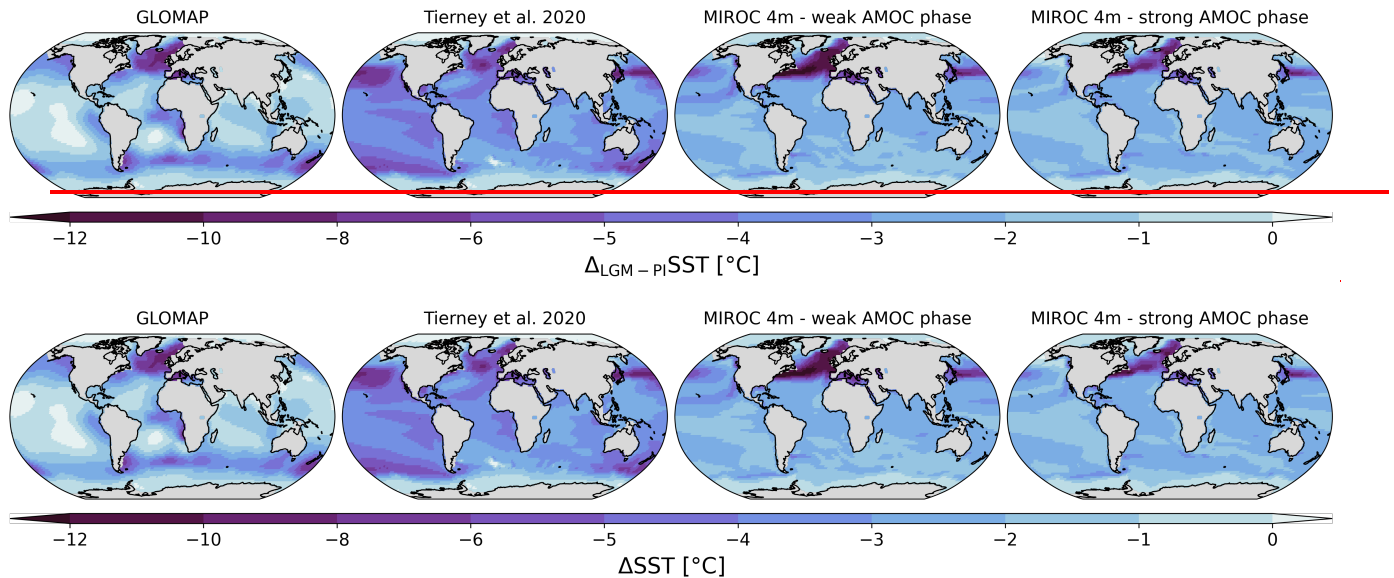


**Figure 1: LGM-PI sea surface temperature changes used as boundary conditions for ECHAM6-wiso simulations. From left to right: GLOMAP (Paul et al., 2021), Tierney et al. (2020), MIROC 4m with weak LGM AMOC phase and MIROC 4m with strong LGM AMOC phase. Anomalies are expressed relative to Atmospheric Model Intercomparison Project mean SST over the period 1870 to 1899.**

### 2.2.2 Sea ice extent

Maps of the averaged sea ice area fraction used as boundary forcings for ECHAM6-wiso are shown in Figure 2. The PI AMIP and LGM GLOMAP sea ice cover is higher around Antarctica compared to MIROC 4m ones, with a further extent in the Southern Ocean especially in the Atlantic sector. On the other hand, sea ice is more extensive in the Northern Hemisphere for MIROC 4m in the weak AMOC phase. For the stronger AMOC case, a decline of the sea ice in the Northern Hemisphere is seen, accompanied by weaker cooling (see section 2.2.1). In its parameterization, MIROC 4m uses a threshold of 95% for the sea ice fraction to allow sub-grid "sea-ice leads". This threshold is not rigid, but it is difficult to exceed sea ice concentrations

of 95% unless there is significant convergence of sea-ice. Consequently, while the sea ice is, on average, more extensive in the north in MIROC 4m for the weak AMOC phase compared to GLOMAP reconstruction, the sea ice area fraction in grid cells near coastal areas like Greenland is lower in MIROC 4m than in GLOMAP (95-98% against 100%, respectively).
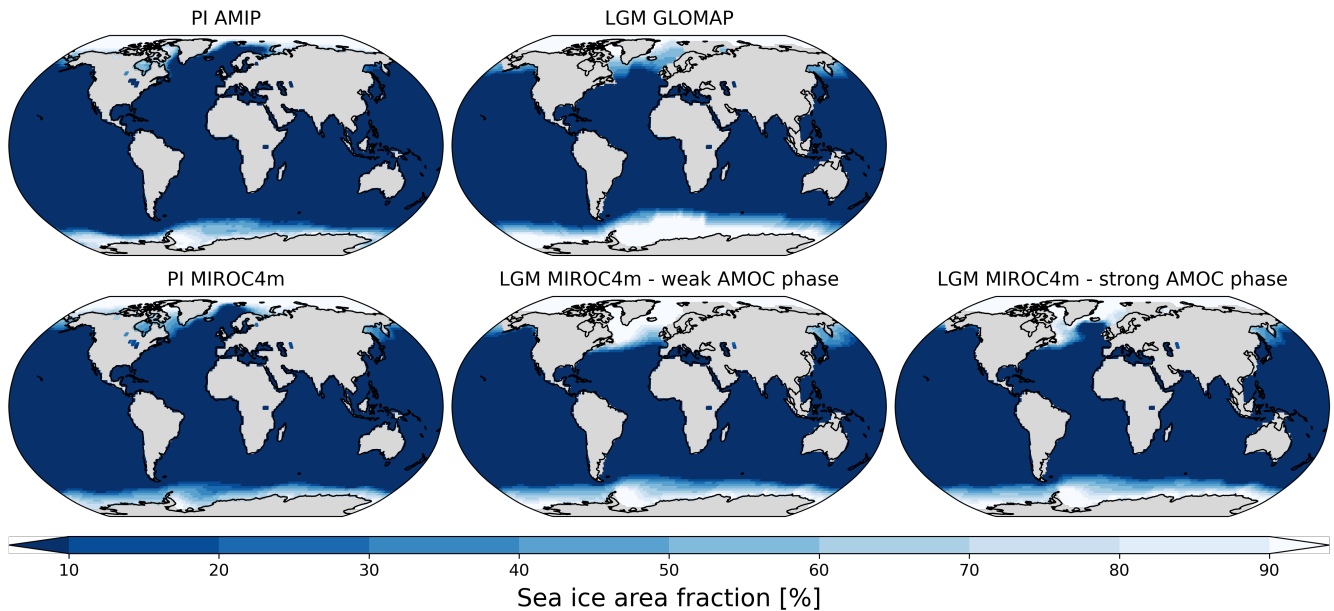


**Figure 2: LGM and PI sea ice area fractions used as boundary conditions for ECHAM6-wiso simulations.**

## 2.3 Model setup and experiments

We performed an ensemble of LGM simulations with ECHAM6-wiso, forced with different combinations of SST and sea ice boundary forcings presented in section 2.2. All our PI and LGM simulations are designed with respect to the PMIP4 protocol (e.g., greenhouse gas and orbital conditions). The LGM SST boundary fields are expressed relative to the Atmospheric Model Intercomparison Project (AMIP, Eyring et al., 2016) mean SST (averaged over the period 1870 to 1899) used for the preindustrial simulations. (Table 1). The GLOMAP reconstruction has the advantage of providing a monthly climatology of LGM SST and sea ice extent, while only annual mean SST is available from the reconstruction by Tierney et al. (2020), without sea ice map distribution. So, Tierney et al. LGM SST for ECHAM6-wiso was produced by taking the annual mean SST anomaly from Tierney et al. (2020) and adopting the monthly climatology temperature variability from GLOMAP. WeSince we also used theGLOMAP sea ice extent data from GLOMAP in this case, too.the SST was adjusted slightly to maintain consistency (e.g., SST set to freezing temperature where there is sea ice). In order to investigate the impact of sea ice extent on our isotope results and the related isotope-temperature slopes for LGM-to-PI climate change, we used LGM SST outputs from MIROC 4m simulations combined with sea ice extent data from the same MIROC 4m simulations or from GLOMAP dataset. Similarly for PI conditionsTo evaluate the LGM-PI anomalies, we performed several PI simulations with different sea ice boundary conditions depending on the setup of LGM experiments, using climatological monthly mean sea ice area fractions

from AMIP or MIROC 4m coupled simulations~~.~~ (Table 1). The prescribed LGM ice sheet is GLAC-1D (Tarasov and Peltier, 2002; Tarasov et al., 2012, 2014; Abe-Ouchi et al., 2013; Briggs et al., 2014) for all LGM simulations. As with SST and sea ice distribution, mean $\delta^{18}O$ of surface seawater needs to be prescribed. For the PI simulations, we used the $\delta^{18}O$ reconstruction from the global gridded data set of LeGrande and Schmidt (2006). As no equivalent data set of the $\delta^2H$ composition of seawater exists, the deuterium isotopic composition of the seawater in any grid cell has been set equal to the related $\delta^{18}O$ composition, multiplied by a factor of 8, in accordance with the observed relation for meteoric water on a global scale (Craig, 1961). As in Werner et al. (2018), a prescribed glacial seawater enrichment of +1 ‰ and +8‰ is assumed for $\delta^{18}O$ and $\delta^2H$ in the LGM simulations, respectively. ~~Finally, the greenhouse gas and orbital conditions were prescribed according to PMIP4 protocol.~~
The PI and LGM simulations were run for 60 and 120 model years, respectively, and we used the last 30 model years for our analyses. The simulations' characteristics are summarized in Table 1. Two additional sensitivity simulations have been performed to evaluate the impacts of lower MIROC 4m sea ice area fraction in coastal grid cells (section 2.2.2) and the consideration of the isotopic composition of snow on sea ice in ECHAM6-wiso (section 2.1) on the modeled $\delta^{18}O_P$-temperature temporal slopes between LGM and PI (see ~~text in~~ Supplementary ~~Material~~Text S1). Also, a LGM simulation using the PMIP3 ice sheet reconstruction instead of GLAC-1D (see Figures 3b and 3d of Werner et al. (2018), respectively) has been performed to evaluate the impact of ice sheet topography on the isotopically enriched bias in Antarctica (~~see text in~~ Supplementary ~~Material~~Text S1).

**Table 1: Characteristics of the ECHAM6-wiso simulations in the present study.**

| LGM simulation name | SST | Sea ice | PI control simulation characteristics | Comments |
|---|---|---|---|---|
| LGM_GLOMAP | GLOMAP | GLOMAP | Mean PI SST and sea ice from AMIP | ~~Lower~~Less global SST cooling |
| LGM_tierney2020 | Tierney et al., 2020 | GLOMAP | Mean PI SST and sea ice from AMIP | ~~Higher~~More global SST cooling |
| LGM_miroc4m_sst_glomap_sic | MIROC 4m | GLOMAP | Mean PI SST and sea ice from AMIP | AMOC oscillation: weak phase |
| LGM_miroc4m_sst_and_sic | MIROC 4m | MIROC 4m | Mean PI SST from AMIP and PI sea ice from MIROC 4m | AMOC oscillation: weak phase |
| LGM_miroc4m_strong_AMOC_sst_glomap_sic | MIROC 4m | GLOMAP | Mean PI SST and sea ice from AMIP | AMOC oscillation: strong phase |
| LGM_miroc4m_strong_AMOC_sst_and_sic | MIROC 4m | MIROC 4m | Mean PI SST from AMIP and PI sea ice from MIROC 4m | AMOC oscillation: strong phase |

## 2.4 Observational data

To evaluate the modeled $\delta^{18}O$ of precipitation and snow values at ice core stations, we use here a selection of 6 Greenland and ~~10~~11 Antarctic ice cores for the preindustrial and LGM climates (Figure 3). The observed $\delta^{18}O$ values were defined as averages over the last 200 years for the preindustrial period, and in the $21 \pm 1$ ka period for the LGM. We also use LGM-PI $\delta^{18}O$ anomalies from 5 (sub-)tropical ice cores that are reported in Table 2 of Risi et al. (2010). The ice core data used in this study are summarized in Table 2. Similarly, we use temperature reconstructions based on borehole reconstructions or $\delta^{15}N$ at 7

In order to mitigate the seasonal bias when comparing observed $\delta^{18}O$ from snow in ice cores with modeled $\delta^{18}O$ of precipitation or deposited snow, the modeled $\delta$ values are calculated as a precipitation (or snow)-weighted mean with respect to the V-SMOW scale. For the evaluation of modeled $\delta^{18}O$ of precipitation at a global spatial scale, we extracted 14 entities from the SISALv2 speleothem dataset (Comas-Bru et al., 2020) where both PI and LGM $\delta^{18}O$ values of calcite or aragonite are available. ~~As recommended by Comas-Bru et al. (2019), we defined here averaged PI and LGM values as the means of the 1850-1990 CE and 21 ± 1 ka periods, respectively. To compare the $\delta^{18}O$ of speleothem data with our modeled $\delta^{18}O$ of precipitation ($\delta^{18}O_p$), the measured $\delta^{18}O$ of calcite or aragonite are converted into $\delta^{18}O$ of drip-water using equations 1 or 2 of Comas-Bru et al. (2019), respectively, after conversion from V-PDB to VSMOW scale (equation 3 of Comas-Bru et al. (2019)). The annual mean surface air temperature from ECHAM6-wiso is used for the conversion. A seasonal bias can appear in the isotopic composition of drip water archived in speleothem records due to the re-evaporation of the precipitated water (Wackerbarth et al., 2010). An additional fractionation between the drip water and the formed calcite/aragonite can also be observed for many speleothems (Dreybrodt and Scholz, 2011~~ (see Supplementary Text S2 for the details).
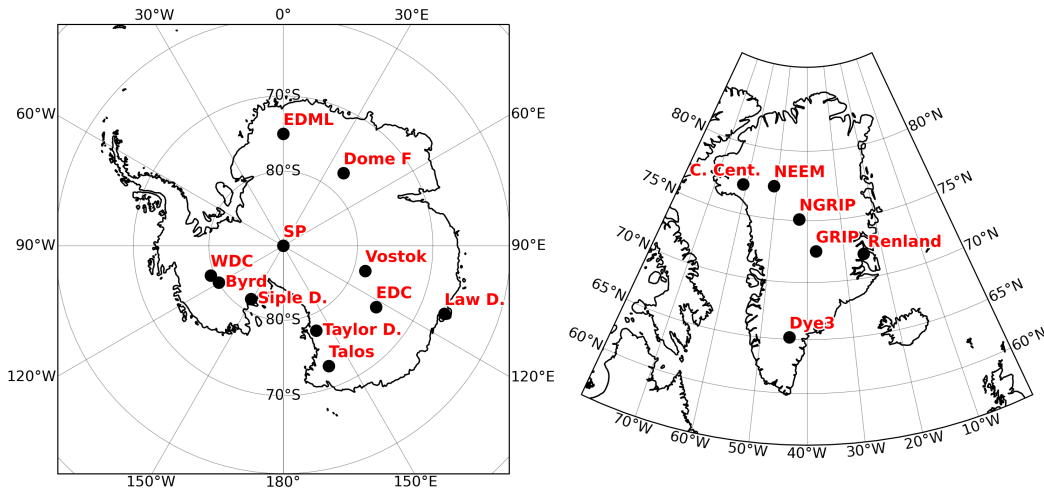


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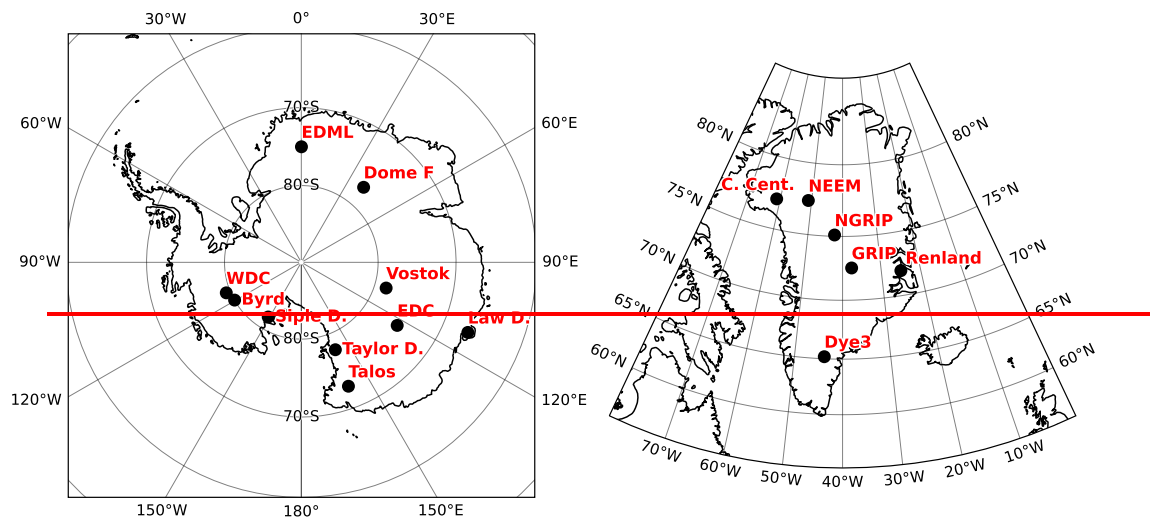
**Figure 3: Location of polar ice core sites in Antarctica (left) and Greenland (right).**

**Table 2: Selected ice cores records and their geographical coordinates, reported PI values of δ¹⁸O ~~and~~, changes in δ¹⁸O between LGM and PI and modeled range of Δδ¹⁸O among our simulations reported in Table 1.**

| Site | Longitude | Latitude | $\delta^{18}O_{PI}$ (‰) | ~~$A_{LGM-PI}$~~$\Delta\delta^{18}O$ (‰) | Modeled $\Delta\delta^{18}O$ range (‰) |
|---|---|---|---|---|---|
| South Pole[a] | 0 | -90 | -51.0 | -6.4 | -3.33 ; -0.65 |

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| | | | | | |
|---|---|---|---|---|---|
| ~~Vostok[a,b]~~Vostok[b,c] | 106.87 | -78.47 | -56.8 | -4.8 | -2.45 ; -0.08 |
| Dome ~~F[e]~~F[d] | 39.70 | -77.32 | -54.6 | -4.9 | -2.75 ; -0.37 |
| ~~EDC[d,e]~~EDC[e,f] | 123.35 | -75.10 | -50.4 | -5.6 | -2.30 ; 0.48 |
| ~~EDML[b,d]~~EDML[c,e] | 0.07 | -75.00 | -44.8 | -6.3 | -3.95 ; -0.91 |
| Law ~~Dome[b]~~Dome[c] | 112.83 | -66.73 | -22.4 | -5.5 | -5.13 ; -0.27 |
| Taylor ~~Dome[f]~~Dome[g] | 158.72 | -77.8 | -40.5 | -3.5 | -5.73 ; -2.87 |
| ~~Talos[g]~~Talos[h] | 159.18 | -72.82 | -36.1 | -5.4 | -2.28 ; -0.37 |
| ~~Byrd[h]~~Byrd[i] | -119.52 | -80.02 | -32.9 | -7.3 | -4.55 ; -1.40 |
| Siple ~~Dome[b]~~Dome[c] | -148.82 | -81.67 | -25.6 | -7.8 | -6.23 ; -3.23 |
| ~~WDC[b]~~WDC[c] | -112.14 | -79.46 | -34 | -7.3 | -5.03 ; -1.66 |
| ~~GRIP[a,j]~~GRIP[b,k] | -37.63 | 72.58 | -35.3 | -5.4 | -3.23 ; -0.43 |
| ~~NGRIP[a,k]~~NGRIP[b,l] | -42.32 | 75.10 | -35.2 | -7.4 | -6.37 ; -3.71 |
| ~~NEEM[l,m]~~NEEM[m,n] | -51.06 | 77.45 | -33 | -10 | -6.48 ; -5.17 |
| Camp | -61.13 | 77.17 | -29.3 | -12.9 | -9.40 ; -6.73 |
| ~~Century[i]~~Century[j] | | | | | |
| ~~Dye3[j]~~Dye3[k] | -43.81 | 65.18 | -27.7 | -7.3 | -5.58 ; -3.79 |
| ~~Renland[i]~~Renland[j] | -25.00 | 72.00 | -27.4 | -3.8 | -9.14 ; -5.72 |
| ~~Huascaran[a]~~Huascaran[b] | -77.61 | -9.11 | - | -6.3 | -3.33 ; -0.65 |
| ~~Sajama[a]~~Sajama[b] | -68.97 | -18.1 | - | -5.4 | -2.45 ; -0.08 |
| ~~Illimani[a]~~Illimani[b] | -67.77 | -16.62 | - | -6 | -2.75 ; -0.37 |
| ~~Guliya[a]~~Guliya[b] | 81.48 | 35.28 | - | -5.4 | -2.30 ; 0.48 |
| ~~Dunde[a]~~Dunde[b] | 96 | 38 | - | -2 | -3.95 ; -0.91 |

References: [a]-reported in [a] Steig et al. (2021), [b] Risi et al. (2010), [~~b~~c] WAIS Divide project members (2013), [~~d~~d] Kawamura et al. (2007), [~~d~~e] Stenni et al. (2010), [~~e~~f] Landais et al. (2015), [~~f~~g] Steig et al. (2000), [~~g~~h] Stenni et al. (2011), [~~h~~i] Blunier and Brook (2001), [i ~~Vinther et al. (2009),~~ j] Vinther et al. (2009), [k Vinther et al. (2006), [~~k~~l] North Greenland Ice Core project members (2004), [~~l~~m] Guillevic et al. (2013), [~~m~~n] Schüpbach et al. (2018).

# 3 Results of the LGM-PI ECHAM6-wiso simulations

## 3.1 Evaluation of ECHAM6-wiso under LGM conditions

We evaluate here the global distribution of δ$^{18}$O of precipitation ($\delta^{18}O_p$) changes between LGM and PI ~~($A_{LGM-PI}$~~($\Delta\delta^{18}O_p$) from our different ECHAM6-wiso simulations. Figure 4 shows the comparison of modeled $\delta^{18}O_p$ anomalies with isotope measurements from ice cores and speleothems for the simulation LGM_miroc4m_sst_and_sic (i.e., SST and sea ice boundary conditions from MIROC 4m simulation at weak AMOC phase). Well-known patterns of global ~~$A_{LGM-PI}$~~$\Delta\delta^{18}O_p$ distribution are found in ECHAM6-wiso, like the negative anomalies across Canada, Greenland and Northern Europe due to the presence of glaciers in these areas during LGM period (Figure 4c). Generally, negative $\delta^{18}O_p$ anomalies are also simulated over Antarctica and the Southern Ocean ~~due to lower temperatures in~~, where the LGM cooling is stronger compared to ~~PI~~

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240  ~~period~~<u>lower southern latitudes</u> (Figure 4a). In the mid-to-low latitudes, ~~$\text{A}_{\text{LGM-PI}}\delta$~~$\underline{\Delta\delta}^{18}\text{O}_p$ is mainly controlled by precipitation anomalies (Figure 4b). For example, lower modeled precipitation in the Amazonian area, over parts of South East Asia and in the western Pacific Ocean during the LGM leads to positive modeled $\delta^{18}\text{O}_p$ anomalies. Despite some biases in modeled ~~$\text{A}_{\text{LGM-PI}}\delta$~~$\underline{\Delta\delta}^{18}\text{O}_p$, like in Southern Amazonia (Figures 4c and d) where negative anomalies are measured in ice cores (between -2 and -6‰, see green dots in Figure 4d) while positive anomalies are simulated (between 0 and 4‰), modeled $\delta^{18}\text{O}_p$ anomalies are

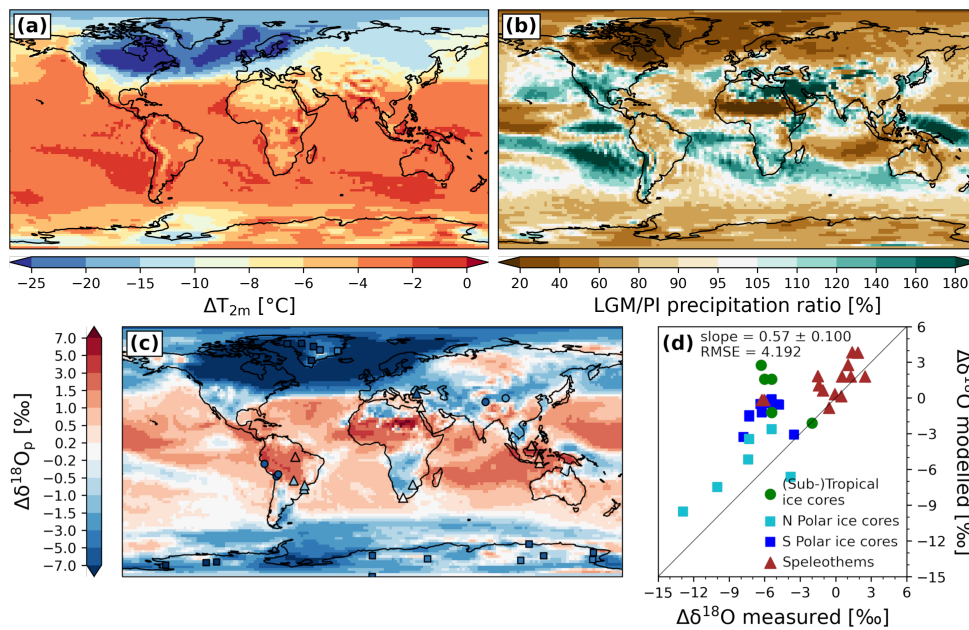245  in rather good agreement with observations from ice cores and speleothems (Figure 4d).



11
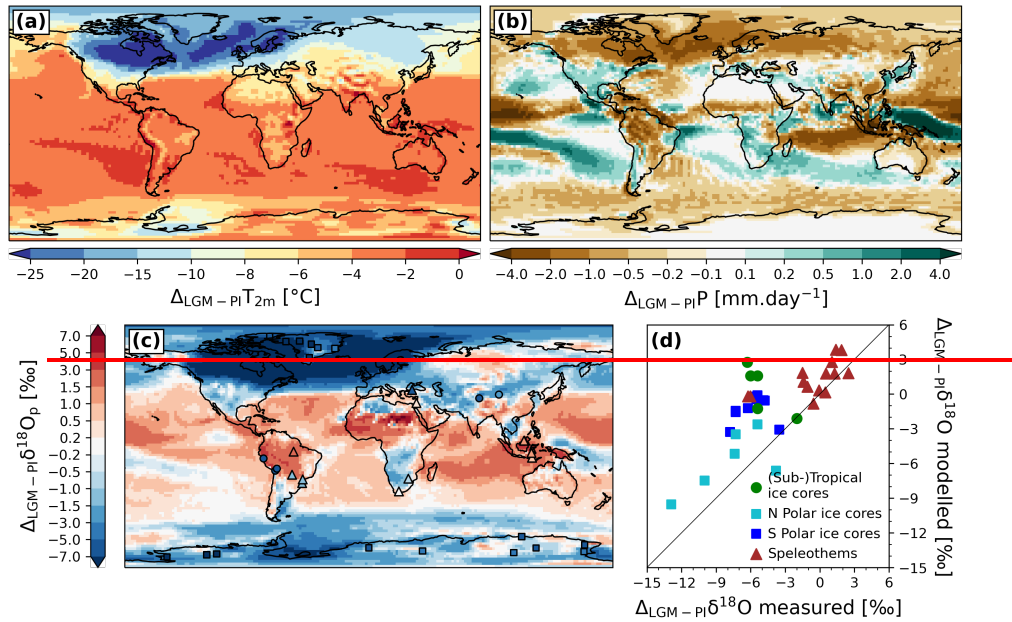
**Figure 4:** ~~Changes in modeled~~ (a) **LGM-PI** 2m air temperature ~~changes~~, (b) **LGM/PI** precipitation **ratio** and (c) **LGM-PI** $\delta^{18}O_p$ ~~between the LGM and PI climates~~**anomalies** from the LGM_miroc4m_sst_and_sic simulation (background colors). In (c), the squares, dots and triangles represent $\delta^{18}O$ changes measured in polar ice cores, (sub-)tropical ice cores and speleothems, respectively. Measured $\delta^{18}O$ in calcite or aragonite from speleothems have been converted into $\delta^{18}O$ of drip-water before comparison with modeled $\delta^{18}O_p$ (see Section 2.4). (d) Scatter plot showing a comparison of observed $\delta^{18}O$ changes with modeled $\delta^{18}O_p$ anomalies at the nearest grid cell of the archives' locations. Northern and southern polar ice core locations are distinguished by cyan and blue colors, respectively.

The isotope ~~distribution~~fractionation is mainly controlled by changes in temperature and in the water cycle. Even though all the ECHAM6-wiso simulations show similar global distribution of 2m air temperature ($T_{2m}$) and precipitation responses to the various SST and sea ice boundary fields, we find some differences too (Figures ~~S1~~S2 and ~~S2~~S3 in Supplementary Material). As expected, the modeled global cooling using SST from GLOMAP is lower while it is stronger when using SST from Tierney et al. (2020) (cooling of -4 and -6.3°C, respectively). Average $T_{2m}$ anomalies in the middle range are obtained when using the MIROC 4m SST fields (between -4.4 and -5.3°C depending on the MIROC 4m data used). The temperature over sea ice covered areas are largely impacted by the sea ice forcings used (i.e., GLOMAP or MIROC 4m). The modeled $T_{2m}$ anomalies over the Southern Ocean vary between -10 and -15°C with ~~GLOMAP~~an extensive LGM sea ice (GLOMAP) while the cooling is only between -4 and -10°C when ECHAM6-wiso is forced by a less extensive one (MIROC 4m sea ice~~. The opposite is true for~~). For the Arctic region~~. A~~, a strong cooling is simulated with the very extensive sea ice from MIROC 4m with a weak AMOC phase (a cooling of more than 20°C), more than with the sea ice from GLOMAP (between -20 and -10°C). The different SST boundary conditions have a strong influence on the precipitation anomalies, especially at mid-to-low latitudes including the western Pacific area and the East Asian monsoon region (Figure ~~S2~~S3). All these differences in $T_{2m}$ and precipitation responses have profound impacts on modeled $\delta^{18}O_p$ anomalies (Figure ~~S3~~S4) and their agreements with observations (Figures 4c and d).

Table 3: Values of $\Delta_{\text{LGM-PI}}\delta^{18}O$ ~~$\delta^{18}O$~~ model-data slope (1 is better), coefficient of determination $r^2$ and root mean square error (RMSE) for our ECHAM6-wiso simulations using different SST and sea ice boundary fields. For each column, worst to best model-data agreements are shown with a yellow-to-green colormap.

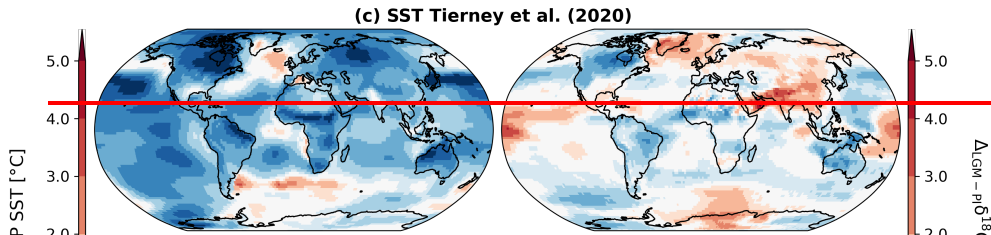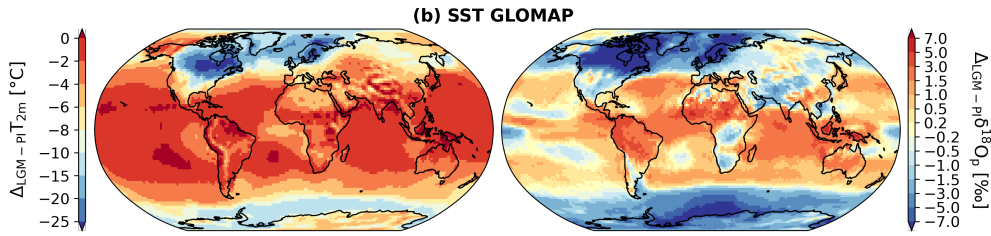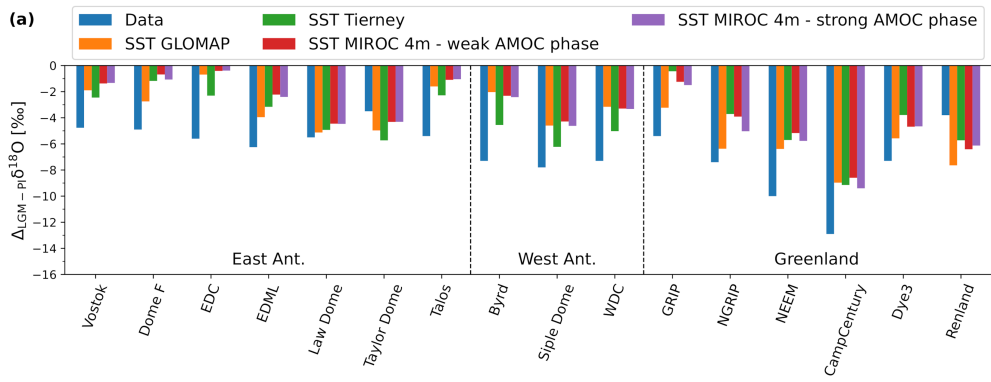| LGM simulation name | Slope | $r^2$ | RMSE (‰) |
|---|---|---|---|
| LGM_GLOMAP | 0.699 | 0.~~540~~543 | 3.~~943~~938 |
| LGM_tierney2020 | 0.646 | 0.~~584~~588 | 3.~~507~~500 |
| LGM_miroc4m_sst_glomap_sic | 0.~~592~~589 | 0.~~562~~561 | 3.~~907~~930 |
| LGM_miroc4m_sst_and_sic | 0.~~582~~574 | 0.~~498~~490 | 4.~~135~~192 |
| LGM_miroc4m_strong_AMOC_sst_glomap_sic | 0.~~660~~656 | 0.~~580~~579 | 3.~~910~~931 |
| LGM_miroc4m_strong_AMOC_sst_and_sic | 0.~~558~~551 | 0.~~532~~527 | 4.~~299~~348 |

The statistics of ~~$\Delta_{\text{LGM-PI}}\delta$~~$\Delta\delta^{18}O_p$ model-data agreements are shown for our different ECHAM6-wiso simulations in Table 3. The best model-data agreement in terms of model-data slope (1 is perfect match) is found when using SST and sea ice from GLOMAP (slope = 0.70) as boundary conditions for ECHAM6-wiso, but better coefficient of determination ($r^2$) and root mean square error (RMSE) are obtained with LGM SST from Tierney et al. (2020) ($r^2 = 0.58$ and RMSE = 3.5 ‰). We notice a worse model-data agreements in $\delta^{18}O_p$ changes when both SST and sea ice changes from MIROC 4m simulations are provided as sea surface boundary conditions (slopes lower than 0.582 and RMSE higher than 4.1 ‰). This is in agreement with Werner et al. (2018) who showed a worse model-data agreement when using SST and sea ice boundary conditions from a coupled model instead of reconstructed ones. The substitution of MIROC 4m sea ice changes by GLOMAP ones, _which are more extensive in the south and less extensive in the north,_ improves the ~~$\Delta_{\text{LGM-PI}}\delta$~~$\Delta\delta^{18}O_p$ model-data agreement for all cases (i.e., weak or strong AMOC phase). For example, the model-data slope when using SST changes from MIROC 4m simulation during strong AMOC phase is similar to the one for the simulation with the Tierney et al. SST reconstruction (0.66 and 0.65, respectively).

## 3.2 Impacts of SST boundary conditions on the ~~$\Delta_{\text{LGM-PI}}\delta$~~$\Delta\delta^{18}O$ model-data agreement at polar ice core stations

The modeled values of ~~$\Delta_{\text{LGM-PI}}\delta$~~$\Delta\delta^{18}O$ of snow ($\delta^{18}O_{sn}$) _and $\Delta T_{2m}$_ at polar ice cores stations for different boundary conditions in LGM-PI SST changes are compared to isotopic observations _and temperature reconstructions_ in Figure 5a. Only simulations using sea ice from GLOMAP are selected here. Except for Renland station in the North and Taylor Dome in the South (that are both coastal sites), ECHAM6-wiso generally under-estimates $\delta^{18}O_{sn}$ changes. ~~The main explanation for this general bias is the use of GLAC-1D ice sheet reconstruction for our ECHAM6-wiso simulations.~~ _We find a generally good agreement with the reconstructed temperature data, except for GRIP station where the cooling is 6-10°C too weak (dots and crosses in Figure 5a)._ The substitution of GLAC-1D reconstruction by the PMIP3 one strongly improves the model-data agreement of $\delta^{18}O$ in Antarctica (Figure ~~S4~~S5), leading to a better model-data agreement at global scale (slope = 0.87, $r^2 = 0.62$ and RMSE = 3.2 ‰) compared to the LGM_GLOMAP experiment. This agrees with the findings of Werner et al. (2018) who showed that the

isotopic model-data correlation for Antarctic ice core stations is weaker when using GLAC-1D instead of PMIP3 ice sheet reconstruction (RMSE = 2.1 and 1.1 ‰ for 11 Antarctic stations, respectively). However, this better $\Delta\delta^{18}O$ model-data agreement is more likely due to a bias compensation than a more realistic ice sheet because the simulation of Antarctic temperatures by ECHAM6-wiso is degraded at the same time (markers in Figure S5). Except for the Taylor Dome station, all

300 modeled ~~$A_{LGM-PI}\delta$~~$\Delta\delta^{18}O_{sn}$ at polar ice core stations are in better agreement with measurements (blue bars in Figure 5a) when reconstructed SST fields ~~from~~(i.e., GLOMAP or Tierney et al~~.~~.) are used (orange and green bars in Figure 5a, respectively), confirming the results of Werner et al. (2018) about the worse model-data agreement when using sea surface boundary conditions from a coupled model instead of reconstructed ones. The change from one MIROC 4m SST field to another one (i.e., weak or strong AMOC phase) as input for ECHAM6-wiso does not modify the modeled ~~$A_{LGM-PI}\delta$~~$\Delta\delta^{18}O_{sn}$ values much

305 (red and purple bars in Figure 5a).

~~As expected, the modeled~~ For a stronger SST cooling ~~is globally lower when using SST from GLOMAP to drive ECHAM6-wiso (left maps in subplots (e) to (e) of Figure 5). However, a strong cooling is obtained with GLOMAP SST~~ in the Southern Ocean~~, which is the evaporative source of isotopic signals measured in polar areas. As a consequence, temperature changes in Antarctica are stronger when using SST from GLOMAP or~~ (GLOMAP and Tierney et al~~., giving~~.), ECHAM6-wiso simulates

310 higher ~~modeled~~ $\delta^{18}O_p$ changes ~~compared to modeled results using SST fields from MIROC 4m~~ (right maps of Figure 5~~), and~~) that are in better agreement with the observations. ~~The~~A stronger cooling in the Atlantic sector of the Southern Ocean ~~with~~ (GLOMAP SST compared to ~~the~~ Tierney et al. one) has the consequence of enhancing the $\delta^{18}O_p$ depletion in the Atlantic-Indian Ocean sector of Antarctica (right map of Figure 5c) despite similar temperatures ~~between the two simulations~~ (left map of Figure 5c). This area includes the ice core stations Dome Fuji and EDML, ~~and a~~where better $\delta^{18}O$ and similar temperature

315 model-data ~~agreement is~~ agreements are found ~~there when GLOMAP SST values are provided as boundary conditions (~~(orange markers and bars in Figure 5a). ~~The opposite~~This is also true for other stations further to the east and west, like WDC and EDC~~. As in the Southern Ocean, a higher cooling is simulated~~ (green bars and markers in Figure 5a). When using a SST field with stronger cooling in the Northern North Atlantic Ocean (i.e., GLOMAP), a stronger cooling is simulated there to the south of Greenland ~~if the SST from GLOMAP is used~~. A stronger cooling is simulated in the southern and central ~~part~~parts of inner

320 Greenland~~,~~ too (left maps of Figure 5~~). As a consequence,~~) with consequently higher $\delta^{18}O$ changes between LGM and PI ~~are simulated~~ in Greenland ~~with the SST from GLOMAP~~ (right maps of Figure 5~~),~~) except in Northern ~~Greenland~~part like at Camp Century station (Figure 5a). A better agreement is obtained with the Greenland $\delta^{18}O$ observations under this configuration (orange bars in Figure 5a), ~~except for Renland and Camp Century (worse and similar model-data agreements among the simulation results, respectively~~without improving the overly too weak cooling in the model (colored markers for GRIP station
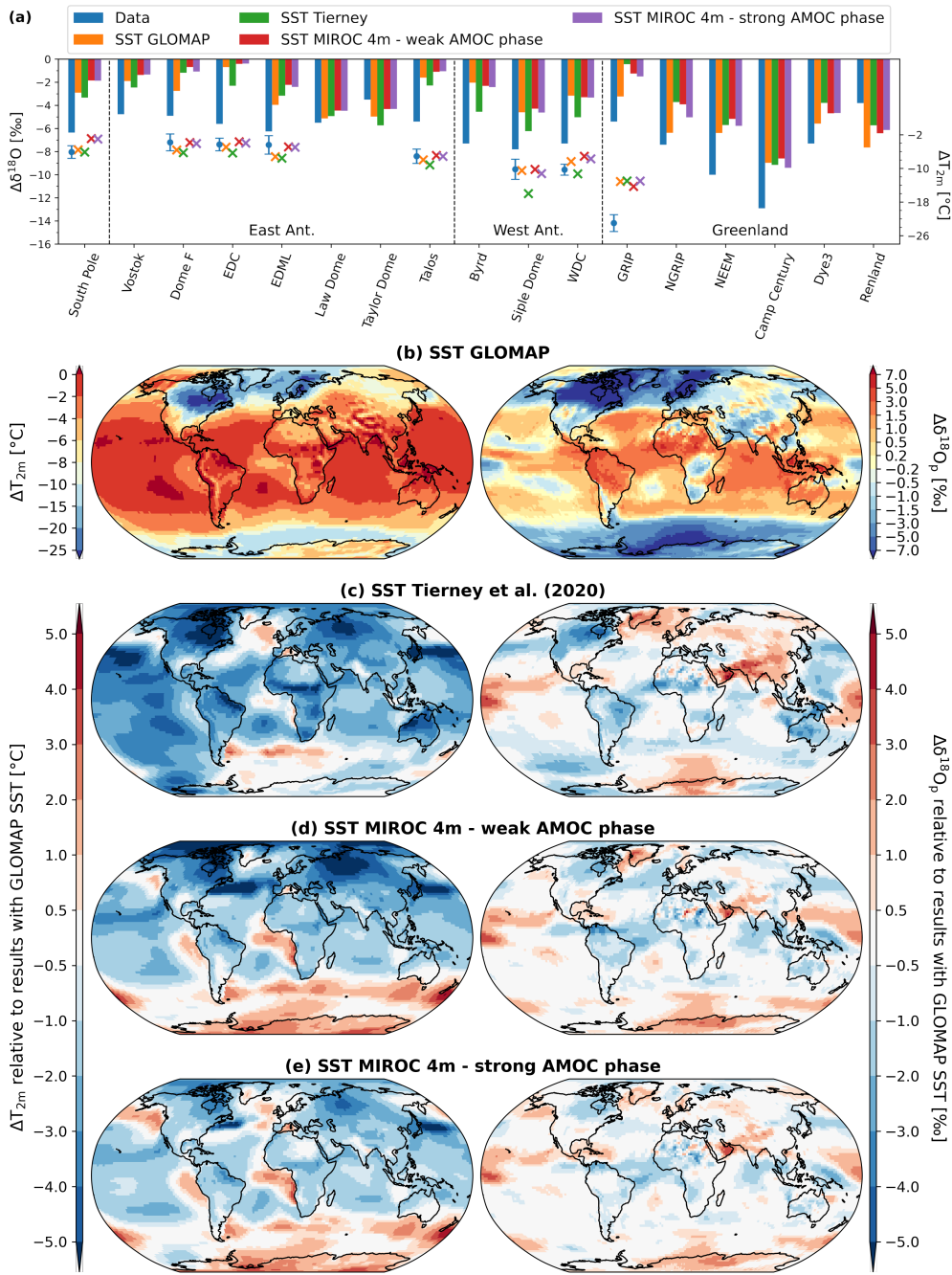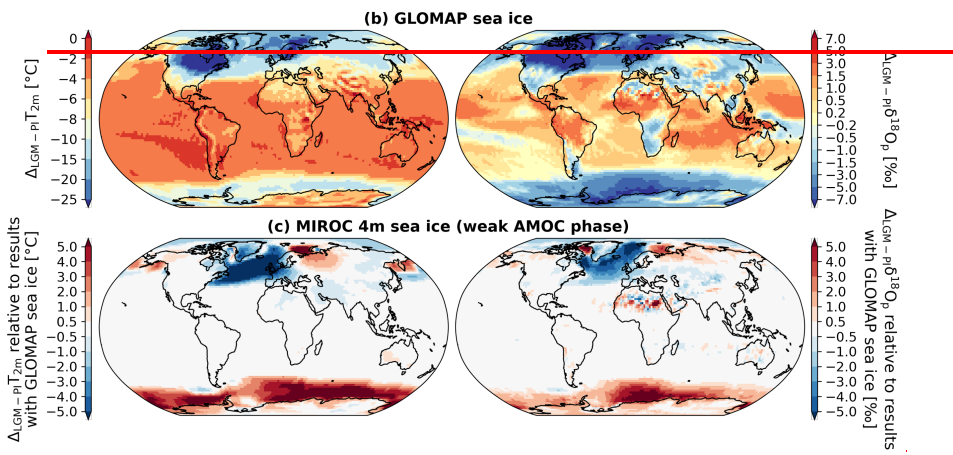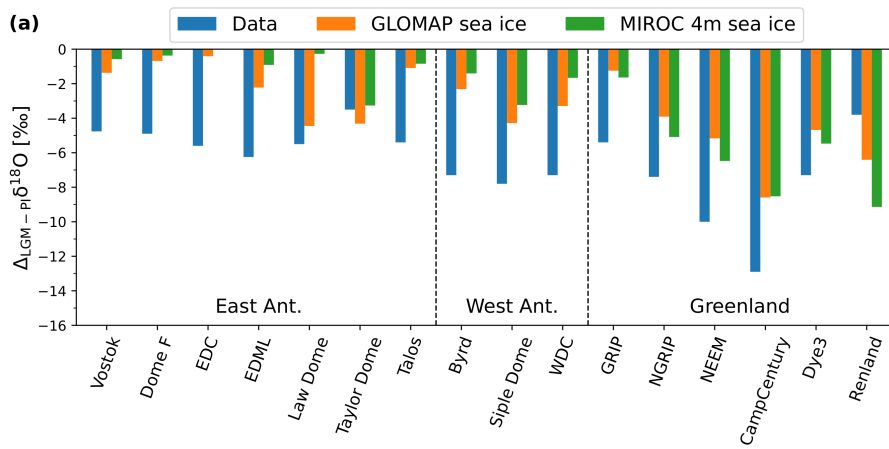
325 in Figure 5a).

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**(a)**

**(b) SST GLOMAP**

**(c) SST Tierney et al. (2020)**

**(d) SST MIROC 4m - weak AMOC phase**
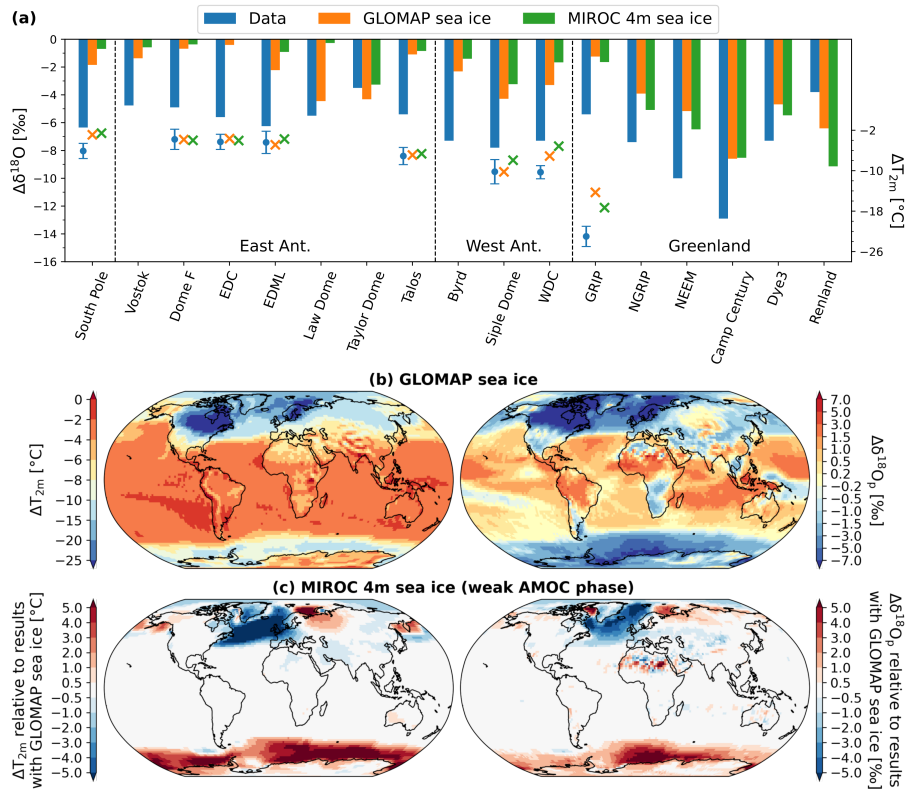
**(e) SST MIROC 4m - strong AMOC phase**

15
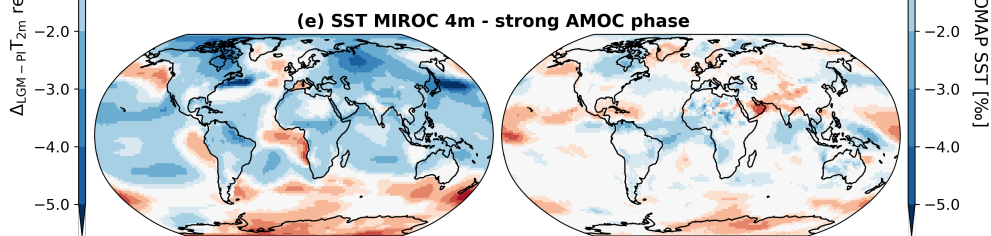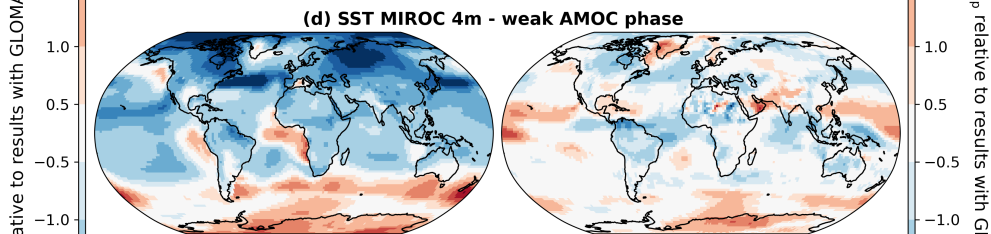
**Figure 5: (a) Comparison of modeled anomalies in T₂ₘ (crosses, right vertical axis) and** $\delta^{18}O_{sn}$ **(bars, left vertical axis) between LGM and PI with temperature and** $\delta^{18}O$ **anomalies ~~measured in~~from polar ice cores (blue dots and bars, respectively). All modeled results are from simulations with the same sea ice boundary conditions from GLOMAP but with different SST forcings: GLOMAP (orange), Tierney et al. (2020) (green), MIROC 4m with weak LGM AMOC phase (red) and MIROC 4m with strong LGM AMOC phase (purple). (b) Modeled T₂ₘ and** $\delta^{18}O_p$ **changes between LGM and PI using GLOMAP SST (left and right maps, respectively). Maps in plots (c) to (e) show the impacts on T₂ₘ and** $\delta^{18}O_p$ **anomalies using the other SST boundary conditions. The values are expressed relative to the modeled results from (b).**

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### 3.3 Impacts of sea ice ~~changes~~ boundary conditions on the ~~$A_{LGM-PI}$δ~~$\Delta\delta^{18}O$ model-data agreement at polar ice core stations

To analyze the effects of sea ice boundary conditions on the modeled $\delta^{18}O$ changes in polar regions between LGM and PI, we compare the results from the simulations using the same SST (here from the MIROC 4m simulation with the weak AMOC phase) but different sea ice area fraction fields: GLOMAP and MIROC 4m (i.e., LGM_miroc4m_sst_glomap_sic and LGM_miroc4m_sst_and_sic simulations, respectively). For all Antarctic ice core stations, a stronger depletion in $\delta^{18}O_{sn}$ between LGM and PI is simulated ~~with GLOMAP sea ice distribution (~~ when the LGM sea ice in the Southern Ocean is more extensive (GLOMAP, orange bars in Figure 6a). Except for Taylor Dome, a better agreement with isotopic observations is then found. ~~The LGM sea ice from GLOMAP in the Southern Ocean is more extensive than the one from MIROC 4m (Figure 2).~~ as well as a better agreement with temperature reconstructions in West Antarctica. It has a huge impact on modeled $T_{2m}$ anomalies over the Southern Ocean (between 2 and 10°C), and the simulated cooling is higher by 1 to 4°C in Western Antarctica and in coastal regions of the continent (left map of Figure 6c). As a consequence, higher LGM-PI anomalies in $\delta^{18}O$ of precipitation and of snow are simulated: more than 5 ‰ over the Southern Ocean and around 1-2 ‰ on the continent, especially in the western part (right map of Figure 6c). ~~The situation is opposite to that of the Arctic Ocean and Greenland with the sea ice from MIROC 4m (weak AMOC phase) being~~ A more extensive ~~than the one from GLOMAP (Figure 2). Cooling is stronger~~ northern sea ice (i.e., MIROC 4m) increases the cooling by 5 to 10°C in the Arctic Ocean and from 0.5 to 5 °C in Greenland (left map of Figure 6~~c) when ECHAM6-wiso is forced by MIROC 4m sea ice boundary conditions,~~), giving higher $\delta^{18}O_p$ anomalies of up to 2 ‰ (right map of Figure 6~~c). The cooling is slightly lower near the Greenland coast because the LGM-PI sea ice change is more important~~) generally in ~~GLOMAP compared to MIROC 4m. This is due to the lower~~ better agreement with the temperature and $\delta^{18}O$ observations (green bars and markers in Figure 6a). The opposite is true with the less extensive sea ice distribution from MIROC 4m under a strong AMOC phase, weakening the model-data agreement for Greenland (Figure S6). Smaller sea ice area fraction in grid cells near coastal areas in MIROC 4m ~~(~~ compared to GLOMAP (95-98% against 100~~% in GLOMAP, see~~%, respectively. See section 2.2.2~~).~~), resulting in less important LGM-PI sea ice change, makes the cooling near the Greenland coast slightly lower. This lower sea ice change in MIROC 4m combined with the isotopic content of snow on sea ice taken into account for sublimation processes in sea ice covered regions leads to a reduction of the LGM-PI $\delta^{18}O_p$ changes in Baffin Bay (right map of Figure 6c). This aspect is investigated in detail in section 4.2. ~~Finally, if the less extensive sea ice distribution from MIROC 4m under a strong AMOC phase is used to force ECHAM6-wiso, modeled $\delta^{18}O_{sn}$ changes at Greenland ice core locations become smaller than the ones with GLOMAP sea ice, weakening the model-data agreement for this region (Figure S5).~~

**(a)**

Data · GLOMAP sea ice · MIROC 4m sea ice

$\Delta_{\mathrm{LGM-PI}}\delta^{18}\mathrm{O}$ [‰]

East Ant. · West Ant. · Greenland

Vostok · Dome F · EDC · EDML · Law Dome · Taylor Dome · Talos · Byrd · Siple Dome · WDC · GRIP · NGRIP · NEEM · CampCentury · Dye3 · Renland

**(b) GLOMAP sea ice**

$\Delta_{\mathrm{LGM-PI}}T_{\mathrm{2m}}$ [°C]

$\Delta_{\mathrm{LGM-PI}}\delta^{18}\mathrm{O}_{\mathrm{p}}$ [‰]

**(c) MIROC 4m sea ice (weak AMOC phase)**

$\Delta_{\mathrm{LGM-PI}}T_{\mathrm{2m}}$ relative to results with GLOMAP sea ice [°C]

$\Delta_{\mathrm{LGM-PI}}\delta^{18}\mathrm{O}_{\mathrm{p}}$ relative to results with GLOMAP sea ice [‰]

**Figure 6: (a) Comparison of modeled anomalies in** ~~T_{2m} (crosses, right vertical axis) and~~ $\delta^{18}O_{sn}$ **(bars, left vertical axis) between LGM and PI with** **temperature and** $\delta^{18}O$ **anomalies** ~~measured in~~**from** **polar ice cores (blue** **dots and** **bars**, **respectively). Modeled results are from the simulations using the SST changes of MIROC 4m with weak LGM AMOC but different sea ice boundary conditions (GLOMAP and MIROC 4m with weak LGM AMOC phase in orange and green, respectively). (b) Modeled T_{2m} and** $\delta^{18}O_p$ **changes between LGM and PI using GLOMAP sea ice (left and right maps, respectively). Maps in (c) show the impacts on T_{2m} and** $\delta^{18}O_p$ **anomalies using sea ice from MIROC 4m instead. Values are expressed relative to the modeled results from (b).**

### 3.4 Impacts of LGM AMOC strength on the ~~A_{LGM-PI}~~$\delta\Delta\delta^{18}O$ model-data agreement at polar ice core stations

Here, we investigate the impacts of AMOC strength on the modeled ~~A_{LGM-PI}~~$\delta\Delta\delta^{18}O$ in polar regions. For that, sea surface outputs (i.e., both SST and sea ice spatial distribution) from the MIROC 4m simulation with different LGM AMOC strengths are used as boundary conditions for ECHAM6-wiso. We focus first on the North Pole region because the AMOC strength mainly influences the climate of the Northern Hemisphere, as shown in SST and sea ice distributions used in this study (Figures 1 and 2). A weaker AMOC during LGM involves less heat transported in the north and thus lower LGM temperatures (i.e., larger cooling relative to PI), as shown in the left map of Figure 7c. A difference in T_{2m} of up to 10°C in the North Atlantic and Arctic Oceans is seen in the LGM_miroc4m_strong_AMOC_sst_glomap_sic and LGM_miroc4m_strong_AMOC_sst_and_sic simulations. Cooling in Greenland is reduced by 2~~-4~~ to 6 °C when the AMOC is increased~~.~~, increasing the warm bias in GRIP (markers in Figure 7a). LGM to PI changes in $\delta^{18}O$ in Greenland is mainly controlled by this change in mean temperature with an increase in LGM $\delta^{18}O_{sn}$ of between 1.2 and 2.5 ‰ at Greenland ice

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core stations for a stronger LGM AMOC (orange and green bar in Figure 7a). As ECHAM6-wiso generally underestimates the LGM-PI $\delta^{18}$O changes at the poles, a weaker AMOC generally improves the model-data agreement (blue and orange bars in Figure 7a). In the Southern Ocean and Antarctic regions, only small $T_{2m}$ changes are simulated by ECHAM6-wiso due to a change in AMOC strength during LGM (left map of Figure 7c~~).~~), as shown by the absence of any effect on model-data temperature agreement (markers in Figure 7a). As a consequence, modeled ~~$A_{LGM\text{-}PI}\delta$~~$\Delta\delta^{18}O_{sn}$ values are very similar between the 2 simulations (orange and green bars in Figure 7a). These small differences are due to the selection of the strong AMOC phase period in the middle of the peak in MIROC 4m simulation (see section 2.2.1). The impact of the period selection for the strong AMOC phase (e.g., the start or the end of the interstadial) on surface temperature and $\delta^{18}$O in Antarctica will be investigated more in detail in a future study. Finally, the changes of SST values alone due to AMOC strength variations change by only less than 1 ‰ of the modeled ~~$A_{LGM\text{-}PI}\delta$~~$\Delta\delta^{18}O_{sn}$ (red and purple bars in Figure 5a). This shows that the LGM to PI changes in sea ice distribution, related to the AMOC strength variations, have a large impact on modeled $T_{2m}$ anomalies and consequently on the isotopic signals in the North Pole region.
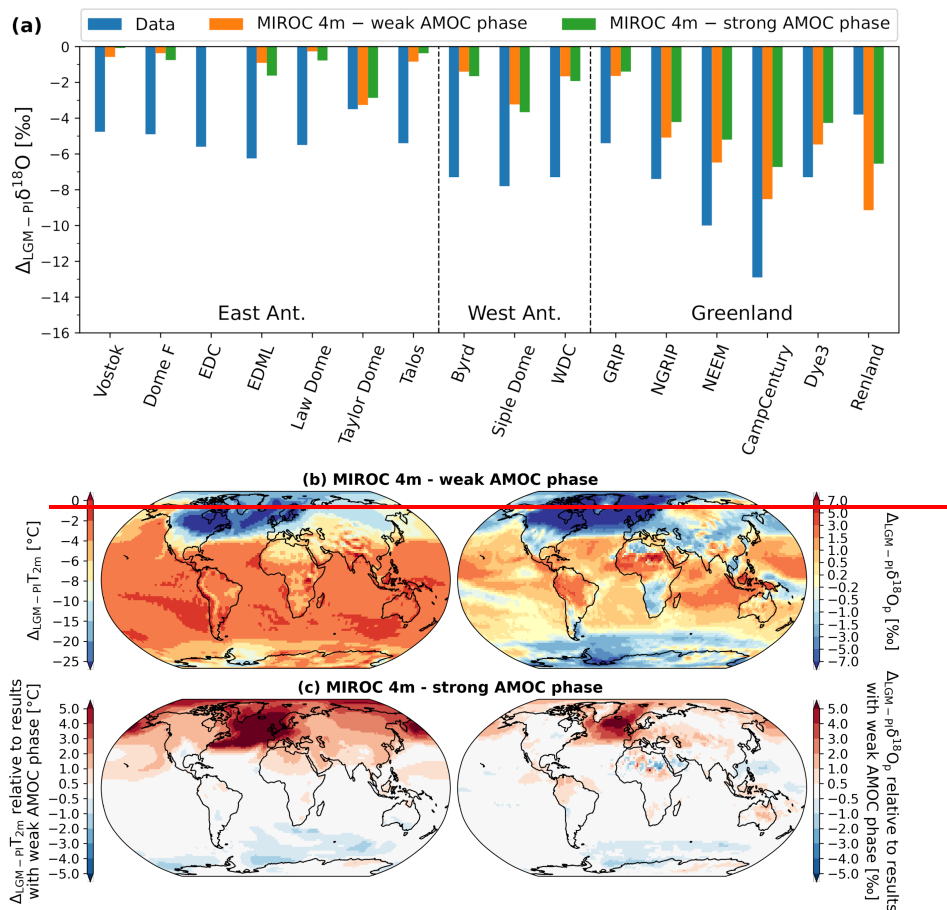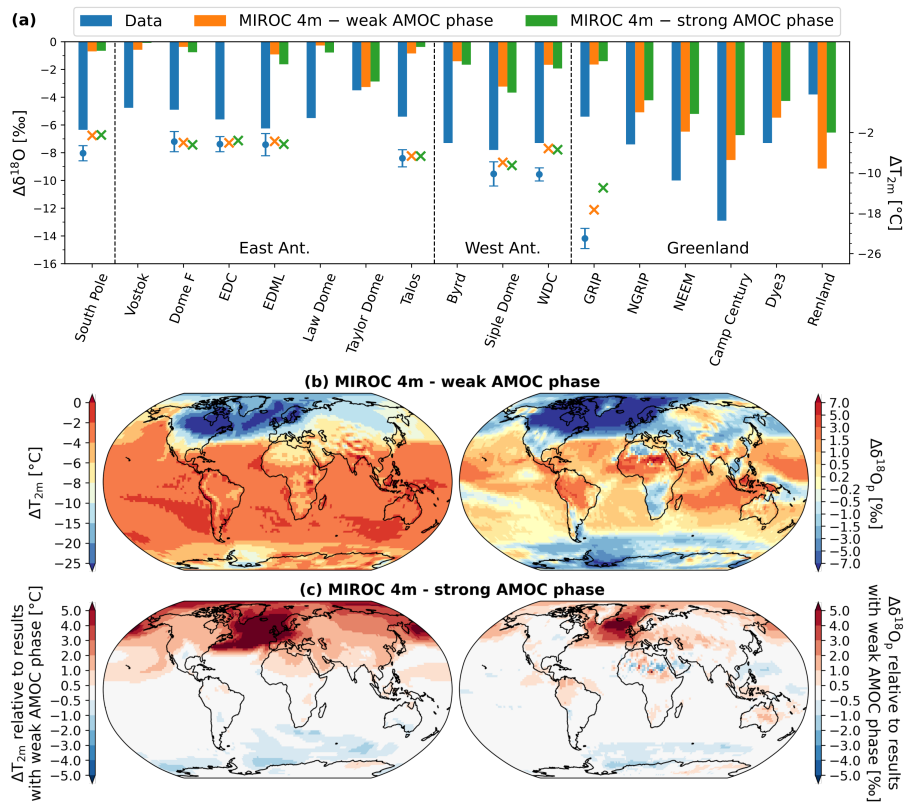
**Figure 7: (a)** Comparison of modeled anomalies in <u>$T_{2m}$ (crosses, right vertical axis) and</u> $\delta^{18}O_{sn}$ <u>(bars, left vertical axis)</u> between LGM and PI with <u>temperature and</u> $\delta^{18}O$ anomalies ~~measured in~~<u>from</u> polar ice cores (blue <u>dots and</u> bars<u>, respectively</u>). Modeled results are from simulations using the sea surface boundary conditions from the MIROC 4m coupled simulations: LGM_miroc4m_sst_and_sic and LGM_miroc4m_strong_AMOC_sst_and_sic in orange and green, respectively. **(b)** Modeled $T_{2m}$ and $\delta^{18}O_p$ changes between LGM and PI using MIROC 4m (weak AMOC phase) sea surface boundary conditions (left and right maps, respectively). Maps in **(c)** show the impacts on $T_{2m}$ and $\delta^{18}O_p$ anomalies using sea surface boundary conditions from MIROC 4m at strong LGM AMOC phase. Values are expressed relative to the modeled results from **(b)**.

**4 Impacts of sea surface boundary conditions on $\delta^{18}O$–$T_{2m}$ temporal slope for LGM-PI climate change**

We have analyzed the effects of LGM to PI changes in SST and sea ice distribution on modeled $\Delta_{\text{~~LGM-PI~~}}T_{2m}$ and $\Delta\delta^{18}O$ of precipitation and snow in the polar regions, as well as the impacts of the LGM AMOC strength. Next, we investigate the repercussions on modeled $\delta^{18}O_p$–$T_{2m}$ temporal slopes~~,~~ <u>and the underlying causes in terms of changes in moisture transport.</u> In other words, are $T_{2m}$ and $\delta^{18}O$ signals in the polar regions influenced ~~in the same way~~ by LGM to PI changes in SST and sea ice distribution <u>in the same way? What are the underlying dynamics, for example, in terms of changes in concentrations and transport of water vapor</u>? A correction for the prescribed glacial seawater change of 1 ‰ has been applied to LGM $\delta^{18}O$ values before temporal slope calculation, according to equation 1 of Stenni et al. (2010). ~~$\delta^{18}O_p$ values in the polar regions might be biased by strong changes in the seasonality or intermittency of the precipitation rate (Sime et al., 2009; Kino et al., 2021). To~~

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take into account this effect, modeled $T_{2m}$ values were weighted by the modeled monthly mean precipitation rates for the calculation of $\delta^{18}O_P$-$T_{2m}$ slopes (see Cauquoin et al., 2019b). As in Cauquoin et al. As in Cauquoin et al. (2019b), the calculation

415 of temporal slopes was restricted to grid cells where simulated annual mean temperatures are below +20°C for both PI and LGM. Moreover, we selected only the grid cells showing an absolute LGM-PI annual mean $T_{2m}$ difference of at least of 0.5°C. As a comparative point, PI spatial $\delta^{18}O_P$-$T_{2m}$ slopes of 0.72 and 0.94 ‰ °C$^{-1}$ are modeled by ECHAM6-wiso in East and West Antarctic ice core stations, respectively (calculated by considering the 25 grid cells centered on each drill location, excluding the ocean grid points), consistent with the mean observed value of 0.8 ‰ °C$^{-1}$ (Masson-Delmotte et al., 2008) and previous

420 modeling studies (Schmidt et al., 2007, Werner et al., 2018, Cauquoin et al., 2019b). For Greenland ice core stations, we find a modeled spatial slope of 0.71 ‰ °C$^{-1}$, also in agreement with previous model results (Schmidt et al., 2007, Cauquoin et al., 2019b).

## 4.1 Antarctica

The values of $\delta^{18}O_P$-$T_{2m}$ slope in East Antarctica are influenced in different ways by sea surface boundary conditions. LGM to

425 PI changes in sea ice area fractions have a strong impact on the slopes in coastal regions, as shown by the comparison between the plots (c)-(d) and (e)-(f) of Figure 8. Law Dome ice core is representative of this impact, with a slope of 0.29 and 0.62 ‰ °C$^{-1}$ depending if MIROC-4m (LGM_miroc4m_sst_and_sic, Figure 8d) or GLOMAP (LGM_miroc4m_sst_glomap_sic, Figure 8c) sea ice is used, respectively. The change of sea ice forcing has only a small effect on the temporal slopes modeled by ECHAM6-wiso in the East Antarctic plateau. The most sensitive case is EDC where the temporal slope is increased from

430 0.19 to 0.3 ‰ °C$^{-1}$ when switching from MIROC 4m sea ice with a strong AMOC to the GLOMAP one (Figures 8f and 8e, respectively). On average, the modeled temporal $\delta^{18}O_P$-$T_{2m}$ slopes of East Antarctic ice core stations are increased by more than 25% when MIROC 4m sea ice (red and brown markers in Figure 10) is replaced by the GLOMAP one (green and purple markers in Figure 10) due mainly to the coastal stations (i.e., Law Dome, Taylor Dome, Talos Dome and, to a lesser extent, EDML). The conclusions remain the same if instead of taking the averages of the slopes at ice cores stations, we use the

435 average slope across the entire East Antarctic area (Figure S6). The SST forcings have various impacts on the temporal slopes simulated by ECHAM6-wiso. The SST forcing from Tierney et al. (2020) enhances the LGM cooling in the eastern part of the Southern Ocean area compared to other SST forcings (left map of Figure 5c). It influences both the LGM $T_{2m}$ and $\delta^{18}O_P$ in the same direction (i.e., toward lower values) but with different magnitudes at EDC, Vostok and Talos Dome. Temporal slopes at these stations are increased by 0.17, 0.11 and 0.08 ‰ °C$^{-1}$, respectively, when ECHAM6-wiso is forced by SST from Tierney

440 et al. (2020) instead of the one from GLOMAP. The higher cooling in the Atlantic sector of the Southern Ocean when ECHAM6-wiso is forced by GLOMAP SST (Figure 8a) makes the Antarctic temporal slope values higher between 0 and 90°E of longitude compared to the other simulations. It impacts especially the Dome Fuji and EDML ice core sites, where values of temporal $\delta^{18}O_P$-$T_{2m}$ slopes reach 0.8 and 0.67 ‰ °C$^{-1}$ (i.e., an increase of at least 60 and 34 %, respectively, compared to the other simulations). As a consequence, the modeled $\delta^{18}O_P$-$T_{2m}$ slopes in East Antarctic ice core stations are on average higher

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$\Delta_{LGM-PI}\delta^{18}O_p/\Delta_{LGM-PI}T_{2m}$ [‰.°C⁻¹]

$\delta^{18}O$ in coastal and western low-elevated sites are derived from nearby local sources while the sources of high-elevation East
460 Antarctic ice cores are typically further north around 40-45° S (Sodemann and Stohl, 2009). So, the values of δ¹⁸Oₚ-T₂ₘ slope

in Antarctica are expected to be influenced by sea surface boundary conditions in different ways depending on the ice core locations and topography.
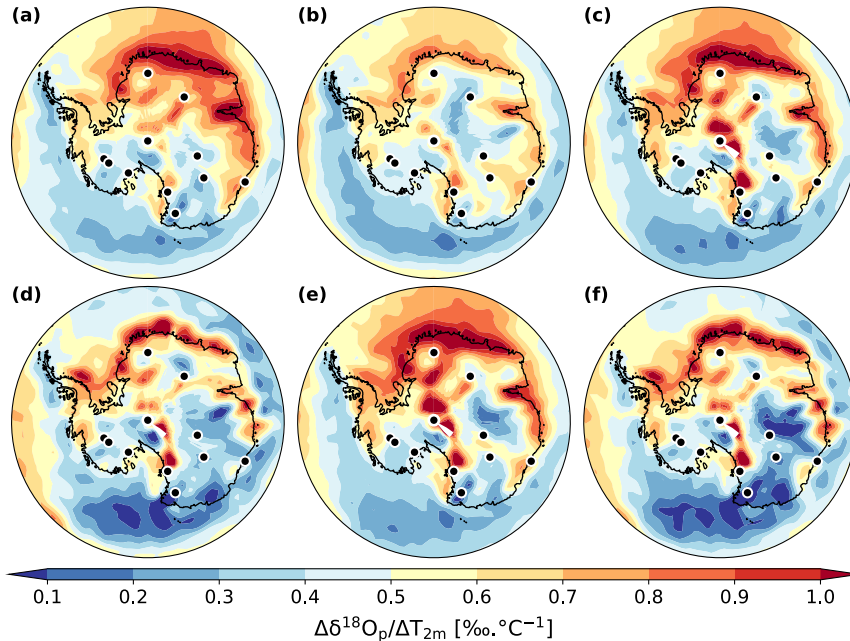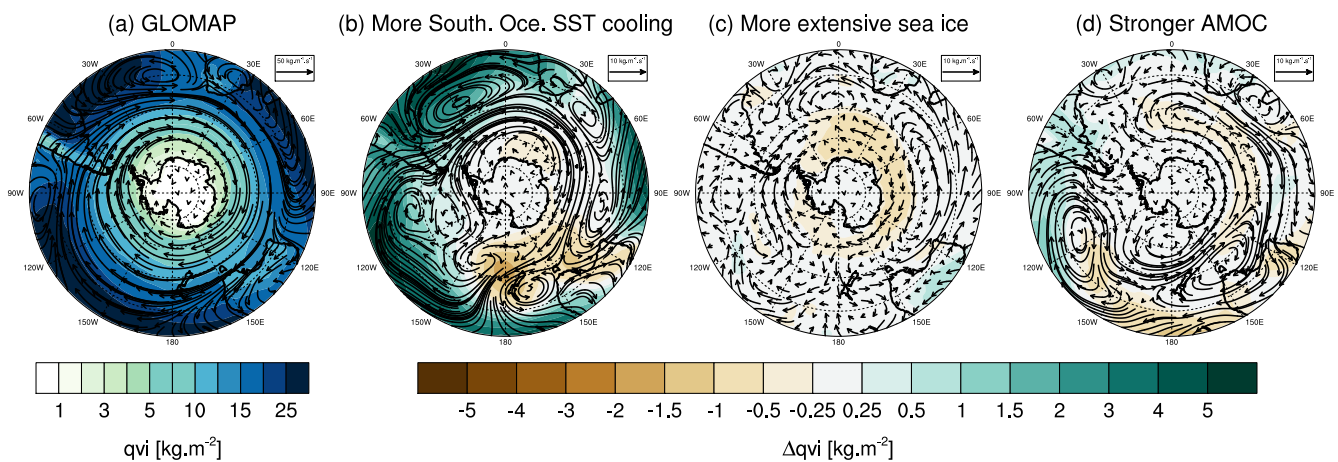


Figure 8: Spatial distribution of $\delta^{18}O_p$-$T_{2m}$ temporal slope in Antarctica for LGM-PI changes according to our different ECHAM6-wiso simulations: (a) LGM_GLOMAP, (b) LGM_tierney2020, (c) LGM_miroc4m_sst_glomap_sic, (d) LGM_miroc4m_sst_and_sic, (e) LGM_miroc4m_strong_AMOC_sst_glomap_sic, and (f) LGM_miroc4m_strong_AMOC_sst_and_sic. The dots indicate the location of the ice core stations.

Greater SST cooling in the Southern Ocean influences both the LGM $T_{2m}$ and $\delta^{18}O_p$ in the same direction (i.e., toward lower values) but with different magnitudes. Stronger cooling in the eastern part of the ocean (Tierney et al. LGM SST colder than the GLOMAP one) increases the temporal slopes at EDC and Talos Dome by 0.23 and 0.07 ‰ °C$^{-1}$, respectively. A higher cooling in the Atlantic sector of the Southern Ocean (GLOMAP SST in Figure 8a) makes the Antarctic temporal slope values higher between 0 and 90°E longitude compared to the other simulations. It impacts especially the Dome Fuji and EDML ice core sites, where values of temporal $\delta^{18}O_p$-$T_{2m}$ slopes reach 0.66 and 0.69 ‰ °C$^{-1}$, representing an increase of at least 25% compared to simulations using SST with less cooling. This can be explained by a change in moisture transport. With a stronger SST cooling in the Southern Ocean (GLOMAP against MIROC 4m), the westerlies around Antarctica are enhanced and the atmosphere is wetter in the mid-latitudes while a drier belt appears around the continent (Figure 9b). More water vapor from the Atlantic sector at 30-40°S, where the cooling is less strong (compared to MIROC 4m SST) and the water vapor more depleted in $\delta^{18}O$ (Figure S4a) because of enhanced evaporation, is transported to eastern Antarctica. As a consequence, the larger cooling in GLOMAP Southern Ocean SST increases the temporal slope at Dome Fuji and EDML stations. A larger SST cooling near the Amundsen Sea (i.e., Tierney et al. SST compared to GLOMAP, Figure 5c) decreases the water vapor transport from this region to western Antarctica sites (Figure S7). The $\delta^{18}O$ change of the water vapor from this source area is relatively

weaker (2 ‰ at maximum, right map of Figure 5c) compared to the decrease in local temperature (2 to 4 °C, left map of Figure 5c). So, less contribution from this source region to West Antarctica inland increases the temporal slopes by ~0.1 % °C$^{-1}$ at WDC and Byrd stations (orange marker in Figure 12). At the same time, this water vapor influences the $\delta^{18}O_p$ of nearby coastal regions like the Antarctic peninsula, making decrease their temporal slopes (Figure 8b and orange marker in Figure S9).

LGM to PI changes in sea ice area fractions have a strong impact on the slopes in coastal regions as shown by the comparison between the plots (c)-(d) and (e)-(f) of Figure 8. Law Dome ice core is representative of this impact in coastal areas, with a slope of 0.35 and 0.64 ‰ °C$^{-1}$ depending if a less extensive sea ice (LGM_miroc4m_sst_and_sic, Figure 8d) or a more extensive one (LGM_miroc4m_sst_glomap_sic, Figure 8c) is used, respectively. A more extensive sea ice in the Atlantic and eastern part of the Southern Ocean changes the transport of vapor only slightly (Figure 9c). On the other hand, the presence of more sea ice covered area close to the continent increases the amount of local water vapor depleted in $\delta^{18}O$ (compared to the $\delta^{18}O$ of open seawater) that influences the isotopic composition of snow in coastal sites, increasing their temporal slopes. A more extensive sea ice increases the slope in a geographical band from Law Dome to Vostok and EDC stations (Figures 8c and d). This can be explained by a decrease of water vapor transport with higher $\delta^{18}O$ concentrations from the Indian Ocean and the south of Australia especially in austral winter (Figure S9). On average, the modeled temporal $\delta^{18}O_p$-$T_{2m}$ slopes of East Antarctic ice core stations are increased by more than 25% when a more extensive sea ice is used (Figure 12, dark and light purple markers against the red and grey ones, respectively). This conclusion remains the same when using the average slope across the entire East Antarctic area but with an enhanced difference because the entire coastal area is under consideration (Figure S9). No clear influence of sea ice extent on temporal slopes at West Antarctica ice core stations is found (Figure 12). On the other hand, effects in the Antarctic peninsula, Ronne Ice Shelf and on the coast of the Amundsen Sea (Figure 8), influencing the average slope values in the entire western part of the continent (Figure S9), are simulated. These are due to changes in the nature of the water source nearby (i.e., open water or sea ice) depending if the sea ice used is less extensive or not.



(a) GLOMAP  (b) More South. Oce. SST cooling  (c) More extensive sea ice  (d) Stronger AMOC

qvi [kg.m$^{-2}$]

Δqvi [kg.m$^{-2}$]

505 **Figure 9: (a) Vertically integrated water vapor transport (arrows) and total column water vapor (qvi, colored background) in LGM_GLOMAP simulation over Antarctic region. Anomalies in transport and concentration of moisture are shown for (b) a stronger SST cooling in the Southern Ocean (LGM_GLOMAP – LGM_miroc4m_sst_glomap_sic), (c) a more extensive sea ice (LGM_miroc4m_sst_glomap_sic – LGM_miroc4m_sst_and_sic) and (d) a stronger LGM AMOC (LGM_miroc4m_strong_AMOC_sst_and_sic – LGM_miroc4m_sst_and_sic).**

510 **4.2 Greenland**

Using ~~both~~a SST field with more LGM cooling off the coast of Greenland and ~~sea ice fields from~~ in the Northern North Atlantic Ocean (GLOMAP) as forcing for ECHAM6-wiso, we model higher $\delta^{18}O_p$-$T_{2m}$ temporal slope values at all Greenland ice core stations (Figure ~~9a~~10a) compared to all other simulations. ~~This is generally true on the entire continent, too (blue marker in Figure S6).~~ The average of the temporal slope values at ice core stations is 0.~~62~~50 ‰ °C$^{-1}$ with GLOMAP sea surface

515 boundary forcing (blue marker in Figure ~~10~~12), and less than 0.~~52~~42 ‰ °C$^{-1}$ in the other ECHAM6-wiso simulations. ~~The influence of both~~ This difference is enhanced by considering the entire Greenland area (blue marker in Figure S9). For Greenland, most of the moisture comes from Northern North Atlantic Ocean at latitudes 30-40° N (Drumond et al., 2016), south of the ice sheet (Figure 11a). A stronger SST cooling in the Arctic – North Atlantic enhances the transport of isotopically depleted water vapor from northern USA (Figure 11b) and strengthens the water vapor transport belt around Iceland. Also,

520 ECHAM6-wiso simulates a slight decrease in local water vapor content in the Greenland Sea (with lower LGM $\delta^{18}O$ because the cooling is stronger there) and its transport to the land. It increases the isotope-temperature temporal slope from the sea to inland, passing through the eastern Greenland coast where Renland station is located (Figures 10b and 10c compared to Figure 10a).
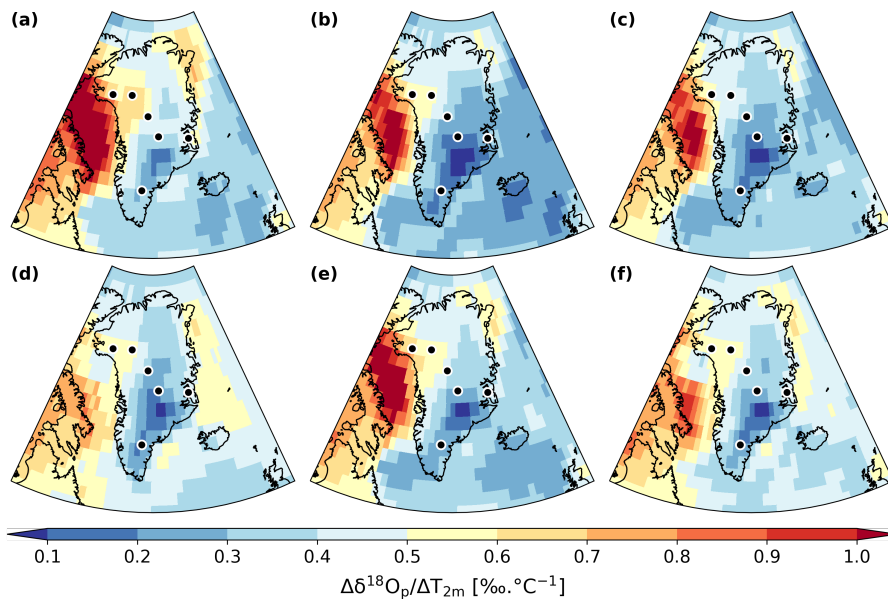


525 **Figure 10: Same as Figure 8 but for the Greenland region.**

A more extensive sea ice ~~boundary fields explains this result.~~

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The LGM-PI SST anomalies off the coast of Greenland are larger in GLOMAP compared to the other reconstructed and modeled SST fields (section 2.2.1), enhancing the cooling over North Atlantic Ocean (left maps of Figure 5) and so the LGM-PI anomalies in $\delta^{18}O_P$ (maps on the right side of Figure 5). The water vapor from this region is transported further north over the Greenland Sea during summer when sea ice shrinks. In Greenland Sea, local SST change is small while $\delta^{18}O_P$ anomalies are strong because of this water vapor transport. Then, the use of GLOMAP SST to force ECHAM6-wiso results in less cooling in the Greenland Sea area but stronger $\delta^{18}O_P$ anomalies compared to ECHAM6-wiso simulations using other SST boundary conditions (Figure 5). As a consequence, the modeled $\delta^{18}O_P$-$T_{2m}$ temporal slopes are higher than 1 ‰ °C⁻¹ over the Greenland Sea with GLOMAP SST (Figure 9a). The modeled slopes range only between 0.4 and 0.8 ‰ °C⁻¹ in the other simulations (Figures 9b to 9f). This affects the temporal slope at the Renland coastal station, where a temporal slope of 0.81 ‰ °C⁻¹ is simulated with GLOMAP SST, while this slope is below 0.6 ‰ °C⁻¹ in other simulations. Moreover, the larger cooling off the coast of Greenland with GLOMAP SST influences the modeled $\delta^{18}O_P$-$T_{2m}$ temporal slopes in inland Greenland ice core stations (Figure 9a) through changes in inland temperature (left maps of Figure 5) and also the inland transport of oceanic water vapor from the North Atlantic Ocean and the Baffin Bay.

The use of GLOMAP or MIROC 4m sea ice boundary as forcing for ECHAM6-wiso simulations lead to mixed results in terms of modeled $\delta^{18}O_P$-$T_{2m}$ temporal slopes. The MIROC 4m sea ice in Greenland Sea shrinks less in summer compared to the one from GLOMAP (Figure 2). The effect on temperature is low but it enhances the LGM-PI anomalies in isotopic composition of precipitation over this area (Figure 6c), increasing the modeled $\delta^{18}O_P$-$T_{2m}$ temporal slopes (Figures 9d and 9f). (i.e., MIROC 4m compared to GLOMAP) makes the Arctic Ocean area drier, especially at 50° N, and it slightly slows down the transport of water vapor from the North Atlantic to Greenland area (Figure 11c). On the other hand, all this area is covered by sea ice in the "more extensive sea ice" simulation (i.e., MIROC 4m sea ice). It decreases the $\delta^{18}O$ of water vapor above this surface, increasing the isotope-temperature temporal slope on the eastern Greenland coast and in the Greenland Sea (Figures 10d and 10f compared to Figures 10c and 10e, respectively). In the latter, a more extensive sea ice, especially in summer, decreases the LGM $\delta^{18}O$ too, while the effect on temperature is small, again increasing the local temporal slope. It has a slight effect on modeled temporal slopes (~0.1 ‰ °C⁻¹) over the eastern coastal regions of Greenland, including the Renland station. In ECHAM6-wiso, the isotopic composition of sea ice surfaces also reflects the isotopic composition of snow deposited on this surface. Then the formation of new sea ice from PI to LGM acts as a positive feedback effect in the decrease of surrounding $\delta^{18}O_P$, leading to steeper modeled $\delta^{18}O_P$-$T_{2m}$ temporal slopes (see textText S1 in Supplementary Material and Figure S7S10). Finally, ECHAM6-wiso forced with MIROC 4m sea ice, whose fractional areas are artificially lower (i.e., not 100% sea ice covered) in coastal grid cells, simulates lower $\delta^{18}O_P$-$T_{2m}$ temporal slope values over Baffin Bay (between 0.3 and 0.69 ‰ °C⁻¹, Figures 9d and 9f) compared to when the model is forced with GLOMAP sea ice (between 0.76 and 1.25 ‰ °C⁻¹, Figures 9c10c and 9e10e). If the MIROC 4m sea ice is corrected by setting sea ice fraction as 100‰ as in GLOMAP (see textText S1 in Supplementary Material and Figure S8S11), we obtain temporal slope values similar to those in the simulations forced by GLOMAP sea ice (Figure S9). It also slightly increases the $\delta^{18}O_P$-$T_{2m}$ temporal slopes of inland ice core stations like NGRIP

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(0.53 and 0.67 ‰ °C$^{-1}$ with original (Figure 9d) and modified MIROC 4m sea ice, respectively). This result shows that the presence or absence of sea ice very near the coast has a large influence on the modeled temporal slopes in some Greenland ice core stations (Figure 11).S12).

The AMOC strength during LGM influences both the SST and the sea ice distribution in the Arctic region. While stronger LGM AMOC weakens the isotopic model-data agreement in Greenland because the predominantly less extensive sea ice reduces the modeled surface cooling (see section 3.4), it generally does not impact the temporal slopes modeled by ECHAM6-wiso (red and brown lines in Figure 10) over inner Greenland. The ice core station that is most sensitive to the change in LGM AMOC strength is Renland, where the modeled $\delta^{18}O_p$-$T_{2m}$ temporal slope is decreased from 0.60 to 0.55 ‰ °C$^{-1}$ (Figures 9d and 9f).



$$\Delta_{LGM-PI}\delta^{18}O_p/\Delta_{LGM-PI}T_{2m} \ [‰.°C^{-1}]$$
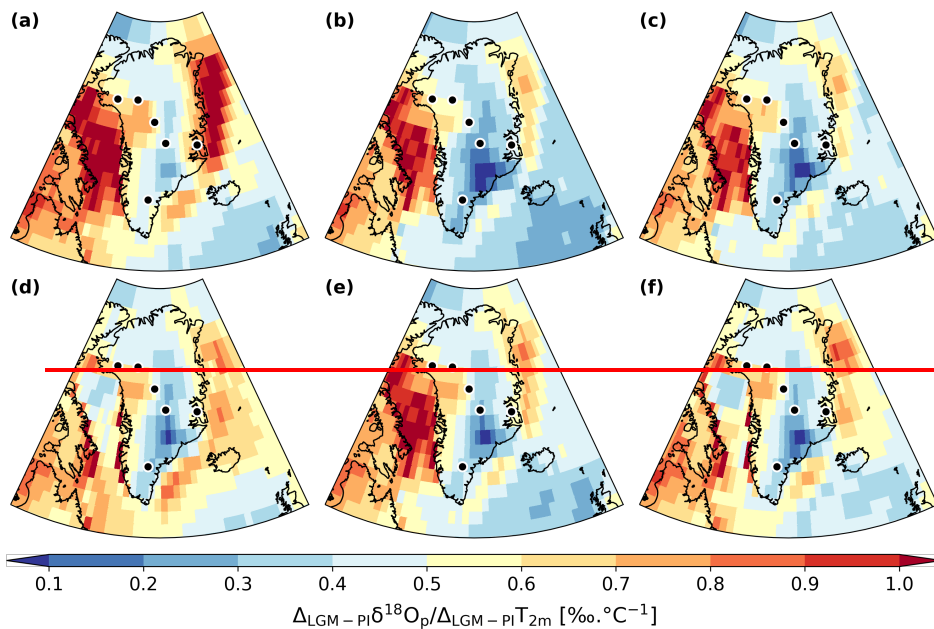
Figure A stronger AMOC increases the amount of water vapor and enhances its transport from the North Atlantic to European coastal areas because of the less extensive sea ice (Figure 11d). More water vapor with higher $\delta^{18}O$ is present southeast of Greenland because sea ice is replaced by open water. However, there is only a slight increase in the transport of this water vapor toward north in the Greenland interior (Figure 11d) while the cooling inland is largely reduced (Figure 7c). So, isotope-temperature temporal slopes are slightly increased over the interior of Greenland for stronger AMOC (dark and light purple markers in Figures 12 and S9). On the other hand, temporal slopes are decreased over the Greenland Sea (Figure 10f) because of the presence of open water instead of sea ice, enhancing the LGM $\delta^{18}O_p$ locally. Note that stronger AMOC worsens the model-data agreement for both $\Delta T_{2m}$ and $\Delta\delta^{18}O$ (Figure 7a).
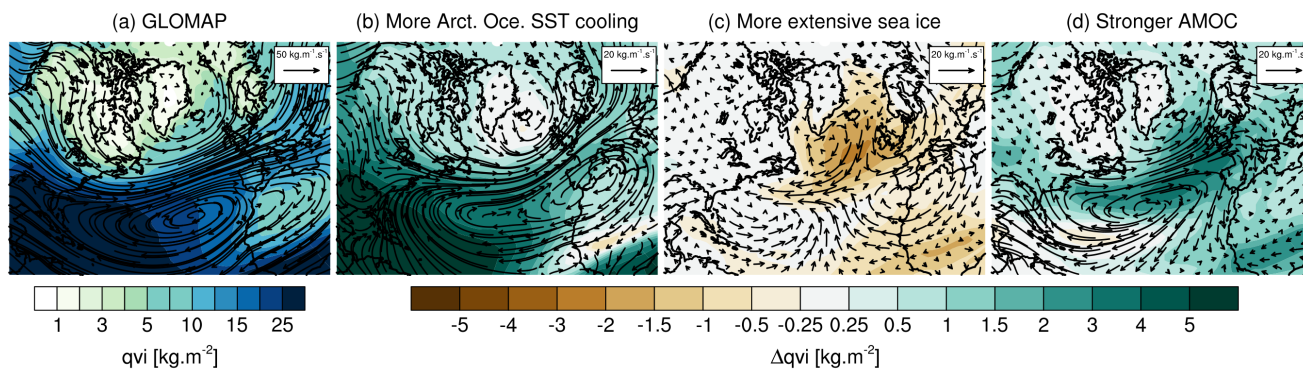
(a) GLOMAP     (b) More Arct. Oce. SST cooling     (c) More extensive sea ice     (d) Stronger AMOC

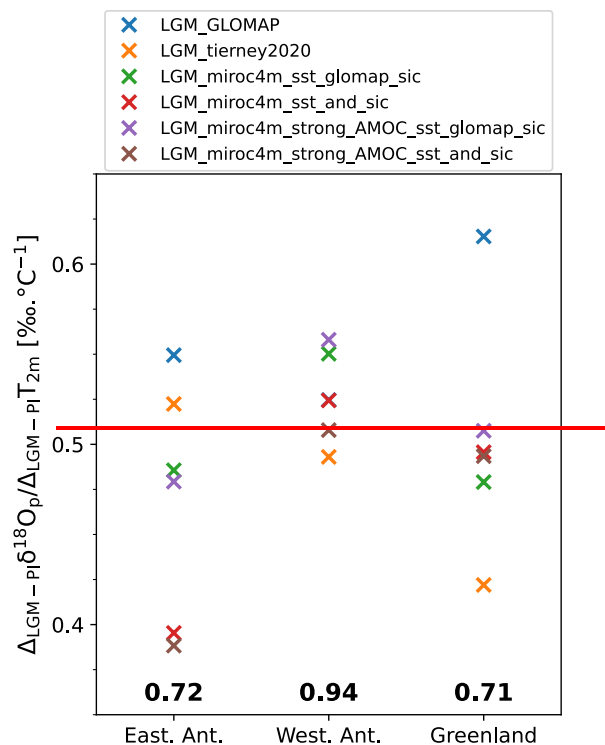qvi [kg.m$^{-2}$]

$\Delta$qvi [kg.m$^{-2}$]

**Figure 11~~911: Same as Figure 8 but for the Greenland region.~~**



**: (a) Vertically integrated water vapor transport (arrows) and total column water vapor (qvi, colored background) in LGM_GLOMAP simulation over Arctic region. Anomalies in transport and concentration of moisture are shown for (b) a stronger SST cooling in Arctic Ocean (LGM_GLOMAP − LGM_tierney2020), (c) a more extensive sea ice (LGM_miroc4m_sst_and_sic − LGM_miroc4m_sst_glomap_sic) and (d) a stronger LGM AMOC (LGM_miroc4m_strong_AMOC_sst_and_sic − LGM_miroc4m_sst_and_sic).**

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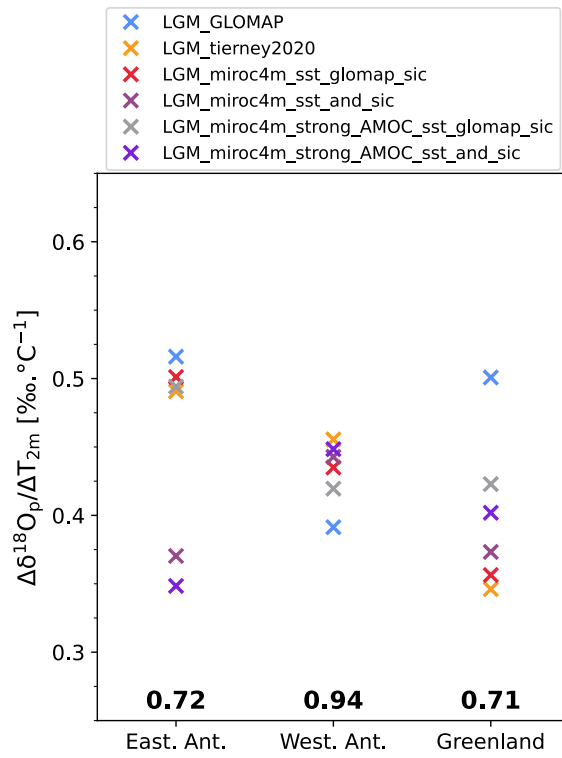**Figure 12~~1012~~**: Average modeled values of $\delta^{18}O_p$-$T_{2m}$ temporal slope for East Antarctic, West Antarctic and Greenland ice core stations according to our different simulations. Numbers in bold are the values of the corresponding modeled mean PI spatial slopes.

## 5 Conclusions and Perspectives

In this study, we ~~raised~~elevated the importance of sea surface boundary conditions on the relationship between near-surface air temperature and $\delta^{18}O_p$ for LGM to PI climate change. Figure ~~11~~13 illustrates the main findings of our study. In ~~East~~ Antarctica, ~~we noted~~ the ~~contrast between coastal regions and inland area in terms of control on the $\delta^{18}O_p$-$T_{2m}$~~ temporal slopes ~~(left map of Figure 11). The coastal site Law Dome is greatly affected by~~ in coastal area like the Antarctic peninsula, the ~~LGM~~ Ronne Ice Shelf, near the Amundsen Sea and the eastern coast are influenced by sea ice extent~~, with more than double the~~. For example, ECHAM6-wiso simulates almost a doubling in the value of the temporal slope at Law Dome and a better isotopic model-data agreement if ~~GLOMAP~~more extensive sea ice is used instead of the ~~MIROC 4m~~less extensive one (Figures 8c (GLOMAP sea ice) and 8d~~,~~ (MIROC 4m), respectively). The transport of moisture to Antarctica is generally changed only slightly changed with variations in the sea ice extent. On the other hand, ~~no noteworthy change in the $\delta^{18}O_p$-$T_{2m}$~~the distribution of open water and sea ice covered areas around the continent mainly control the nearby $\delta^{18}O_p$ and so the isotope-temperature temporal slopes. We found that temporal slope ~~is simulated regardless of the SST boundary conditions. A cooling larger by 1-2 °C in the Southern Ocean near Law Dome (Figure 5c)~~ further inland, at the EDC and Vostok sites, are influenced by changes ~~the temporal slope value at this station only by a small amount (less than 0.05 ‰.°C⁻¹).~~in sea ice extent through a weakening

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of moisture transport from the Indian Ocean and the south of Australia when sea ice is more extensive (left map Figure 13).

The values of $\delta^{18}O_p$-$T_{2m}$ temporal slopes in inland East Antarctic ice core stations like Dome F, EDML, EDC, ~~Vostok~~ and Talos are mainly controlled by the change of SST in our ECHAM6-wiso simulations. ~~Stronger cooling in Atlantic sector of the Southern Ocean (GLOMAP) leads to steeper temporal slopes in Dome F and EDML (between 0 and 40° E). Similarly, stronger cooling in the eastern part of the Southern Ocean (Tierney et al., 2020) increases significantly the $\delta^{18}O_p$-$T_{2m}$ temporal slopes at EDC, Vostok and Talos. The steeper $\delta^{18}O_p$-$T_{2m}$ temporal slopes are accompanied by better isotopic model-data agreement at these ice core stations (Figure 5a). The sea ice distribution can impact the $\delta^{18}O_p$-$T_{2m}$ temporal slopes at some inland stations, like EDML, but to a lesser extent compared to coastal regions. In West Antarctica, we showed that sea surface boundary conditions have a mixed effect on the $\delta^{18}O_p$-$T_{2m}$ temporal slopes (left map of Figure 11). For example, a steeper slope at WDC is simulated for more extensive sea ice in the western part of the Southern Ocean, while no significant effect on the $\delta^{18}O_p$-$T_{2m}$ temporal slope at Byrd, located very near WDC station, is seen. We note that a larger change in sea ice extent increases the mean $\delta^{18}O_p$-$T_{2m}$ temporal slope across the entire West Antarctic region (Figure S6). Larger sea surface cooling in the western part of the Southern Ocean slightly increases the $\delta^{18}O_p$-$T_{2m}$ temporal slopes at Byrd and WDC stations, and reduces it at Siple Dome. The lower change in $\delta^{18}O_p$ compared to temperature could be explained by the influence of water vapor transport in this region. In Greenland, our~~Stronger cooling in the Atlantic sector of the Southern Ocean (GLOMAP) leads to steeper temporal slopes in Dome F and EDML due to enhanced water vapor transport from lower latitudes. Strong cooling in the Admunsen Sea weakens the transport of relatively less depleted water vapor (compared to the large cooling) inland West Antarctica. It slightly increases the temporal slopes at the WDC and Byrd sites. At the same time, this water vapor contributes to nearby coastal region, decreasing the temporal slopes there (left map of Figure 13). These various effects on the temporal slopes can complexify the interpretation of $\delta^{18}O_p$ signals in West Antarctica, depending on the sites considered. Our modeled results demonstrate clearly that the $\delta^{18}O_p$-$T_{2m}$ temporal slopes in Greenland are influenced by the sea surface temperatures very near the coasts. The greater the LGM cooling off the coast of southeast Greenland, the larger the $\delta^{18}O_p$-$T_{2m}$ temporal slopes (right map of Figure ~~11). The cooling in this region of the North Atlantic in the GLOMAP reconstruction is larger than in the Tierney et al. reconstruction or MIROC 4m model results, giving steeper modeled temporal slopes (0.62 and 0.42 ‰ °C⁻¹, respectively, on average on all Greenland stations) and generally a better agreement with isotopic data (Figure 5a).~~13) because of an enhancement in the transport of isotopically depleted moistu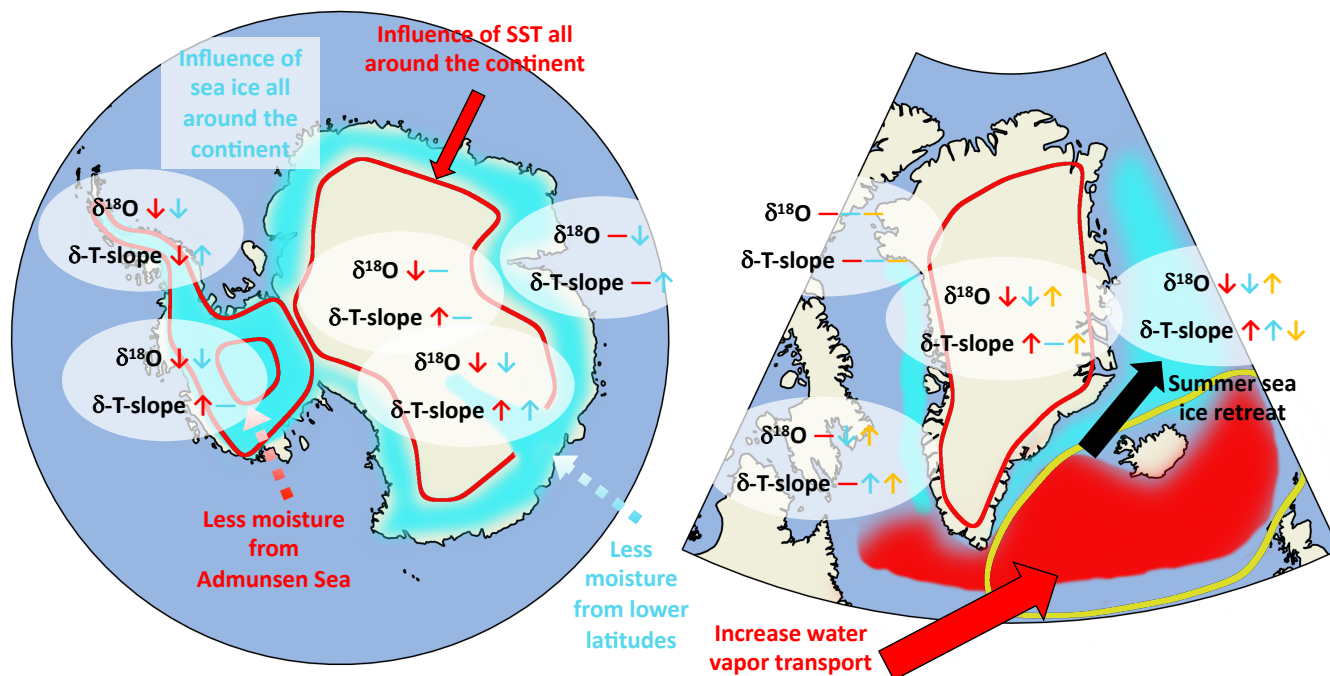re from the western North Atlantic Ocean. Similarly, the presence or absence of sea ice very near the coast can impact the modeled temporal slopes in some Greenland ice core stations~~, and~~. It has a large influence in Baffin Bay and the Greenland Sea.~~ The~~, too. A more or less large southern extent of the sea ice ~~is not so important, as shown by the similar~~(MIROC 4m in weak AMOC phase or GLOMAP, respectively) has only a slight impact on modeled temporal slope values ~~using GLOMAP or MIROC 4m ~~(light purple and red markers in Figure 12, respectively). Water vapor transported to the Arctic region is more depleted when sea ice ~~(weak AMOC phase).~~ ~~On~~is more extensive, but less is transported to the ~~other hand, the~~Greenland region. The seasonal variation of the sea ice distribution, especially its retreat in summer, changes the origin type of the water source (from evaporation of open water or

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sublimation of snow on sea ice) and so influences the ~~water~~$\delta^{18}$O of local vapor ~~transport in this region and~~contributing to the ~~modeled~~ $\delta^{18}$O$_P$-T$_{2m}$ temporal slopes in the eastern part of ~~the~~ inner Greenland (right map of Figure 11). Finally, ~~while~~a stronger LGM AMOC ~~reduces the isotopic model-data agreement, it generally does not impact~~ impacts slightly the temporal slopes modeled by ECHAM6-wiso over Greenland mainly because of the changes in the sea ice distribution. Higher temporal slopes are modeled by ECHAM6-wiso mainly because the cooling in Greenland is largely reduced while changes in $\delta^{18}$O anomalies remain localized in the Greenland Sea. Variations in the temporal slope values are located in the Greenland Sea, where the changes in surface temperature and sea ice distribution due to the AMOC strength mainly occur. For Antarctica, only small changes in surface temperature and $\delta^{18}$O are modeled by ECHAM6-wiso because the strong phase period was selected ~~in~~from within the middle of the AMOC peak. The impact of the precise period ~~selection for~~chosen to represent the strong AMOC phase, ~~like~~for example the start or the end of the interstadial, will be investigated in more ~~in~~ detail in a future study.
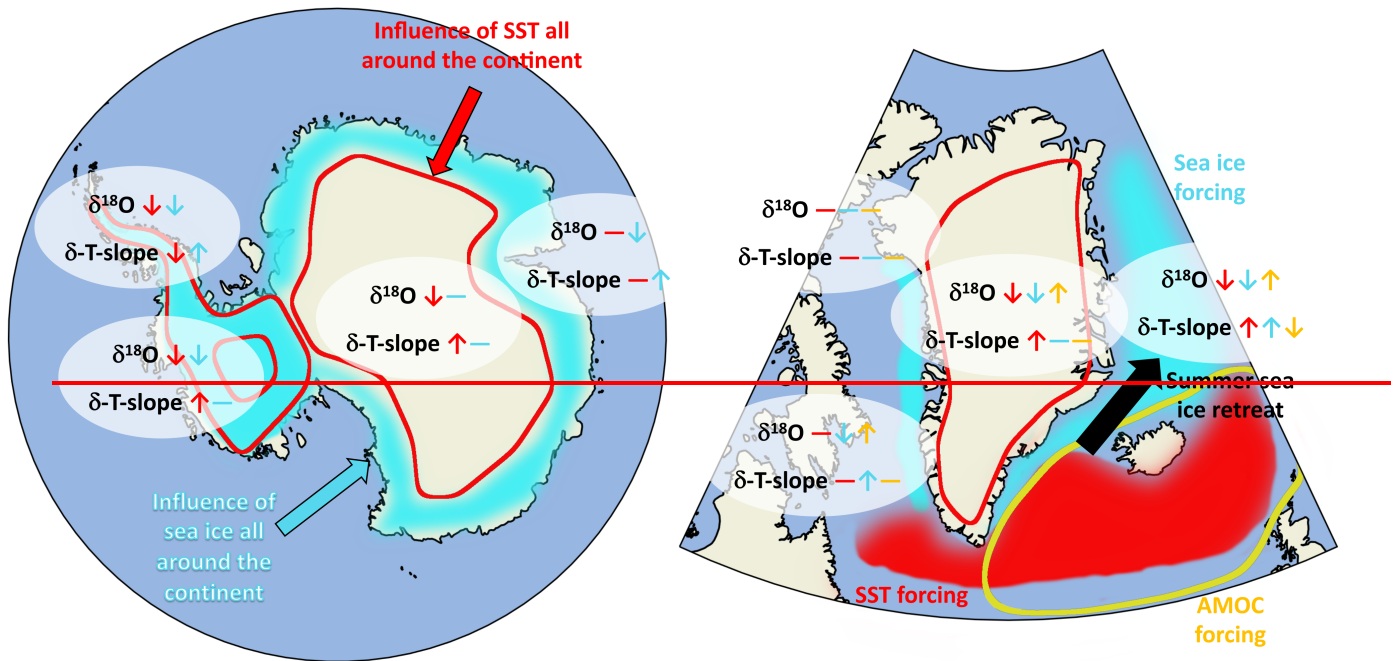
**Figure 13~~1113~~:** Summary figure illustrating the influence of higher sea surface cooling, larger sea ice extent and stronger LGM AMOC (in red, cyan and yellow, respectively) on the modeled LGM $\delta^{18}O$ and temporal $\delta^{18}O_p$-$T_{2m}$ slopes in the Antarctic and Greenland regions (left and right, respectively). The up and down arrows indicate higher and lower values, respectively. The horizontal lines indicate no significant change.

In Greenland, ECHAM6-wiso simulates ~~on average~~ $\delta^{18}O_p$-$T_{2m}$ temporal slopes <u>oscillating between 0.2 and 0.7 ‰ °C$^{-1}$ inland and at northwestern coastal sites, respectively,</u> lower than the spatial one (0.71 ‰ °C$^{-1}$, Figure ~~10~~12), as already reported in previous studies (Buizert et al., 2014; Cauquoin et al., 2019b; Jouzel et al., 1999; Werner et al., 2000). <u>Our modeled temporal slope values for stations NEEM (around 0.7 ‰ °C$^{-1}$) and NGRIP (between 0.37 and 0.57 ‰ °C$^{-1}$) are in agreement with previous reconstructions (Buizert et al., 2014), too.</u> In Antarctica, the ECHAM6-wiso modeled $\delta^{18}O_p$-$T_{2m}$ temporal slopes for LGM-to-PI climate change are on average lower than the PI spatial slopes of the same model by at least 0.~~15~~20 and 0.~~38~~48 ‰ °C$^{-1}$ for eastern and western ice core locations, respectively (Figure ~~10~~12), regardless <u>of</u> the simulation <u>being</u> considered. By extension, we found much lower temporal slope values than the ones estimated by Buizert et al. (2021). We simulate a maximum temporal slope value of 0.~~8~~9 ‰ °C$^{-1}$ for ~~Dome Fuji in~~ the ~~LGM_GLOMAP simulation~~<u>South Pole</u>, while Buizert et al. (2021) found temporal slopes in Antarctic ice core stations ranging from 0.9 to 1.4 ‰ °C$^{-1}$, which are higher than the observed spatial $\delta^{18}O_p$-$T_{2m}$ slope of 0.8 ‰ °C$^{-1}$ (Masson-Delmotte et al., 2008). ~~Compared to~~ <u>The use of the thicker</u> PMIP3 ice sheet reconstruction,~~ the use of~~ <u>compared to</u> GLAC-1D ~~to run LGM simulations reduces the isotopic model-data agreement for Antarctica (Figure S4). Also, the use of the old PMIP3 ice sheet reconstruction in ECHAM6-wiso~~ increases the resulting modeled $\delta^{18}O_p$-$T_{2m}$ temporal slopes <u>in ECHAM6-wiso</u> (Figure ~~S10~~<u>S13</u>) with mean values for East and West Antarctic ice core

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stations equal to 0.68 and 0.92 ‰ °C$^{-1}$, respectively, ~~which are closer to the modeled mean PI spatial slopes (0.71 and 0.94 ‰ °C$^{-1}$, respectively) but still lower than the Buizert et al. (2021) results. So, the variability in LGM ice sheet reconstructions affects our modeled δ$^{18}$O$_p$-T$_{2m}$ temporal slopes for LGM-to-PI climate change in Antarctica, as already shown by Werner et al. (2018). On the other hand~~by decreasing the isotopically enriched bias in the model for LGM (Figure S5). However, the temperature model-data agreement is reduced in this case. Even if ECHAM6-wiso show biases in the representation of the isotopic content of Antarctic precipitation, we insist that the purpose of our study was to investigate the relative effects of sea surface conditions and AMOC strength on the links between δ$^{18}$O$_p$ and near-surface air temperature, regardless ~~the~~of agreement or disagreement with other slope reconstructions.

In addition to orography effects, fractionation during the sublimation of surface ice is not taken into account in ECHAM6-wiso as in many isotope-enabled AGCMs. This process would lead to a further decrease in the δ$^{18}$O of water vapor in the polar regions, contributing to steeper modeled δ$^{18}$O$_p$-T$_{2m}$ temporal slopes in regions with low temperature. The mismatch between our model slopes and the reconstructed ones from Buizert et al. (2021) could be related to the representation of the atmospheric boundary layer and the related inversion temperature (Krinner et al., 1997; Masson-Delmotte et al., 2006; Cauquoin et al., 2019a), too. Still, despite these biases that potentially affect our modeled δ$^{18}$O$_p$-T$_{2m}$ temporal slopes for LGM-to-PI climate change, our ensemble of simulations provides information on how sea surface conditions partially control the links between δ$^{18}$O$_p$ and near-surface air temperature in the polar regions, through changes in the sources or transport of moisture arriving at the poles.

Because only ECHAM6-wiso is used in this study, we cannot exclude the model-dependency of our results. So, the use of isotope-enabled AGCMs other than ECHAM6-wiso would be beneficial to confirm or refute our findings. A set of SST reconstructions for the LGM, based on both model results and observations, are now available. We ~~raise~~stress the importance of providing sea ice cover reconstruction, which contributes not only to the δ$^{18}$O$_p$ of coastal sites but also of some inland ice core stations, for this period too. The sea ice cover simulated by coupled GCMs for the LGM period takes various forms. An alternative reconstruction to the GLOMAP one, also based on observations, would help to better assess the impact of sea ice cover on the δ$^{18}$O$_p$-T$_{2m}$ relationship for LGM to PI climate change. We also emphasize that more proxy measurements of temperature and sea ice are necessary for the Southern Ocean. Relatively large uncertainties remain in the reconstruction of the climatology in this area while the water vapor from this region contributes largely to δ$^{18}$O$_p$ in Antarctica. As a first step, the focus of this study was to identify and quantify the important factors influencing the isotope-temperature relationship in the polar areas for the LGM to PI climate change. Future studies will investigate the evolution of this relationship along the whole of the last deglaciation. For that, an ensemble of equilibrium isotopic simulations using the sea surface and ice sheet boundary conditions from MIROC 4m for different time periods between the LGM and PI will be performed.

*Author contributions.* AC designed the model experiments and performed the simulations using the MIROC 4m sea surface boundary conditions with the help of AAO, TO and WLC. AC performed the simulations using the GLOMAP or Tierney et al. sea surface boundary conditions with the help of MW and AP. AC and all the co-authors analyzed the model outputs. AC wrote the manuscript with contributions from all co-authors.

*Competing interests.* One of the co-author (André Paul) is editor in Climate of the Past.

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