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# 1 Importance of the advection scheme for the simulation of water 2 isotopes over Antarctica by atmospheric general circulation 3 models: a case study for present-day and Last Glacial Maximum 4 with LMDZ-iso

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13 **Abstract.** Atmospheric general circulation models (AGCMs) are known to have a warm and isotopically enriched  
14 bias over Antarctica. We test here the hypothesis that these biases are consequences of a too diffusive advection.  
15 Exploiting the LMDZ-iso model, we show that a less diffusive representation of the advection, especially on the  
16 horizontal, is very important to reduce the bias in the isotopic contents of precipitation above this area. The  
17 choice of an appropriate representation of the advection is thus essential when using GCMs for paleoclimate  
18 applications based on polar water isotopes. Too much diffusive mixing along the poleward transport leads to  
19 overestimated isotopic contents in water vapor because dehydration by mixing follows a more enriched path  
20 than dehydration by Rayleigh distillation. The near-air surface temperature is also influenced, in a more minor  
21 way, by the diffusive properties of the advection scheme directly via the advection of the air and indirectly via  
22 the radiative effects of changes in high cloud fraction and water vapor. A too diffusive horizontal advection  
23 increases the temperature and so also contributes to enrich the isotopic contents of water vapor over Antarctica  
24 through a reduction of the distillation. The temporal relationship, from Last Glacial Maximum (LGM) to present-  
25 day conditions, between the mean annual near-air surface temperature and the water isotopic contents of  
26 precipitation for a specific location can also be impacted, with significant consequences on the paleo-  
27 temperature reconstruction from observed changes in water isotopes.

## 28 1 Introduction

29 Water stable isotopologues (hereafter designated by the term “water isotopes”), are integrated tracers of the  
30 water cycle. Especially, the isotopic composition recorded in polar ice cores enabled the reconstruction of past  
31 temperature variations (Jouzel, 2013 and references therein). For example, low accumulation sites that are  
32 typical on the East Antarctic Plateau ( $< 10$  cm water-equivalent  $\text{yr}^{-1}$ ) provided the longest ice core records,  
33 allowing to reconstruct past climate over several glacial-interglacial cycles (Jouzel et al., 2007). However, the  
34 interpretation of isotope signals remains challenging because of the numerous and complex processes involved

35 (water vapor transport, fractionation during the phase changes in the water cycle, distillation effect...). This is  
36 particularly the case for Antarctica because this part of the world is subject to extreme weather conditions.

37 To improve our knowledge on the mechanisms controlling the water isotopes distribution, atmospheric general  
38 circulation models (AGCMs) enhanced by the capability to explicitly simulate the hydrological cycle of the water  
39 isotopes ( $\text{H}_2^{16}\text{O}$ ,  $\text{HDO}$ ,  $\text{H}_2^{17}\text{O}$ ,  $\text{H}_2^{18}\text{O}$ ) are now frequently used (Joussaume et al., 1984; Risi et al., 2010a; Werner  
40 et al., 2011). Water isotopes in climate models have been used, for example, to better understand how the  
41 climatic signal is recorded by isotopes in polar ice cores at paleoclimatic time scales (Werner et al., 2001).

42 However, some issues remain concerning the simulation of the climate over the Antarctic continent by AGCMs.  
43 For example, they frequently present a near-surface warm bias over this area (Masson-Delmotte et al., 2006)  
44 and isotopic values in precipitation that are not depleted enough compared to observations (Lee et al., 2007; Risi  
45 et al., 2010a; Werner et al., 2011). This raises the question why many of the AGCMs have these warm and  
46 enriched in heavy water isotopes biases over Antarctica.

47 In this paper, we hypothesize that one part of these biases is associated with an excessively diffusive water vapor  
48 transport, i.e. transport that is associated with too much mixing. According to previous studies, the diffusive  
49 properties of the advection scheme in the AGCMs, on the horizontal as well on the vertical, can have an impact  
50 on the simulation of humidity and of its water isotope contents. On the horizontal, dehydration of air masses by  
51 mixing with a drier air mass leads to more enriched water vapor than dehydration by condensation and  
52 associated Rayleigh distillation (Galewsky and Hurley, 2010). For the same reason, poleward water vapor  
53 transport by eddies (which act as mixing) leads to more enriched water vapor in Antarctica than transport by  
54 steady advection (Hendricks et al., 2000). On the vertical, the excessive diffusion during water vapor transport  
55 seems to be the cause of the moist bias found in most AGCMs in the tropical and subtropical mid and upper  
56 troposphere, and of the poor simulation of isotopic seasonality in the subtropics (Risi et al., 2012). The diffusivity  
57 of the advection scheme in the vertical has also important consequences on modeling of tracers like tritium by  
58 affecting greatly its residence time in the stratosphere, and so its downward transport from the stratosphere to  
59 the troposphere (Cauquoin et al., 2016).

60 The goal of this paper is to test whether the warm and enriched biases in Antarctica are associated with an  
61 excessively diffusive water vapor transport, both on the horizontal and on the vertical. The diffusive character of  
62 the advection can be varied by modifying either the advection scheme or the resolution of the simulation, and  
63 we test both possibilities. Finally, we explore if a too diffusive water vapor transport can affect the temporal  
64 water isotopes – temperature slope between the Last Glacial Maximum (LGM, 21 ka) and present-day periods.

## 65 2 Model, simulations and data

### 66 2.1 Model and simulations

67 We use here the isotopic AGCM LMDZ-iso (Risi et al., 2010a) at a standard latitude-longitude R96 grid resolution  
68 ( $2.5^\circ \times 3.75^\circ$ ), and with 39 layers in the vertical spread in a way to ensure a realistic description of the  
69 stratosphere and of the Brewer-Dobson circulation (Lott et al., 2005). Water isotopes are implemented in a way  
70 similar to other state-of-the-art isotope-enabled AGCM (Risi et al., 2010a). The isotopic composition of glacier  
71  $R_{\text{glacier}}$  is calculated prognostically in the model. It is a precipitation-weighted average of the previous snow fall.  
72 At each time step and in each grid box, it is updated as:

$$73 R_{\text{glacier}}(t + dt) = \frac{h_{\text{glacier}} \times R_{\text{glacier}}(t) + iso_{\text{snowfall}} \times dt}{h_{\text{glacier}} + H_2O_{\text{snowfall}} \times dt}, \quad (1)$$

74 with  $h_{\text{glacier}} = 20 \text{ kg.m}^{-2}$  the height scale for the glacier,  $iso_{\text{snowfall}}$  and  $H_2O_{\text{snowfall}}$  the snowfall of isotopes and  
75 standard water in  $\text{kg.m}^{-2}.\text{s}^{-1}$  and  $dt = 30 \times 60 \text{ s}$ . No fractionation is assumed during the runoff and sublimation of  
76 glaciers ice. The model has been validated at global scale for the simulation of both atmospheric (Hourdin et al.,  
77 2006) and isotopic (Risi et al., 2010a) variables, and has been extensively compared to various isotopic  
78 measurements in polar regions (Casado et al., 2013; Steen-Larsen et al., 2013, 2014, 2017; Bonne et al., 2014,  
79 2015; Touzeau et al., 2016; Ritter et al., 2016; Stenni et al., 2016). LMDZ-iso is also able to simulate the  $\text{H}_2^{17}\text{O}$   
80 distribution (Risi et al., 2013) but we do not consider it here because the limitations inherent to the AGCMs lead  
81 to strong uncertainties and numerical errors on the spatio-temporal distribution of this isotope. 2 years of spin-  
82 up have been performed for all the simulations presented hereafter.

83 To quantify the effects of the prescribed advection scheme on water stable isotope values over Antarctica, we  
84 first performed three sensitivity simulations with LMDZ-iso under present-day conditions for the post spin-up  
85 period 1990-2008 (i.e. 19 model years), following the model setup from Cauquoin et al. (2016) i.e. simulations  
86 follow the AMIP protocol (Gates, 1992), forced by monthly observed sea-surface temperatures and nudged by  
87 the horizontal winds from 20CR reanalyses (Compo et al., 2011): (1) one control simulation with the van Leer  
88 (1977) advection scheme (called VL), which is a second order monotonic finite volume scheme prescribed by  
89 default in the standard version of the model (Risi et al., 2010a); and two other simulations whose the van Leer  
90 advection scheme has been replaced by a single upstream scheme (Godunov, 1959) on (2) the horizontal plane  
91 (UP\_xy) and on (3) the vertical direction (UP\_z). Depending on one tunable parameter, the LMDZ model can be  
92 used with these 2 versions of the advection scheme according to the object of study (Risi et al., 2012). The  
93 advection scheme in the simulations presented in the LMDZ-iso reference paper from Risi et al. (2010a) was set  
94 erroneously to the simple upstream scheme rather than to the van Leer's scheme (Risi et al., 2010b), and has  
95 little influence on their simulated spatial and temporal distributions of water isotopes at a global scale. However,  
96 as we will show here, this has considerable effect on the spatial distribution of these proxies over region with  
97 extreme weather conditions such as Antarctica. The 2-year spin-up time is enough to reach equilibrium. In the  
98 VL simulation, the globally and annually average values of temperature and  $\delta^{18}\text{O}$  in precipitation for 1990 are

99 13.06°C and -7.53‰ respectively, very close to the average values over the whole period 1990-2008 (13.14°C  
100 and -7.57 ‰) within the average interannual variability of 0.15°C and 0.06 ‰. The conclusion is the same if we  
101 focus on the 60°S-90°S area instead: the average values of temperature and  $\delta^{18}\text{O}$  in precipitation for the year  
102 1990 are -17.89°C and -25.45 ‰ respectively, very close to the average values over the whole period 1990-2008  
103 (-17.71°C and -25.34 ‰) within the average interannual variability of 0.43°C and 0.29 ‰.

104 The detailed description of the mixing ratio by the van Leer's (1977) scheme and its comparison with the  
105 upstream scheme (Godunov, 1959) can be found in the Appendix A of Cauquoin et al. (2016). To resume, the  
106 mixing ratio at the left boundary of box  $i$ ,  $q_{i-1/2}$ , is calculated as a linear combination of the mixing ratio in the  
107 boxes  $i-1$  and  $i$  in the van Leer's scheme whereas in the upstream scheme  $q_{i-1/2} = q_{i-1}$ . This means that in the  
108 upstream scheme, even if the air mass flux from grid box  $i-1$  to grid box  $i$  is very small, the air that is advected  
109 into box  $i$  has the same water vapor mixing ratio as grid box  $i-1$  as a whole. This makes the upstream scheme  
110 much more diffusive. An example of the effect of these 2 different advection schemes in a very idealized case is  
111 given in the figure 3 (top panel) of Hourdin and Armengaud (1999). In a single dimension case, a tracer  
112 distribution is initially rectangle and is advected with a constant velocity such that Courant number is equal to  
113 0.2. After 70 iterations, the initial rectangle shape is almost unchanged with the van Leer scheme, whereas it  
114 spreads into a flat Gaussian shape with the upstream scheme. Quantitatively, almost half of the tracer mass  
115 becomes outside the initial rectangle shape with the upstream scheme.

116 Increasing the grid resolution is equivalent to using an advection scheme that is less diffusive. Indeed, these  
117 finite-difference schemes are discretization methods and so depend on the chosen spatial resolution. To check  
118 that our findings and conclusions are consistent, we performed two more UP\_xy and VL simulations under  
119 present-day conditions with the same configuration as presented above but at the R144 resolution (latitude-  
120 longitude grid resolution of  $1.27^\circ \times 2.5^\circ$ ).

121 Finally, we evaluate the impacts of applying advection schemes with different diffusive properties on the  
122 temporal relationship between water isotopic contents of precipitation and mean air surface temperature,  
123 essential for paleo-temperature reconstructions. For this, we effected 12 years long (the two first years being  
124 used for the spin-up) UP\_xy and VL simulations for present-day "free" (i.e. not nudged by the 20CR reanalyses,  
125 still following the AMIP protocol) and LGM conditions at the R96 resolution. For the LGM simulations, the PMIP3  
126 protocol is applied (Braconnot et al., 2012). Orbital parameters and greenhouse gas concentrations are set to  
127 their LGM values. ICE-5G ice sheet conditions are applied (Peltier, 1994). LMDZ-iso is forced by the climatological  
128 sea-surface temperatures (SSTs) and sea ice from the IPSL-CM4 model (Marti et al., 2005). The SST simulated by  
129 IPSL-CM4 for pre-industrial conditions (PI) has a global bias of -0.95 Kelvin with a cold bias in the mid-latitudes,  
130 a warm bias on the eastern side of the tropical oceans and on the Southern Ocean, and a particularly strong cold  
131 bias in the North Atlantic (Hourdin et al., 2013). To avoid confusing these biases with LGM – present-day signals,  
132 we use the SSTs from an IPSL PI simulation in the following way to cancel out the biases in the IPSL model common  
133 to both the LGM and PI simulations (see details in Risi et al., 2010a):  $\text{SST forcing} = \text{SST}_{\text{LGM}} - \text{SST}_{\text{PI}} + \text{SST}_{\text{AMIP}}$ . We  
134 set the sea surface  $\delta^{18}\text{O}$  to +1.2‰ and assume no glacial change in the mean deuterium excess in the ocean, as

135 in Risi et al. (2010a). Note once again that the 2-year spin-up time is enough to reach equilibrium. In the VL  
136 present-day “free” simulation, the globally and annually averaged values of temperature and  $\delta^{18}\text{O}$  in  
137 precipitation for the first post spin-up model year are 12.54°C and -8.27‰ respectively, very close to the average  
138 values over the 10 years of simulation (12.54°C and -8.34‰) in comparison to the interannual variabilities of  
139 0.06°C and 0.06‰. The conclusion still holds if we focus on the 60°S-90°S area instead: the average values of  
140 temperature and  $\delta^{18}\text{O}$  in precipitation for the first post spin-up year are -22.03°C and -26.08‰ respectively, very  
141 close to the average values over the 10 years of simulation (-22.04°C and -26.04‰) within the average  
142 interannual variability of 0.10°C and 0.21‰.

143 We express the isotopic composition of difference water bodies in the usual  $\delta$ -notation as the deviation from the  
144 Vienna Standard Mean Ocean Water (V-SMOW). So for  $\text{H}_2^{18}\text{O}$ , the  $\delta^{18}\text{O}$  value is calculated as  $\delta^{18}\text{O} =$   
145  $([\text{H}_2^{18}\text{O}]/[\text{H}_2^{16}\text{O}])_{\text{sample}} / ([\text{H}_2^{18}\text{O}]/[\text{H}_2^{16}\text{O}])_{\text{V-SMOW}} - 1) \times 1000$ . Long-time mean  $\delta$  values are then calculated as  
146 precipitation-weighted mean. For the quantitative model-data comparisons, we retrieve the model values at  
147 data geographical coordinates by bilinear interpolation. Without such an interpolation, i.e. considering the  
148 nearest grid point instead, the root-mean-squared errors (RMSE) of mean  $\delta^{18}\text{O}$  and temperature from the VL  
149 simulation differ only by 0.5‰ and 0.14°C compared to the results presented below, so the uncertainty  
150 associated with the model-data co-location is small.

## 151 **2.2 Observations**

152 For analyzing the model performance over Antarctica under present-day conditions, we make use of the  
153 observational isotope database compiled by Masson-Delmotte et al. (2008). We also focus especially on the East-  
154 Antarctic plateau (defined by the black bold contour of 2500 m above sea level elevation in Figure 1) because  
155 this area provides the main reconstructions of past climate based on the interpretation of water stable isotope  
156 records. To compare the model-data agreement of our simulations with van Leer and upstream advection  
157 schemes for atmospheric boundary layer, inversion temperatures, we use additional datasets for the EPICA  
158 Dome C station (EDC: [75.10° S; 123.35° E]): the surface temperature, the 10m-temperature and the downward  
159 longwave radiative flux at the surface ( $\text{LW}_{\text{dn}}$ ) over the period 2011-2018 thanks to the CALVA program and the  
160 Baseline Surface Radiation Network (BRSN) (Vignon et al., 2018 and references therein); and the observed  
161 precipitable water from radiosoundings data for the period 2010-2017. We are aware that the observations  
162 periods are not the same than our model period (1990-2008), giving a possible bias in the model-data  
163 comparison. For the cloud cover, we use the CALIPSO-GOCCP observations (Chepfer et al., 2010) over the period  
164 2007-2008. We compare these data with the high-level cloud fraction over EDC simulated by LMDZ-iso through  
165 the COSP package (Bodas-Salcedo et al., 2011), which allows to simulate what the CALIOP instrument on-board  
166 CALIPSO would measure if it was into orbit above the simulated atmosphere. The difference in  $\text{LW}_{\text{dn}}$  between  
167 the simulations VL and UP\_xy at EDC is also present all over the East Antarctic plateau, confirming that EDC is  
168 representative of this area.

## 169 **3 Results and discussion**

### 170 **3.1 Model-data comparison for present-day conditions**

#### 171 **3.1.1 Water stable isotopes**

172 Figure 1 shows the observed annual mean  $\delta^{18}\text{O}$  values in the snow surface in Antarctica compiled by Masson-  
173 Delmotte et al. (2008) (Figure 1a) and the difference with the modeled annual  $\delta^{18}\text{O}$  in precipitation from the  
174 UP\_z (Figure 1b), UP\_xy (Figure 1c) and VL (Figure 1d) simulations. The spatial average over the 60°S – 90°S area  
175 of standard deviations of long-time  $\delta^{18}\text{O}$  and  $\delta\text{D}$  model values is of 0.97 and 7.48 ‰, respectively. The results  
176 from the VL simulation are in better agreement with the  $\delta^{18}\text{O}$  observations over Antarctica (Figure 1d). This is  
177 confirmed by the smaller root-mean-squared error of modeled  $\delta^{18}\text{O}$  in precipitation from the VL simulation,  
178 calculated as the difference between the observed annual mean values and the LMDZ-iso results (RMSE = 4.47‰,  
179 i.e. 12.2% of the observed mean Antarctic  $\delta^{18}\text{O}$  value). The results from the VL simulation for the other isotopic  
180 variable  $\delta\text{D}$  is also the closest of the observations with a RMSE of 40.93‰ (Table 1, red background). Our  
181 simulated  $\delta^{18}\text{O}$  in precipitation is very sensitive to the choice of the advection scheme on the horizontal plane,  
182 with more enriched values when a more diffusive advection scheme is applied (Figure 1c). This is reflected by the  
183 mean  $\delta^{18}\text{O}$  model value and RMSE, higher by 3.16 and 4.31 ‰ than the VL simulation values. On the contrary,  
184 the sensitivity of Antarctic  $\delta^{18}\text{O}$  values to the diffusive properties of vertical advection is weak (Figure 1b) with  
185 an RMSE very close of the VL one (4.84 ‰). The results from the VL simulation for the other isotopic variable  $\delta\text{D}$   
186 is also the closest of the observations with a RMSE of 40.93‰ (Table 1, red background). According to the  
187 observations, the East-Antarctic plateau is where the water isotope values are the lowest (mean  $\delta^{18}\text{O}$  below -  
188 40‰, Figure 1a) due to the very low temperatures. Because of the extreme cold and dry conditions at this area,  
189 one can see that the main disagreements between model outputs and observations are located at this place  
190 (Figure 1 and blue background of Table 1). Again, the isotopic outputs from the VL simulation are in better  
191 agreement with the observations (Table 1, blue background). These first results confirm that an excessively  
192 diffusive water vapor transport influences significantly the simulated isotopic and temperature values over  
193 Antarctica.

#### 194 **3.1.2 Temperature**

195 The bias in temperature is deteriorated about in the same way when applying a more diffusive advection on the  
196 vertical direction or on the horizontal plane, as shown with the RMSE of annual mean temperature of 7.50, 7.31  
197 and 6.60°C for the UP\_z, UP\_xy and VL simulations respectively (Table 1, red background). This tendency is the  
198 same when focusing on the East-Antarctic plateau. However, in average over the East Antarctic plateau, the  
199 temperatures of -30.51, -30.69 and -31.54°C from the UP\_z, UP\_xy and VL simulations are all within the spatial  
200 average of standard deviations of long-time temperature values (0.87°C). These values are all much warmer than  
201 the average observed temperature (-36.93°C). This shows that other factors than advection are responsible for  
202 the warm biases.

203 It has been suggested that the Antarctic warm bias in AGCMs could be linked to the general poor representation  
204 of the polar atmospheric boundary layer and related atmospheric inversion temperatures in these models  
205 (Krinner et al., 1997). Cesana and Chepfer (2012) have also shown that CMIP5 models generate too many high-  
206 level clouds (i.e. above an altitude of 6.72 km), that can partly explain the overestimation of temperatures in  
207 Antarctica, due to their effect on downwelling longwave radiation. To go further, we compare the seasonal  
208 signals of surface temperature, near-surface thermal inversion (defined here as the difference between the 10m-  
209 temperature and the surface temperature), downward longwave radiative flux at the surface ( $LW_{dn}$ ), integrated  
210 water vapor column and high cloud fraction at EPICA Dome C (EDC: [75.10° S; 123.35° E]) from VL and UP\_xy  
211 simulations with meteorological observations (see Figure 2 and the description of the data in section 2.2).

212 Both VL and UP\_xy simulations have a warm bias at the surface (Figure 2a) and underestimate the near-surface  
213 inversion (Figure 2b), especially during the austral winter. This is mostly explained by an overly active turbulent  
214 mixing within stable boundary layers in LMDZ5 (Vignon et al., 2017). The disagreement with the observed near-  
215 surface inversion at EDC is exacerbated in the UP\_xy simulation because the modeled surface temperature is  
216 more affected by the change of advection scheme than the modeled 10m-temperature. This is consistent with a  
217 higher total  $LW_{dn}$  flux over EDC in the UP\_xy simulation than in the VL one (Figure 2c), by around  $5 \text{ W.m}^{-2}$  in  
218 summer and  $8 \text{ W.m}^{-2}$  in austral winter. The difference of  $LW_{dn}$  between our two simulations can be due to two  
219 aspects: the fraction of high cloud and the amount of water vapor (e.g., Vignon et al., 2018). The modeled high-  
220 level cloud fraction over EDC for the period 2007-2008 is overestimated compared to the CALIPSO-GOCCP  
221 observations (Figure 2f). This finding is consistent with the results from Cesana and Chepfer (2012) for CMIP5  
222 models and with Lacour et al. (2018) for dry areas over Greenland. This result also concurs with an overestimated  
223 relative humidity with respect to ice in the troposphere compared to radiosoundings (not shown) during winter  
224 seasons, especially in the UP\_xy simulation. The disagreement of high cloud cover is clearly enhanced in the  
225 UP\_xy case, which is consistent with stronger  $LW_{dn}$  in the UP\_xy simulation. The comparison of  $LW_{dn}$  under clear  
226 sky in our two simulations (Figure 2d) can give us information about the contribution of cloud and water vapor  
227 on the variations of total  $LW_{dn}$  (Figure 2c). The stronger simulated  $LW_{dn}$  in the UP\_xy simulation is due mainly to  
228 the high cloud in winter (70 %) and to both high cloud and water vapor during summer (42.5 and 57.5 %,   
229 respectively). The effect of water vapor on downward longwave radiative flux is also confirmed by the higher  
230 amount of precipitable water over EDC in the UP\_xy simulation (Figure 2e). From these results, we deduce that  
231 the warmer surface temperature in the UP\_xy simulation is due to the higher air temperature (direct effect of  
232 the horizontal upstream advection) and to the radiative amplification from high clouds and in a lesser extent  
233 from water vapor (indirect effect of the horizontal upstream advection).

234 The near-surface warm bias in LMDZ-iso, which is most pronounced for the coldest temperatures (see Figure 3),  
235 has the consequence that the distillation is not strong enough. Some microphysical processes and kinetic  
236 fractionation at very low temperature can be missed too. These different aspects could contribute to an  
237 overestimation of the  $\delta^{18}\text{O}$  and  $\delta\text{D}$  in precipitation over Antarctica. Finally, the 20CR reanalysis assimilate only  
238 surface observations of air pressure and use the observed monthly sea surface temperature and sea ice  
239 concentration as lower boundary conditions. These less strong constraints, compared to other reanalyses, may



240 cause biases on the surface temperature over the poles (A. Orsi, personal communication), that can impact our  
241 isotopic delta values.

### 242 **3.1.3 Spatial $\delta^{18}\text{O}$ –temperature relationship**

243 We compare now our simulated spatial  $\delta^{18}\text{O}$ –temperature relationship and  $\delta^{18}\text{O}$  values for a given temperature  
244 to the ones from the data compiled by Masson-Delmotte et al. (2008). Over the full temperature range, the  
245 spatial gradient is  $0.83 \text{‰} \cdot \text{°C}^{-1}$  in the VL simulation, very close of the observed one ( $0.80 \text{‰} \cdot \text{°C}^{-1}$ ). We make the  
246 same comparison but by restricting the dataset to the ones on the East-Antarctic plateau (Figure 3). As noticed  
247 previously, the average modeled temperature over Antarctica is overestimated whatever the simulation  
248 considered. Especially, no simulated temperature reaches a value below  $-50\text{°C}$ . Yet, simulated  $\delta^{18}\text{O}$  values are  
249 too depleted for the temperature range between  $-49\text{°C}$  and  $-43.5\text{°C}$ . As a consequence, a steeper modeled  $\delta^{18}\text{O}$ –  
250 temperature gradient is observed for the lowest temperatures, giving a modeled global gradient of  $1.24 \text{‰} \cdot \text{°C}^{-1}$   
251 (thin orange line). If we restrict the fit to simulated temperatures higher than  $-43.5\text{°C}$  (vertical dashed line in  
252 Figure 3), corresponding to the change of slope in the simulated  $\delta^{18}\text{O}$ –temperature relationship (thick orange  
253 line), the simulated gradient ( $0.96 \text{‰} \cdot \text{°C}^{-1}$ ) is in more reasonable agreement with the one from the observations  
254 ( $0.85 \text{‰} \cdot \text{°C}^{-1}$ ).

255 We now discuss possible reasons that could explain why the simulated  $\delta^{18}\text{O}$ –temperature slope is too steep at  
256 very low temperatures. First, it could be related to missing representation of fractionation during sublimation  
257 from the surface. As for all AGCMs equipped with water isotopes, fractionation at sublimation is not taken into  
258 account in LMDZ-iso. However, this effect would lead to further decrease of the water vapor  $\delta^{18}\text{O}$  in polar region  
259 and hence contribute to an even steeper  $\delta^{18}\text{O}$ –temperature slope at low temperature (hence further accentuate  
260 the mismatch). Second, the slope mismatch could be related to poorly represented kinetic fractionation. As in all  
261 the other models equipped with water isotopes, the parameterization of kinetic effect during vapor-to-solid  
262 condensation is represented empirically using a linear relationship between the supersaturation and the  
263 condensation temperature (Risi et al., 2010a). A modification of the temperature can thus induce some change  
264 in the  $\delta^{18}\text{O}$  of the condensate, but this effect is of second order compared to the distillation effect explaining  
265 much of the slope between  $\delta^{18}\text{O}$  and surface temperature. Third, the slope mismatch could be related to a poor  
266 representation of the atmospheric boundary layer and of its related inversion temperature (Krinner et al., 1997;  
267 Masson-Delmotte et al., 2006), as shown in the Figure 2. In LMDZ-iso, the warm bias in the simulated  
268 condensation temperature is smaller than that in the simulated surface temperature. Therefore, the water vapor  
269 masses continue to be distilled when moving away from the coast, while the cooling simulated at the surface  
270 from the coast to the remote region of the East Antarctic plateau is much less steep than in the reality.

## 271 **3.2 Comparison of the different simulations**

### 272 **3.2.1 Effects of the diffusive properties of the advection scheme**

273 We compare here the results from our different present-day simulations at a R96 grid resolution. The UP\_z  
274 simulation (upstream vertical advection, Figure 1b) increases the bias a little in  $\delta^{18}\text{O}$ , but its results stay relatively

275 close of the  $\delta^{18}\text{O}$  values from the VL simulation, indicated by the similar average values that differ only by 0.89‰  
276 for all Antarctica (Table 1) that is smaller than the mean of the 60° S – 90° S standard deviations. On the other  
277 hand, the  $\delta^{18}\text{O}$  outputs from the UP\_xy simulation (upstream horizontal advection, Figure 1c) display greater  
278 differences with the VL simulation ones, and so with the isotopic data, as revealed by the mean UP\_xy - VL  
279 difference in  $\delta^{18}\text{O}$  of 4.31‰. This is even more significant when focusing on the East-Antarctic plateau, with a  
280 model-data difference in  $\delta^{18}\text{O}$  reaching 20‰ at some locations. The annual mean  $\delta^{18}\text{O}$  and  $\delta\text{D}$  values from the  
281 UP\_xy simulation are increased by 6.39‰ and 44.26‰ compared to the VL simulation average values, in less  
282 agreement with the observations as shown by their respective RMSE values (Table 1, blue background). It shows  
283 that the diffusive property of the advection scheme on the horizontal plane is essential to better model the water  
284 isotope distribution, especially over Antarctica. To go further, one can also compare the  $\delta^{18}\text{O}$  values at a fixed  
285 temperature for the UP\_xy and VL simulations (Figure 3, red and orange crosses respectively). The  $\delta^{18}\text{O}$  in  
286 precipitation for a temperature of -32°C over the East-Antarctic plateau is already smaller by 6.1‰ in the VL  
287 simulation. This very significant difference in initial  $\delta^{18}\text{O}$  can be attributed to the proportion of mixing against  
288 distillation that affects the water vapor during its transport. This lends support to our hypothesis that too much  
289 diffusive mixing along the poleward transport leads to overestimated  $\delta^{18}\text{O}$  because dehydration by mixing  
290 follows a more enriched path than dehydration by Rayleigh distillation (Hendricks et al., 2000; Galewsky and  
291 Hurley, 2010). We expect the relative contribution of mixing vs. distillation to have the largest impact on  $\delta^{18}\text{O}$  at  
292 latitudes where eddies are the most active. This is probably why the  $\delta^{18}\text{O}$  difference between the VL and UP\_xy  
293 simulation becomes large in mid-latitudes, over the austral ocean before arriving at the Antarctica coast (not  
294 shown), hence the difference in “initial”  $\delta^{18}\text{O}$  in Figure 3.

295 As noticed in section 3.1.2, all our simulations overestimate the average temperature in Antarctica and even  
296 more on the East-Antarctic plateau. A more diffusive advection on the horizontal or on the vertical increases the  
297 mean temperature value by 0.85 and 1.03°C respectively compared to the VL result. To explain such an influence  
298 of the advection on the temperature over Antarctica, even secondary, one can hypothesize that the Antarctic  
299 continent is better isolated, and so colder, when the advection of the model is less diffusive. If we focus now on  
300 the link between the temperature and the  $\delta^{18}\text{O}$  in precipitation over all the continent, the  $\delta^{18}\text{O}$ -temperature  
301 gradients according to our different R96 simulations UP\_z, UP\_xy and VL are at 0.79, 0.69 and 0.83 ‰.°C<sup>-1</sup>  
302 respectively. The difference between the VL and UP\_xy gradient shows an effect of diffusive properties of the  
303 large-scale transport on the distillation process. This difference between the modeled  $\delta^{18}\text{O}$ -temperature  
304 gradients is reduced if we restrict to the temperature range above -43.5°C over the East-Antarctic plateau, with  
305 gradients of 0.93 and 0.96 ‰.°C<sup>-1</sup> according to the UP\_xy and VL simulations respectively (Figure 3, red and  
306 orange thick lines).

307 Since the simulated temperature difference between the UP\_xy and VL simulations is in the error margin, i.e.  
308 less than 1°C, we do not expect that the temperature difference explains the difference in spatial slopes. Rather,  
309 the larger slope in simulation VL is due to the larger relative contribution of Rayleigh distillation compared to  
310 mixing.

### 311 3.2.2 Effects of the horizontal grid resolution

312 We test now the hypothesis that to increase the horizontal resolution is equivalent to using an advection scheme  
313 that is less diffusive. The Antarctica-mean results are summarized in the Table 2. Compared to the UP\_xy R96  
314 simulation, the average value of  $\delta^{18}\text{O}$  in precipitation is decreased by 4.31‰ when the advection scheme is  
315 improved (VL R96), and by 2.42‰ when the horizontal resolution is increased (UP\_xy R144), in better agreement  
316 with the observations. The picture is the same for the  $\delta\text{D}$  outputs. The decrease of the mean modeled  
317 temperature values, smaller than the mean of the long-time standard deviations on the 60° S – 90° S area, is the  
318 same by changing the advection scheme or by increasing the resolution: by 0.85°C and 0.76°C respectively. The  
319 best results are reached by improving both the advection scheme and the horizontal resolution at the same time,  
320 with model-data differences in temperature and  $\delta^{18}\text{O}$  of 4.83°C and 0.88‰ respectively. This confirms that an  
321 increase of the horizontal resolution plays the same role as an improvement in the representation of the  
322 advection scheme on its horizontal plane. It is worth mentioning that the improvement in model-data agreement  
323 using a higher horizontal grid resolution is probably not only due to an improved representation of the advection  
324 in the model but, among others, to a better resolved Antarctic topography and near surface circulation. Our  
325 results are consistent with the study of Werner et al. (2011) that shows, for an increased horizontal resolution,  
326 a better agreement of the simulated isotopic delta values and of the water isotope-temperature gradient with  
327 the observations. The seasonality in modeled Antarctic temperature and precipitation, which influences the  
328 precipitation-weighted isotope values, is not altered by the change of advection scheme or horizontal grid  
329 resolution.

### 330 3.2.3 Effects on the LGM to present-day change in temperature

331 We evaluate now the effects of the diffusive properties of the advection scheme on the  $\delta^{18}\text{O}$  – temperature  
332 temporal slope at different Antarctic locations between LGM and present-day. For this, we compare the VL and  
333 UP\_xy LGM simulations with their present-day “free” counterparts (i.e. without nudging). Figure 4a shows VL  
334 simulated temporal slopes in  $\delta^{18}\text{O}$  – temperature at each location as  $\Delta\delta^{18}\text{O}/\Delta T$ . Over East Antarctica, the  
335 simulated temporal slope becomes larger inland, going from near to 0 ‰.°C<sup>-1</sup> on the coast to 2 ‰.°C<sup>-1</sup> on the  
336 deep East-Antarctic plateau. Over West Antarctica, the temporal slopes are more heterogeneous and show  
337 variations from 0.1 to 1 ‰.°C<sup>-1</sup>. The slopes at WDC (Wais Divide Core, square symbol), Siple Dome (triangle  
338 symbol), Vostok (circle symbol) and EDC (cross symbol) are of 0.80, 0.66, 0.73 and 0.37 ‰.°C<sup>-1</sup> respectively  
339 (calculated by considering the 9 grid cells centered on each drill location). Figure 4b shows the differences  
340 between the UP\_xy and VL simulated slopes. As the spatial slopes from the VL simulation are higher than the  
341 UP\_xy one, it is expected that the LGM to present-day temporal slopes are larger in the VL simulation than in the  
342 UP\_xy one because of the colder temperatures and of the higher relative contribution of the Rayleigh distillation.  
343 This is the case on the western part of the continent where the UP\_xy temporal slopes are smaller than the VL  
344 ones by up to 0.6 ‰.°C<sup>-1</sup>. On the East-Antarctic plateau, the VL  $\delta^{18}\text{O}$  – temperature slopes are larger between the  
345 longitudes 100°E and 140°E than in UP\_xy as well. The slope values are smaller in VL than in UP\_xy only over the  
346 center of the continent, which corresponds to the lowest simulated temperature values and so where the

347 gradient between the  $\delta^{18}\text{O}$  and temperature values is much steeper (see 2<sup>nd</sup> paragraph of section 3.1.3),  
348 especially for LGM conditions.

349 We highlight now the consequences that an excessively diffusive horizontal advection may have on the LGM  
350 cooling reconstructed from simulated temporal slopes. For this, we calculated the difference in the LGM to  
351 present-day change in temperature deduced from the observed changes in  $\delta^{18}\text{O}$  and simulated slopes ( $\Delta\delta^{18}\text{O}/\Delta T$ )  
352 by VL and UP\_xy, at WDC and Siple Dome stations for West Antarctica, and Vostok and EDC stations for East  
353 Antarctica (Table 3). For WDC, Siple Dome and Vostok stations, the average changes in  $\delta^{18}\text{O}$  between LGM and  
354 present-day are larger in the VL simulation, -11.98, -9.88 and -3.90 ‰ respectively, than in the UP\_xy simulation  
355 (-7.52, -6.24 and -2.07 ‰ respectively). For the EDC station, the difference in LGM to present-day change in  
356  $\delta^{18}\text{O}$  between our VL and UP\_xy simulations can be considered in the margin error (-1.97 and -1.70 ‰ for VL and  
357 UP\_xy simulations respectively). The values deduced from the VL simulation are also in better agreement with  
358 the observations (Table 3). As noticed previously, the temporal  $\delta^{18}\text{O}$ -temperature gradients at WDC, Siple Dome,  
359 Vostok and EDC are also larger in the VL simulation by 57%, 65%, 92% and 26% respectively, compared to the  
360 values deduced from the UP\_xy simulation. As a consequence, if one applies the slope from the UP\_xy simulation  
361 instead of the VL one, we would overestimate the present-day-to-LGM cooling at these stations by 4.34, 5.83,  
362 3.76 and 2.95°C respectively. Even if the change in the temporal  $\delta^{18}\text{O}$ -temperature gradient at EDC is relatively  
363 smaller than for the other stations, the consequence on the reconstructed present-day-to-LGM cooling value is  
364 still important. These results show that a more diffusive advection scheme on the horizontal plane can affect  
365 greatly the LGM to present-day temperature change deduced from observed  $\delta^{18}\text{O}$  in precipitation.

#### 366 **4 Conclusions**

367 We have tested with LMDZ-iso if the warm and isotopically enriched biases in Antarctica, frequently observed in  
368 the AGCMs, are associated with the diffusive property of the advection scheme. The simulated water isotope  
369 contents in Antarctica are very sensitive to the diffusive character of the water vapor transport on the horizontal  
370 plane. The higher the contribution of mixing (i.e. diffusion), the more enriched the precipitation. These findings  
371 are even more striking for the East-Antarctic plateau where the main ice cores allowing paleoclimate  
372 reconstructions are located. Moreover, because the diffusive character of the large-scale transport influences  
373 the temperature in this region, even in a light way, this has an impact on the modeled water isotopic composition  
374 through the Rayleigh distillation. So, we conclude here that the excessive numerical diffusion has a large  
375 influence on the enriched isotopic bias. For the spatial isotope-temperature relationship over the East-Antarctic  
376 plateau observed in LMDZ-iso, this latter is improved for the temperatures above -43.5°C, in more reasonable  
377 agreement with the observations. At the lowest temperatures (i.e still over the East-Antarctic plateau), that the  
378 model is not able to reach, the non-linearity observed in our simulations (the spatial  $\delta^{18}\text{O}$  – temperature  
379 relationship is steeper for the lowest temperatures) can be unlikely explained at first order to missing or poorly  
380 represented kinetic fractionation. One can speculate that the water masses continue to be distilled when  
381 moving away from the coast, hence depleting the water vapor in heavy isotopes while the modeled temperature  
382 decrease from the coast to the remote region of the East Antarctic plateau is much less steep than in the reality.

383 This more pronounced effect in the case of a more diffusive horizontal advection scheme could be due to the  
384 deteriorated representation of the inversion temperature. The temporal isotope–temperature relationship at  
385 some locations in Antarctica can be influenced by the diffusive properties of the advection scheme on its  
386 horizontal domain. As for the spatial gradient, an excessive numerical diffusion has the consequence to decrease  
387 the isotope-temperature temporal gradient, leading to a wrong estimation of the LGM to present-day  
388 temperature change deduced from observed  $\delta^{18}\text{O}$ . Our study demonstrates that a representation of the  
389 advection scheme in the AGCMs taking into account water isotopes and isotopic gradients, especially on the  
390 horizontal domain, is an important step toward a more realistic modeling of water isotopes over Antarctica.  
391 Another way to improve this aspect is to increase the spatial resolution, which has the same effect as applying a  
392 less diffusive advection scheme on the water isotopic composition and the temperature. This study shows again  
393 the importance of using water stable isotopes in GCMs for the evaluation and quantification of the processes  
394 influencing the hydrological cycle, including the advection of water vapor. We expect our main results (excessive  
395 diffusive advection leading to warmer temperatures, moister boundary layer and more enriched water vapor) to  
396 be robust and to hold in other models as well. However, the quantitative response to the advection scheme may  
397 be modulated by the representation of boundary layer processes and high-cloud microphysics in each model.

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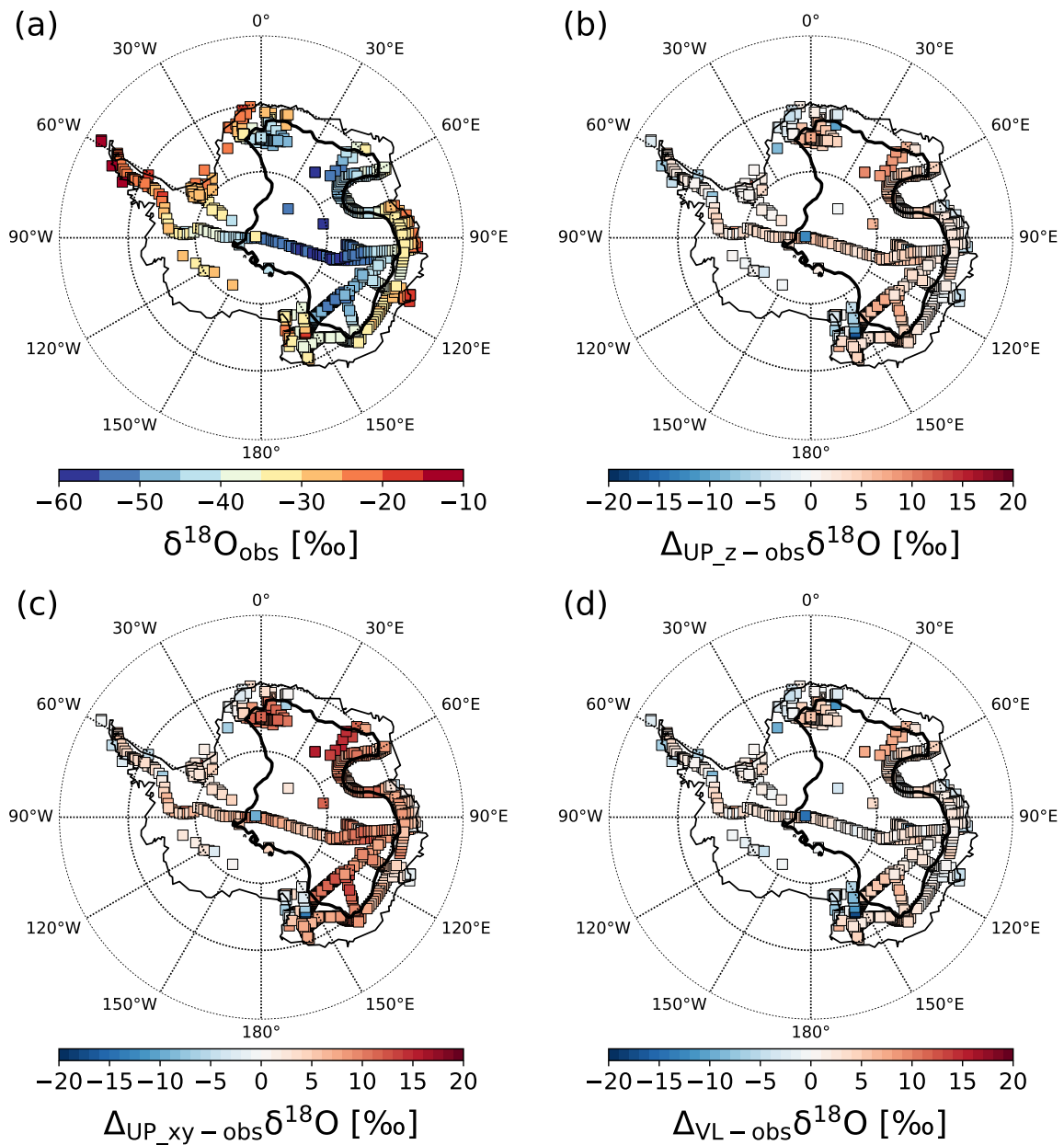
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566 **Figure 1: Map of Antarctica showing (a) the observed  $\delta^{18}\text{O}$  values from the compilation by Masson-Delmotte et al. (2008),**  
 567 **(b) the difference between the simulated  $\delta^{18}\text{O}$  in precipitation and the  $\delta^{18}\text{O}$  observations for the UP\_z, (c) UP\_xy and (d)**  
 568 **VL simulations. The bold black line shows the contour of 2500 m above sea level elevation.**

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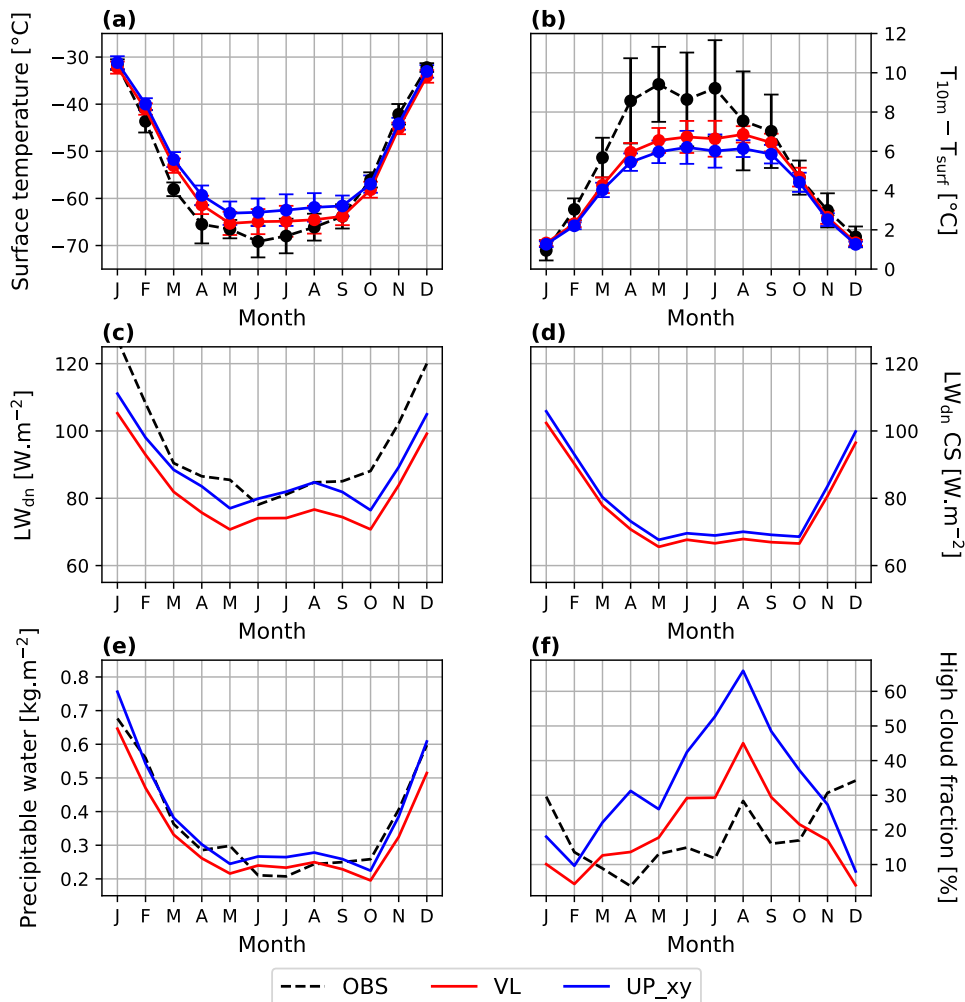
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	Mean data	Mean UP_z	RMSE UP_z	Mean UP_xy	RMSE UP_xy	Mean VL	RMSE VL
T (°C)	-36.93	-30.51	7.50	-30.69	7.31	-31.54	6.60
$\delta^{18}\text{O}$ (‰)	-36.76	-34.85	4.84	-31.43	7.63	-35.74	4.47
$\delta\text{D}$ (‰)	-289.62	-272.28	43.76	-251.34	62.00	-279.49	40.93
T (°C)	-47.46	-39.49	8.38	-39.88	7.99	-40.71	7.23
$\delta^{18}\text{O}$ (‰)	-46.77	-42.27	5.03	-37.37	9.69	-43.76	3.80
$\delta\text{D}$ (‰)	-366.98	-325.37	43.79	-291.99	76.44	-336.25	33.58

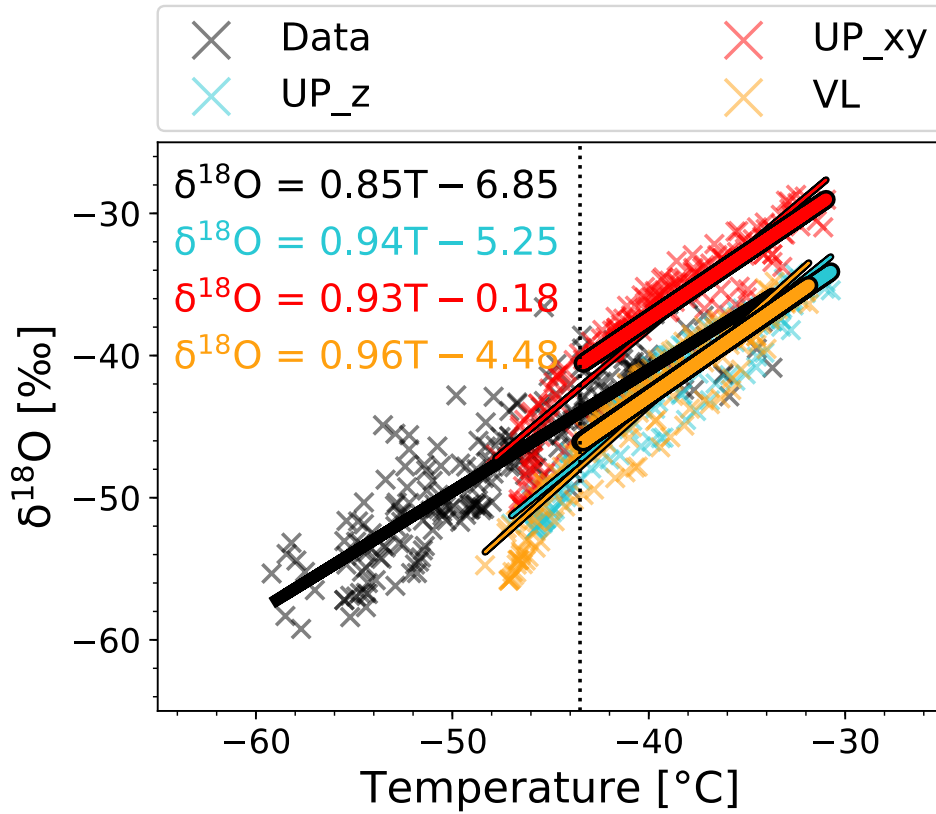
574 Table 1: Observed and simulated annual mean values of temperature (T),  $\delta^{18}\text{O}$  and  $\delta\text{D}$  for the full Antarctic dataset (red  
575 background) and restricted to the East-Antarctic plateau (blue background), and the corresponding RMSE.

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577

578 Figure 2: Multi-year monthly mean variations of (a) surface temperature, (b) near-surface thermal inversion (defined as  
579 the difference between the 10m-temperature and the surface temperature), (c and d) total and clear-sky component of  
580 downward longwave radiative flux at the surface, (e) precipitable water and (f) high cloud fraction above EPICA Dome C.  
581 The red and blue curves correspond to the VL and UP\_xy simulation results, respectively. A comparison with observations  
582 (see section 2.2) is made when possible (dashed black lines).



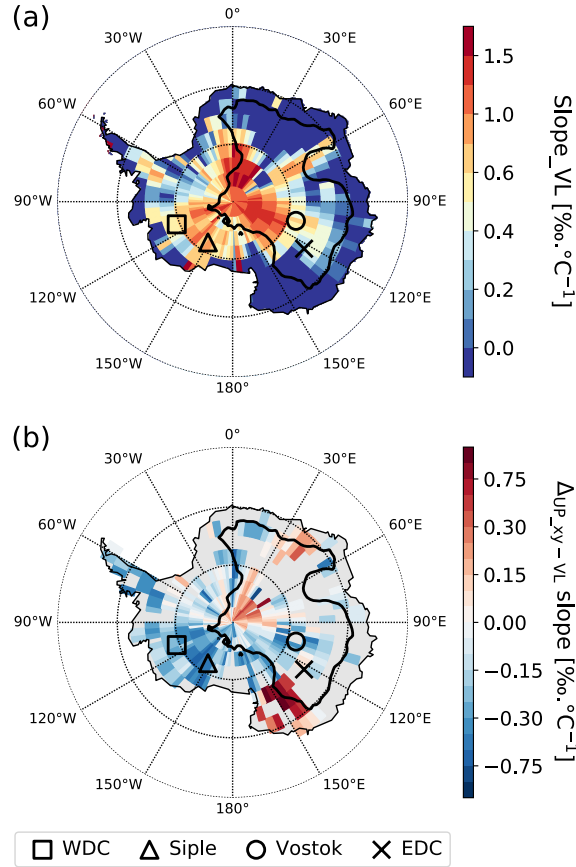
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584 Figure 3: Relationship between  $\delta^{18}\text{O}$  and temperature on the East-Antarctic plateau according to the observations (black)  
 585 and the UP\_z (blue), UP\_xy (red) and VL (orange) simulations. For each simulation outputs, two linear regressions have  
 586 been conducted: one on the full East-Antarctic plateau dataset (thin lines) and one on the same dataset without the  
 587 temperatures below  $-43.5^\circ\text{C}$  (bold lines) indicated by the vertical dashed line. The corresponding formulas of these latter  
 588 are also shown.

	Mean data	Mean UP_xy R96	Mean VL R96	Mean UP_xy R144	Mean VL R144
T ( $^\circ\text{C}$ )	-36.93	-30.69	-31.54	-31.45	-32.10
$\delta^{18}\text{O}$ (‰)	-36.76	-31.43	-35.74	-33.85	-37.64
$\delta\text{D}$ (‰)	-289.62	-251.34	-279.49	-267.17	-289.55

589 Table 2: Comparison of the observed annual mean values of temperature (T),  $\delta^{18}\text{O}$  and  $\delta\text{D}$  for the full Antarctic dataset  
 590 with four different LMDZ-iso simulations, combining different horizontal resolutions (R96 and R144) and different  
 591 advection schemes (UP\_xy and VL).

592



593

594 **Figure 4: (a) Temporal slope ( $\Delta\delta^{18}\text{O}/\Delta T$ ) between the present-day (PD) and LGM according to the VL simulations. (b)**  
 595 **Difference with the temporal slope deduced from the UP\_xy simulations. The grey areas indicate where the LGM-PD**  
 596 **difference in  $\delta^{18}\text{O}$  is lower than the standard deviation of the interannual variability. Square, triangle, circle and cross**  
 597 **symbols represent the WDC, Siple Dome, Vostok and EDC sites respectively.**

Site	Latitude	Longitude	LGM-PD $\delta^{18}\text{O}_{\text{obs}}$ (‰)	LGM-PD $\delta^{18}\text{O}_{\text{UP\_xy}}$ (‰)	LGM-PD $\delta^{18}\text{O}_{\text{VL}}$ (‰)	Slope UP_xy (‰·°C <sup>-1</sup> )	Slope VL (‰·°C <sup>-1</sup> )	Difference in reconstructed $\Delta T$ (°C)
WDC	-79.47	-112.08	-9.44	-7.52	-11.98	0.51	0.80	-4.34
Siple Dome	-81.67	-148.82	-9.8	-6.24	-9.88	0.40	0.66	-5.83
Vostok	-78.47	106.87	-5.66	-2.07	-3.90	0.38	0.73	-3.76
EDC	-75.10	123.35	-6.74	-1.70	-1.97	0.31	0.37	-2.95

598 **Table 3: Sites of interest and their geographical coordinates, observed LGM to present-day changes in  $\delta^{18}\text{O}$  (Wais Divide:**  
 599 **WAIS Divide Project Members (2013), Schoenemann et al. (2014); Siple Dome: Brook et al. (2005), Schoenemann et al.**  
 600 **(2014); Vostok: Vimeux et al. (2001), Landais et al. (2008, 2012); EDC: EPICA Community Members (2004), Stenni et al.**  
 601 **(2010)), simulated LGM to present-day changes in  $\delta^{18}\text{O}$  (UP\_xy and VL simulations), simulated temporal  $\delta^{18}\text{O}$ -temperature**  
 602 **slopes and VL-UP\_xy difference in reconstructed temperature change deduced from the observed  $\delta^{18}\text{O}$  changes and the**  
 603 **simulated temporal slopes. This difference is calculated as  $-(slope_{\text{UP\_xy}} - slope_{\text{VL}}) \times \Delta\delta^{18}\text{O}_{\text{obs}} / slope_{\text{VL}}^2$ . Negative**  
 604 **signs indicate that when using the UP\_xy temporal slope, we overestimate the LGM cooling compared to using the VL**  
 605 **temporal slope. The model values are based on the spatial averages over the 9 grid cells surrounding the ice cores**  
 606 **geographical coordinates.**