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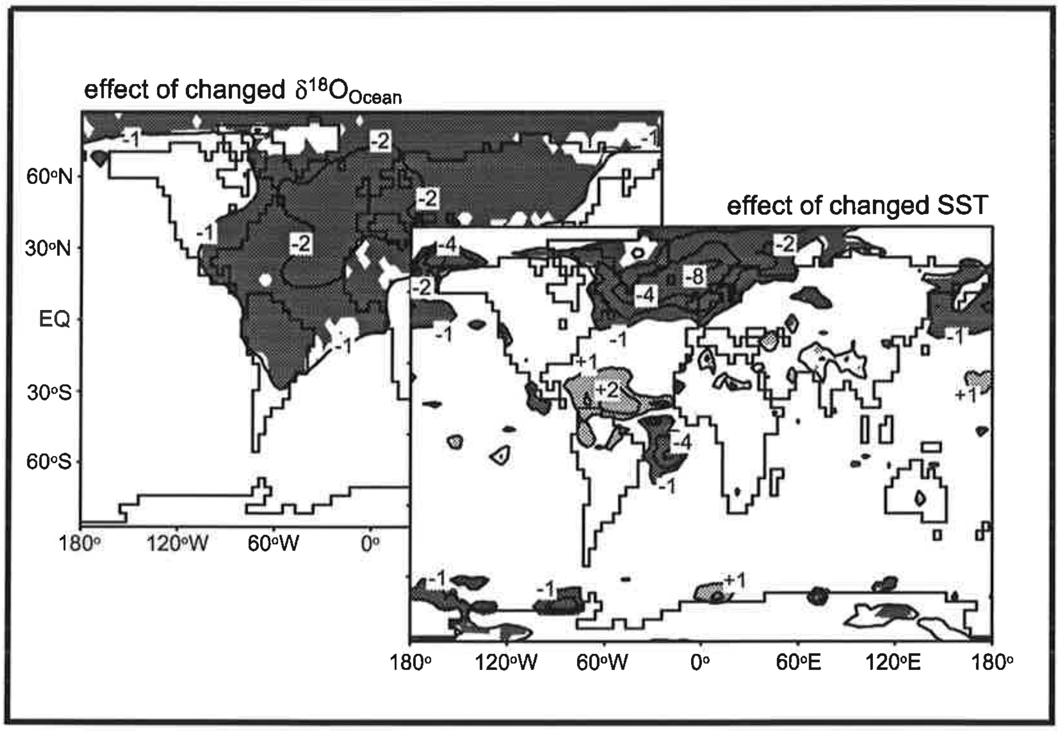

POSSIBLE CHANGES OF δ^{18}O IN PRECIPITATION CAUSED BY A MELTWATER EVENT IN THE NORTH ATLANTIC

by

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Possible Changes of $\delta^{18}O$ in Precipitation Caused by a Meltwater Event in the North Atlantic

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Abstract

The latest version of the Hamburg Atmosphere General Circulation Model ECHAM-4 is used to investigate how a meltwater event in the North Atlantic might alter the signal of stable water isotopes ($H_2{}^{18}O$, HDO) in precipitation. Our results show that such a meltwater event will cause significant changes in the isotopic composition of the precipitation over many parts of the Northern Hemisphere, but also in the tropical Atlantic region. Model simulations suggest that for such a scenario isotope anomalies are not always related to temperature changes, but also to changes in the seasonality of precipitation or the precipitation amount. A changed isotopic composition of evaporating ocean surface waters (caused by a massive meltwater input into the North Atlantic) causes temperature-independent isotope anomalies, too. Changes of the deuterium excess are even more affected by the imposed oceanic isotope anomaly due to the non-linearity of the evaporation process.

Introduction

One of the most puzzling problems of climate research is the question for the cause of the strong and rapid climate changes during the last 70,000 years known as Heinrich events, Dansgaard-Oeschger events and the Younger Dryas. These events can be traced in paleo-records almost from the entire world including the tropics, e.g. in [*Brook et al.*, 1996; *Curry and Oppo*, 1997; *Johnsen et al.*, 1992]. Two main hypotheses for their origin are currently discussed: One is a reduction of northward Atlantic heat transport due to a shutdown / reduction of North Atlantic Deep Water (NADW) formation triggered by strong meltwater / iceberg discharge from the European and / or North American ice shields [*Stocker*, 1998]. The other explanation claims that the cause for this variability lies in the tropics [*Cane*, 1998].

Most of our knowledge about these past climate changes is based on proxy records like e. g. ice cores, pollen records and marine sediment cores. From these proxy records changes in e. g. temperature and precipitation are estimated. The required transfer functions are in general derived from present day spatial variations of proxy (e. g. isotope in ice cores) and physical quantities (e. g. temperature and precipitation). These transfer functions are often nonunique and there is no guarantee that they are also appropriate for temporal variations. However, this approach is widely used to compare estimates of past climate changes with the results of model sensitivity studies. In this paper we will use a different approach. Here we will show a simulation study, which explicitly models the cycling of two stable water isotopes ($H_2{}^{18}O$, HDO) in the hydrological cycle. Focusing on the simulation of a meltwater event into the Labrador Sea we investigate the following questions: (1) In which regions can we detect isotope anomalies in a colder climate forced by a rapid shutdown of the NADW formation? (2) Is a changed isotopic composition in precipitation (usually given as $\delta^{18}O$ resp. δD) always coupled to changed surface temperatures? (3) A massive fresh water input with a strong depletion in heavy isotopes will alter the isotopic ocean surface water composition $\delta^{18}O_{Ocean}$. How much does this effect the $\delta^{18}O$ signal in precipitation? (4) Will changes of the deuterium excess d (defined as $d = \delta D - 8 \cdot \delta^{18}O$) reveal additional information?

Model Experiments

Our results are based on three model experiments using the Hamburg Atmosphere General Circulation Model (AGCM) ECHAM-4 in T30 mode (spatial resolution: 3.75 x 3.75 degrees). Each experiment was run for 10 years in equilibrium state after a spin-up time of one year. In the first experiment (further referred to as control run) both sea surface temperatures (SSTs) and $\delta^{18}O_{Ocean}$ were set to present day values. For the other two experiments we prescribed colder SSTs. But while the $\delta^{18}O_{Ocean}$ values were still set to modern values in the second experiment, we assumed a changed isotopic composition $\delta^{18}O_{Ocean}$ in the third one. A comparison between the second and third experiment will enable us to clearly distinguish between the effects of changed SSTs and additionally changed $\delta^{18}O_{Ocean}$. The prescribed monthly SST fields for all three experiments were derived from simulations with the coupled ocean-atmosphere general circulation model (OAGCM) ECHAM3/LSG [*Voss et al.*, 1998]. The ECHAM3/LSG OAGCM was forced into a colder state by a meltwater spike input into the Labrador Sea with a 500 year long triangle-shaped time history (maximum: 0.625 Sv) [*Schiller et al.*, 1997]. The meltwater input led to a freshening of the North Atlantic surface waters, thus suppressing deep convection and the formation of NADW. As a consequence, the thermohaline circulation of the Atlantic was weakened until the end of the freshwater input. Poleward heat transport in the North Atlantic was strongly reduced, leading to a simulated cooling of almost the entire Northern Hemisphere. We calculated monthly mean SSTs of years 300 to 400 of an OAGCM control run resp. the OAGCM meltwater experiment to use as boundary conditions for the isotope control resp. cold climate experiments (Fig. 1a). For the cold climate scenario with an additional changed $\delta^{18}O_{Ocean}$ field we used results of an GCM experiment with an OGCM coupled to an atmospheric energy balance model Mikolajewicz [*Mikolajewicz*, 1996] which included the $H_2{}^{18}O$ composition of ocean water. This model was forced with a meltwater spike, identical to the one of in the OAGCM experiment mentioned above. Although not identical, mean SST changes between the control and cold climate state of this OGCM experiment were similar to the coupled OAGCM simulations. We used the mean of the $\delta^{18}O_{Ocean}$ changes of years 300 to 400 of the OGCM experiment as a prescribed boundary condition for our third isotope experiment (Figure 1b). The deuterium excess of ocean surface water was set to zero in all our experiments. Such a d-excess value is valid for recent climate conditions but might be slightly higher after a meltwater event. Meltwater from the Laurentian ice shield was probably enriched in the deuterium excess. Measurements on the Dye3 core, Greenland, show d-excess values between 4‰ and 8 ‰ for different climate stages [*Johnsen et al.*, 1989]. Except SSTs and $\delta^{18}O_{Ocean}$ we did not change any other boundary condition such as topography, ice shield distribution or solar insolation, which were all set to present day values in the experiments.

Figure 1:

a:

b:

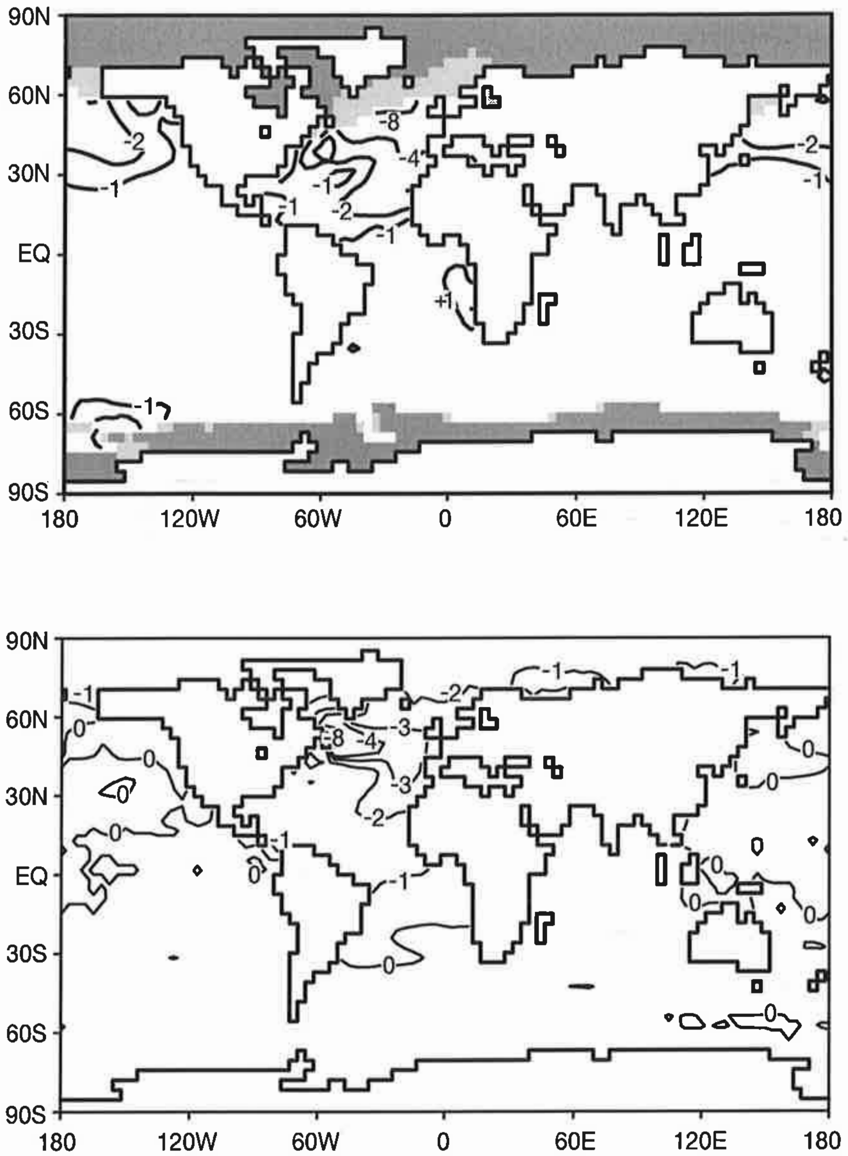

Fig. 1: Differences of the boundary conditions between the control and cold climate experiments: (a) mean anomalies of sea surface temperatures SSTs and sea ice (temperature contour lines at -1, -2, -4, -8, -16 °C; dark gray area: sea ice prescribed in both climates at least half of the year, light gray area: additional sea ice prescribed at least half of the year for the cold climate), (b) anomalies of isotope composition of the ocean surface water $\delta^{18}O_{Ocean}$ (contour lines at 0, -1, -2, -3, -4, -8‰)

We are fully aware that this setup does not represent a realistic simulation of the conditions occurring during a rapid climate change event. Nevertheless the experiments allow an assessment of the first order effects in the isotopic composition of precipitation after a meltwater induced rapid Northern hemisphere cooling event, such as the Younger Dryas. Since most other boundary conditions will probably have remained fairly constant during a rapid climate change, we believe that analyzing the anomalies between the different model experiments can reveal important information.

Results

The mean changes of $\delta^{18}O$ in precipitation between the control climate and the cold climate are shown in Figures 2a-2c. For further analyses we have split the $\delta^{18}O$ anomalies in two parts: (1) the $\delta^{18}O$ changes caused by colder SSTs alone (Fig. 2a), (2) additional $\delta^{18}O$ changes in precipitation for the colder climate caused by the assumed change in $\delta^{18}O_{Ocean}$ (Fig. 2b). Colder SSTs alone affect the $\delta^{18}O$ signal over the Atlantic region, Scandinavia and the western part of Europe (Fig. 2a). The strongest isotope depletion (-8‰) can be observed over the Northern Atlantic in the area of the Norwegian Sea, slightly east of the area of maximum cooling.. Another minimum of isotope values is located over the Northern Pacific region centered at the Bering Street area associated with the prescribed cooling and the increased sea ice cover. This Pacific signal is much weaker than the Atlantic correspondent but still shows a decrease of -4‰. A dipole-like pattern of isotope changes is found in the tropical Atlantic region between 30°N and 30°S. The positive branch (+2‰) is found north of the equator. It is mainly located over the ocean but extends into the northern part of South America. The negative branch (-4‰) is seen over the Atlantic ocean south of the equator. The additional anomalies of $\delta^{18}O$ in precipitation induced by changed $\delta^{18}O_{Ocean}$ values of the Atlantic (Fig. 2b) are very similar to the $\delta^{18}O_{Ocean}$ input field (Figure 1b). Although the extreme ocean water depletion of -8‰ at the coast of Labrador is not reflected in the precipitation signal, the -2‰ and -1‰ contour lines between forcing ($\delta^{18}O_{Ocean}$) and response ($\delta^{18}O$ in prec.) are almost identical, indicating a strong local control of the response signal. Over land surfaces, strongest depletion of -2‰ is found over Western Europe and the Mediterranean region. Weaker anomalies of -1‰ to -2‰ are also found above Eastern Europe and Siberia, Southern Greenland, the East Coast of North America, West Africa and almost half of the South American continent. Combining both effects of changed SSTs and changed $\delta^{18}O_{Ocean}$ (Fig. 2c) results in an increased change in $\delta^{18}O$ over Central Europe and Siberia. The modeled $\delta^{18}O$ changes over the Summit region of the Greenland ice sheet are also slightly larger.

Figure 2:

a:

b:

c:

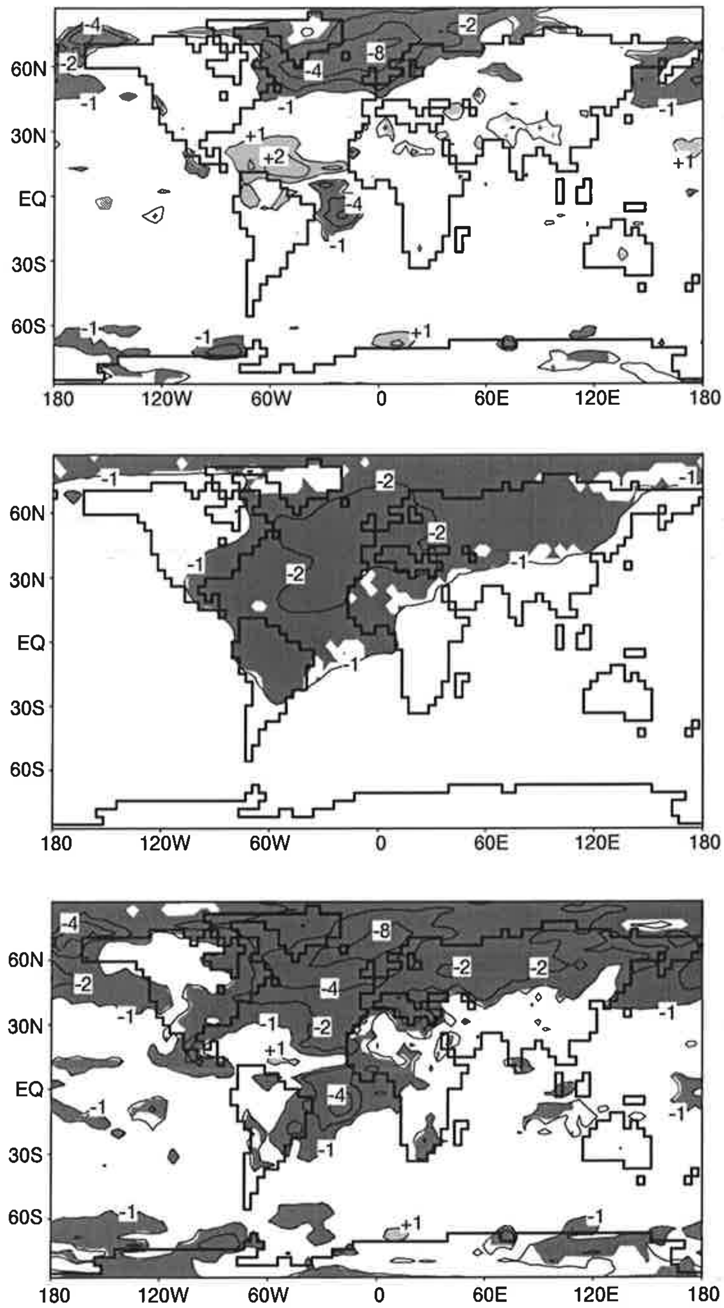

Fig. 2: Changes of the isotopic composition of precipitation between the ctrl and cold climate experiments: (a) $\delta^{18}O$ anomalies in precipitation if only SSTs are changed. (b) Changes of the $\delta^{18}O$ values in precipitation for the cold climate simulation, caused by the additional change of the isotopic composition of ocean surface water $\delta^{18}O_{Ocean}$. (c) $\delta^{18}O$ anomalies in precipitation if both SSTs and $\delta^{18}O_{Ocean}$ are changed for the cold climate simulation. Contour lines in all three plots at ±1, ±2, ±4, ±8‰. Significant $\delta^{18}O$ changes of the long-time mean value compared to the year-by-year variations (two-sided u-test, 95% level) are shaded in light resp. dark gray.

Figure 3:

a:

b:

c:

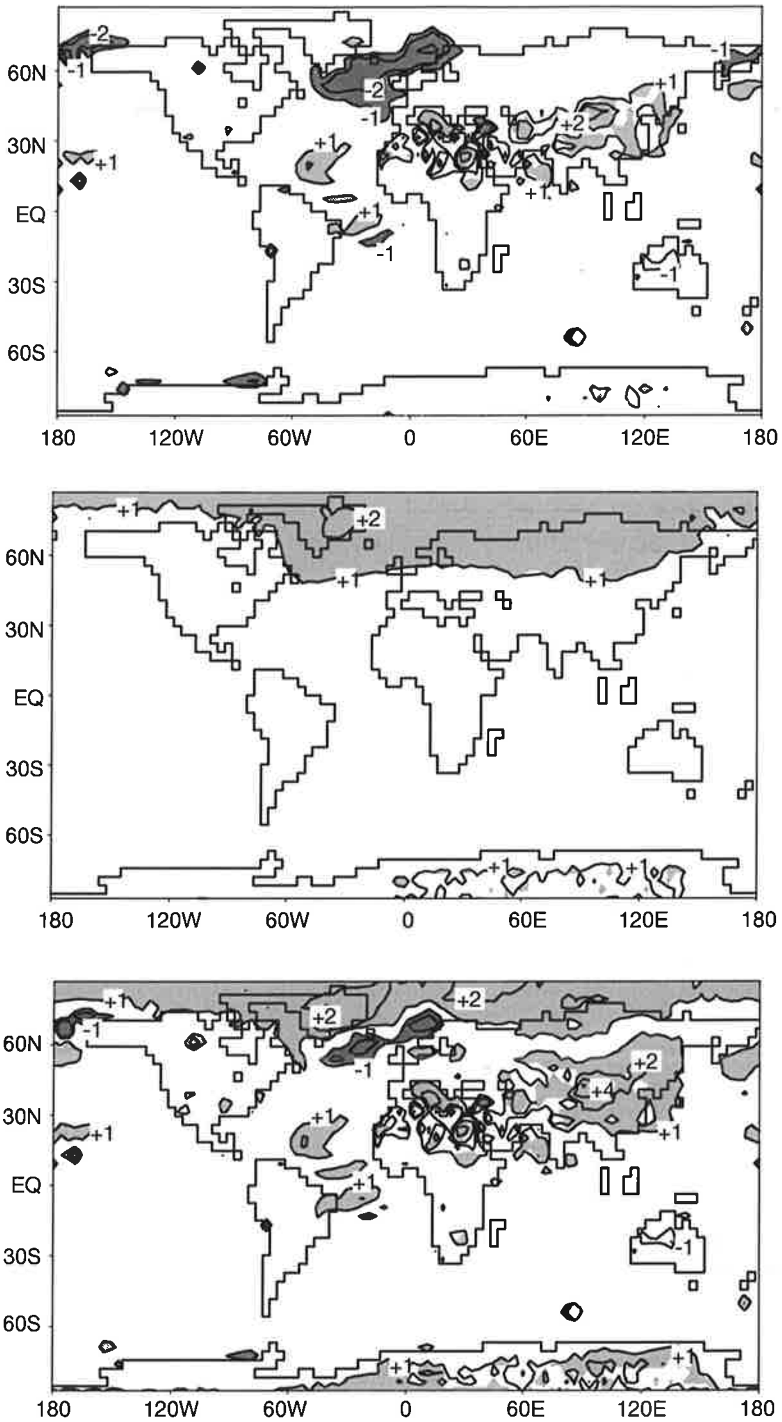

Fig. 3: The same as Fig. 2, but for the changes of the deuterium excess d (contour lines in all three plots at ±1, ±2, ±4‰, shading of significant changes like in Fig. 2a-2c).

The most striking difference is seen in the tropical Atlantic: While the negative $\delta^{18}O$ anomaly south of the equator increased in size, the positive anomaly of +2‰ north of the equator almost completely vanishes.

Figures 3a-3c shows the same sequence of anomalies for the deuterium excess. Three clear signals appear in the excess: (1) A local negative signal over the North Atlantic in the order of –1‰ mainly due to the local temperature changes (see Figures 3a and 3c). (2) A quite strong positive signal over Central and South Asia with a maximum of +4‰ over Tibet due to the long range climate changes caused by the melt water induced lowering of the North Atlantic SSTs (Fig. 3a). (3). A widespread positive anomaly between +1-2‰ over the high northern latitudes which is produced by the isotopic composition of the meltwater (see Fig. 3b). The scattered excess signal over northern Africa is probably related to model deficits for areas with only a few rainfall events over a period of several years [*Hoffmann et al.*, 1998].

Discussion

To understand the modeled isotope anomalies one has to consider the main influences of the $\delta^{18}O$ signal in precipitation. In the extra-tropics, under a present day climate, $\delta^{18}O$ strongly correlates with surface temperatures ("temperature effect", observed mean spatial slope: 0.61 ‰/°C [*IAEA*, 1992]). But in tropical regions surface temperatures are fairly constant over an annual cycle. There, the $\delta^{18}O$ signal shows in observations a weak negative correlation to the amount of precipitation ("amount effect", observed mean slope -1.3‰/100cm/year [*IAEA*, 1992]).

In the cold climate experiments surface temperatures are reduced in many land regions north of 30°N except parts of Asia and Alaska (Fig. 4a). Similar to the cold SST boundary conditions we find the strongest temperature drop (-20°C) in the area of the Greenland and Norwegian Sea. Strong cooling is also seen over the Greenland ice sheet (-8°C to -12°C) and the Bering Street (-8°C). The latter is directly correlated to a minimum in SSTs, too. The cooling in the Northern Pacific is caused by an intensified winter-time outflow of cold air from Siberia [*Mikolajewicz et al.*, 1997]. Cooling in the range of –2°C to –4°C is observed over most parts of Europe, Siberia and the east coast of North America. Such colder temperatures above the North Atlantic and Europe will reduce the amount of precipitation in Southern Greenland, over the Norwegian Sea and some parts of the European continent (Figure 4b). But additionally, the Innertropical Convergence Zone (ITCZ) is strongly shifted

Figure 4:

a:

b:

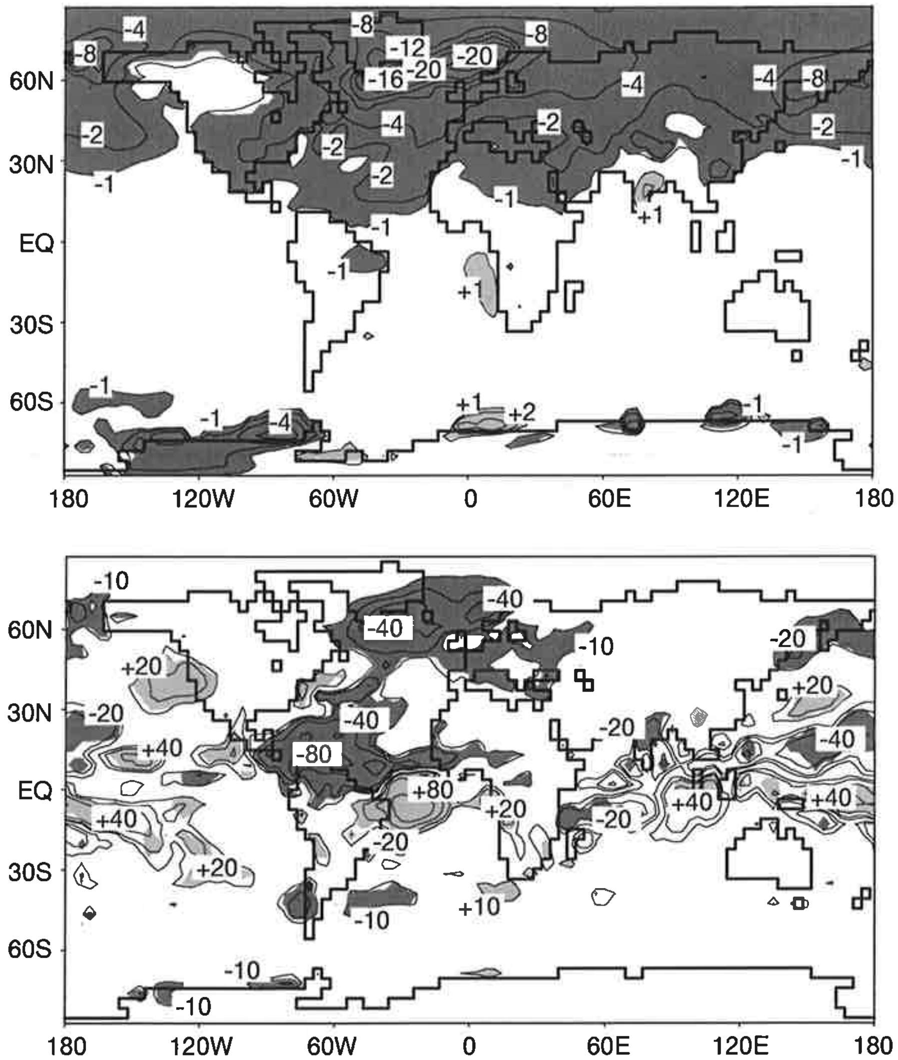

Fig. 4: (a): Changes of surface temperature between the ctrl and cold climate experiments. Contour lines are at ±1, ±2, -4, -8, -12, -16, -20°C. Significant temperature changes of the long-time mean value compared to the year-by-year variations (two-sided u-test, 95% level) are shaded in light resp. dark gray. (b): Same as Fig. 4a, but for the changed amount of precipitation. Contour lines are at ±10, ±20, -40, -80 mm/month, shading of significant changes like in Fig. 4a.

to the south. We observe a dipole-like change of the precipitation amount in the tropical Atlantic region: North of the equator precipitation is strongly reduced (down to -80 mm/month) in a band from Middle America to the Sahel zone while south of the equator precipitation amount increases (up to +140 mm/month). A similar, but weaker shift of the

ITCZ is observed in the Pacific, too. The dipole-like precipitation anomaly in the tropical Atlantic is the reason for the very similar dipole-like $\delta^{18}O$ anomaly seen in Figure 2a. In low latitudes the amount effect dominates the isotope signal. Therefore, lower precipitation amounts north of South America cause a relative enrichment of heavy isotopes in precipitation. Conversely, more precipitation south of the equator is responsible for the stronger depletion in $H_2{}^{18}O$. However, the positive branch of this pattern vanishes if we combine cooler SSTs and a changed $\delta^{18}O_{Ocean}$ (Fig. 2c). The additional isotope depletion caused by the changed $\delta^{18}O_{Ocean}$ field counterbalances the enrichment induced by the amount effect. As a result we see in Fig. 2c only a strong $\delta^{18}O$ anomaly between 0° and 30°S but almost no counterpart north of the equator.

The relatively strong and spatially coherent reaction of the deuterium excess is an astonishing result of our simulation, in particular since the isotopic composition of the meltwater itself was prescribed with an excess of 0‰. As mentioned above, the excess depends not on the temperature gradient between the evaporation and condensation site such as $\delta^{18}O$ and δD but is strongly affected by the isotopic disequilibrium during the evaporation at the sea surface. A slow (rapid) evaporation process, that means low (high) evaporation temperatures and a high (low) relative humidity, produces a small (large) Deuterium Excess [*Johnsen et al.*, 1989; *Merlivat and Jouzel*, 1979]. The direct impact of lower SSTs, therefore, is a lower excess of about -2‰ in the North Atlantic and North Pacific where the imposed temperature anomalies are strongest. The interpretation of the positive signal of up to 4‰ over Central and South Asia, however, is less straightforward. In our control simulation the highest excess values (d >14‰) were calculated in the very same region. This feature was already reported in a former version of the model [*Hoffmann et al.*, 1998] and is in good agreement with observations. Slight changes in the intensity of the monsoon or local precipitation conditions might have played a role in this very sensitive region, but are not understood yet. The most widespread signal, however, is caused by the isotopic anomaly of the meltwater itself (Fig. 3b). In fact the evaporation process reacts non-linear on the imposed changes of the water isotopes at the sea surface. Using the global evaporation model of Merlivat and Jouzel [*Merlivat and Jouzel*, 1979] we estimated the change of deuterium excess of the vapour formed in the region of the meltwater Δd:

$$\Delta d=8\cdot 1./\alpha_{18O}\cdot(1-k_{18O})/(1-k_{18O}\cdot h)\cdot\Delta\delta^{18}O_{Ocean}-1./\alpha_D\cdot(1-k_D)/(1-k_D\cdot h)\cdot\Delta\delta D_{Ocean}$$

with α_{18O} and α_D as equilibrium fractionation factors [*Majoube*, 1971], k_{18O} and k_D as the kinetic fractionation factors during evaporation from the ocean [*Merlivat and Jouzel*, 1979], h as the relative humidity (estimated 80%) and $\Delta\delta D_{Ocean}=8*\Delta\delta^{18}O_{Ocean}$ as the imposed change of the isotopic composition of sea water. In this simple calculation, we estimate the average change in the North Atlantic (north of 30°N) to $\Delta\delta^{18}O_{Ocean}$ =–3‰. The resulting change of the deuterium excess yields to about +1.9‰ pretty close to the change of more than +1‰ simulated by the ECHAM in northern high latitudes.

As mentioned above, these model experiments represent a sensitivity study and we do not expect an exact match with any observed isotopic composition changes during rapid cooling events in the past. However, given our simulation captures some of the atmospheric responses on a sudden cooling in the North Atlantic, the modeled isotope anomalies should at least be in the same direction and order of magnitude as available observations. A typical example of a rapid cooling event possibly caused by a meltwater spike in the North Atlantic might be the Younger Dryas climate reversal (about 12 kyrs BP). It has been shown that several aspects of their OAGCM simulation, from which our SSTs were taken, agree well with observations of the Younger Dryas (YD) cooling [*Mikolajewicz et al.*, 1997; *Schiller et al.*, 1997]. The timing of this cold reversal has been studied in detail on several Greenland ice cores, e.g. in [*Blunier et al.*, 1997; *Severinghaus et al.*, 1998; *Taylor et al.*, 1997]. Observed changes of $\delta^{18}O$, d-excess, surface temperature and precipitation amount on Summit, Central Greenland, are listed in table 1. The decrease in $H_2^{18}O$ in our model simulation is about half of the observed value. But modeled temperature anomalies are three times greater as expected from $\delta^{18}O$ anomalies if the modern (spatial) isotope-temperature-gradient of 0.67 ‰/°C [*Johnsen et al.*, 1989] is applied. Apparently, a lower (temporal) gradient has to be assumed for a cooling by a meltwater event. Such lower gradient has been reported by gas diffusion thermometry for the Younger Dryas [*Severinghaus et al.*, 1998] and also by borehole thermometry for the Last Glacial Maximum (LGM) [*Cuffey et al.*, 1995; *Dahl-Jensen et al.*, 1998; *Johnsen et al.*, 1995]. Latest ECHAM-4 simulations under full glacial boundary conditions are able to reproduce the changed isotope-temperature-relation. The deviation from modern spatial gradient is explained by an increased seasonality of precipitation over Greenland during the LGM [*Werner et al.*, 1999]. Less snowfall during winter season causes a bias of the mean $\delta^{18}O$ values measured in ice cores towards the higher summer signal. Here, we do find a similar change in seasonality for the cold climate simulations (not shown). This confirms previous findings that decreased winter precipitation over Greenland is mainly influenced by cooler SSTs but not by other glacial boundary conditions.

Location		$\delta^{18}O$ (‰)	d (‰)	T_{Surf} (°C)	Prec. (cm/y)
Summit, Greenland	GRIP/GISP2 ice cores	-5.3	+3		-10 to -12
	ECHAM-4	-2.7 (-1.3)	+3.2 (+1.0)	-11.4 (-11.4)	-14.7 (-14.7)
Sajama, Bolivia	ice cores C-1, C-2	-5.2			(positive)[1]
	ECHAM-4	-1.6 (-0.6)	-0.2 (-0.5)	-0.6 (-0.6)	+12.5 (+12.5)

[1] Relative higher net accumulation during the cold climate of the DCR was observed.

Table 1: Ice core data from Summit, Greenland, and Sajama, Bolivia, and corresponding model results of the ECHAM4 simulations. Changes in $\delta^{18}O$ values, deuterium excess d, surface temperature T_{Surf} and precipitation amount from the Younger Dryas stadial YD (resp. the deglaciation climate reversal DCR in the Sajama record) minus early Holocene values are compared to model anomalies caused by changed SSTs and $\delta^{18}O_{Ocean}$ boundary conditions. Model anomalies caused by changed SSTs alone are given in brackets. Ice core data was compiled from [*Alley et al.*, 1993; *Taylor et al.*, 1997; *Thompson et al.*, 1998].

A rapid decrease of about 3‰ in the deuterium excess during the transition from the YD to the Pre-boreal has been reported in the Dye3 core, Southern Greenland [*Dansgaard et al.*, 1989] and a similar value is measured on the GISP2 core [*Taylor et al.*, 1997]. Dansgaard et al. interprete this change as a redistribution of source areas of Greenland's precipitation towards cold high latitudinal regions. In their interpretation, this redistribution was mainly caused by a dramatic retreat of the sea ice border at the end of the YD. Our results, however, imply that for the interpretation of isotope records (and in particular of the deuterium excess) which stem from a region close to the meltwater input the isotopic composition of the meltwater might play a very important role, too.

The YD transition is also archived in different paleo-records in Europe. For example, isotope measurements in the calcite shells of freshwater ostracods from Lake Ammersee, Germany, allow the quantitative reconstruction of the local $\delta^{18}O_{Prec}$ signal. They show a decrease in $\delta^{18}O$ of 3-4‰ indicating in the classical $\delta^{18}O$/T-interpretation a temperature difference of 9°C for the YD [*von Grafenstein et al.*, 1999]. Our model simulations, however, show only a minor cooling over Europe (-2°C to -4°C). An additional anomaly of -2‰ over Central Europe can be related to the changed $\delta^{18}O_{Ocean}$ input. This might explain part of the discrepancies between isotope and pollen based temperature reconstruction [*Lotter*, 1991].

A rapid cooling after the beginning of the last deglaciation period is also observed in two ice cores retrieved from the Andes [*Thompson et al.*, 1998; *Thompson et al.*, 1995]. The well-dated Sajama record shows a deglaciation cold reversal (DCR) comparable to the YD signal observed in Greenland ice cores. However, the beginning of this reversal may have started about 1000 years before the onset of the YD [*Thompson et al.*, 1998]. The possible relevance of temperatures shifts in the tropics for a global climate change is therefore one of the most interesting, but still unanswered questions. Although our model simulations agree qualitatively well with the Samaja record (table 1), our findings in the Andes region are highly uncertain. The orography of the Andes is poorly resolved in the spatial T30 resolution, e.g. the grid box of the Sajama ice cap is only 2300 m a.s.l (the ice cores were drilled on 6542 m a.s.l.). The main water vapor transported to the Andes does originate from the tropical Atlantic and the Amazon region [*Grootes et al.*, 1989]. Therefore the dipole-like changes in the $\delta^{18}O$ signal seen in Fig. 2 will definitely have an imprint on the isotope composition of precipitation over the Andes. But since positive and negative anomalies are located so closely together, it is difficult to determine how the $\delta^{18}O$ signal on Sajama would be altered. But it seems very likely that changes of the $\delta^{18}O$ signal will be induced by the amount effect, and might not be strongly related to changes of the surface temperature.

Conclusions

Clearly, the modeled effects of a meltwater event on the $\delta^{18}O$ signal in precipitation do strongly depend on the applied boundary conditions. Especially, the effect of a changed $\delta^{18}O_{Ocean}$ field might be reduced if the isotope depletion of the fresh water input is weaker than assumed in the presented simulations. Therefore, the following list of possible effects seen in our sensitivity study should be taken with some caution. Nevertheless, it might help to lead to a better interpretation of paleo-records of fast climatic changes:

1. A rapid cooling of the atmosphere by a meltwater spike in the Labrador Sea causes a clear depletion of $H_2{}^{18}O$ in precipitation in most regions of the Northern Hemisphere polewards of 45°N. In general, the depletion of isotopes is related to a cooling of surface temperatures but enhanced due to the lower surface ocean ^{18}O isotopic composition.

2. Surface temperatures on the Greenland ice sheet are much colder than expected from $\delta^{18}O$ values. A change in seasonality of precipitation over Greenland results in a changed temperature-isotope-relation. The use of the present-day spatial relation to convert isotope data into past temperatures seems questionable for fast climate changes recorded in Greenland ice core records (since temperature changes are underestimated).

3. In the tropical Atlantic we do also observe significant changes in the isotopic composition of precipitation. These changes are not directly related to surface temperature but to changes of precipitation amounts induced by a southward shift of the ITCZ.

4. A depletion of the isotopic composition of ocean surface waters by a massive melt water input affects the $\delta^{18}O$ signal in precipitation over most parts of Europe and the Mediterranean Sea, eastern parts of North America and northern parts of South America. This additional decrease will lead to an overestimation of temperature shifts if these are calculated from present spatial $\delta^{18}O$-temperature relations.

5. Changes of the deuterium excess are even more affected by the imposed oceanic isotope anomaly due to the non-linearity of the evaporation process. Only a simultaneous interpretation of the water isotopes ($\delta^{18}O$ or δD) and the deuterium excess will provide us complete information for a full understanding of what happened during these dramatic climate events in the past.

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