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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 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 only 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 nonlinearity of the evaporation process.

1. Introduction

One of the most puzzling problems of climate research is the question of 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 paleorecords almost from the entire world, including the tropics [e.g., *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, for example, ice cores, pollen records, and marine sediment cores. From these proxy records, changes in, for example, temperature and precipitation are estimated. The required transfer functions are in general derived from present-day spatial variations of proxy (e.g., isotopes 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 that explicitly models the cycling of two stable water isotopes ($H_2{}^{18}O$, HDO) in the hydrological cycle. Modeling of both $H_2{}^{18}O$ and HDO enables additional analyses of the deuterium excess d (defined as $d = \delta D - 8\delta^{18}O$), a parameter which is related to temperature and/or humidity at the evaporation site [*Johnsen et al.*, 1989; *Merlivat and Jouzel*, 1979]. 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 $\delta^{18}O$ or δD) always coupled to changed surface temperatures? (3) A massive freshwater input with a strong depletion in heavy isotopes will alter the isotopic ocean surface water composition $\delta^{18}O_{ocean}$. How much does this affect the $\delta^{18}O$ signal in precipitation? (4) Will changes of the deuterium excess d reveal additional information?

2. 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 spinup time of 1 year. In the first experiment (hereinafter 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. However, 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) ECHAM-3/LSG [*Voss et*

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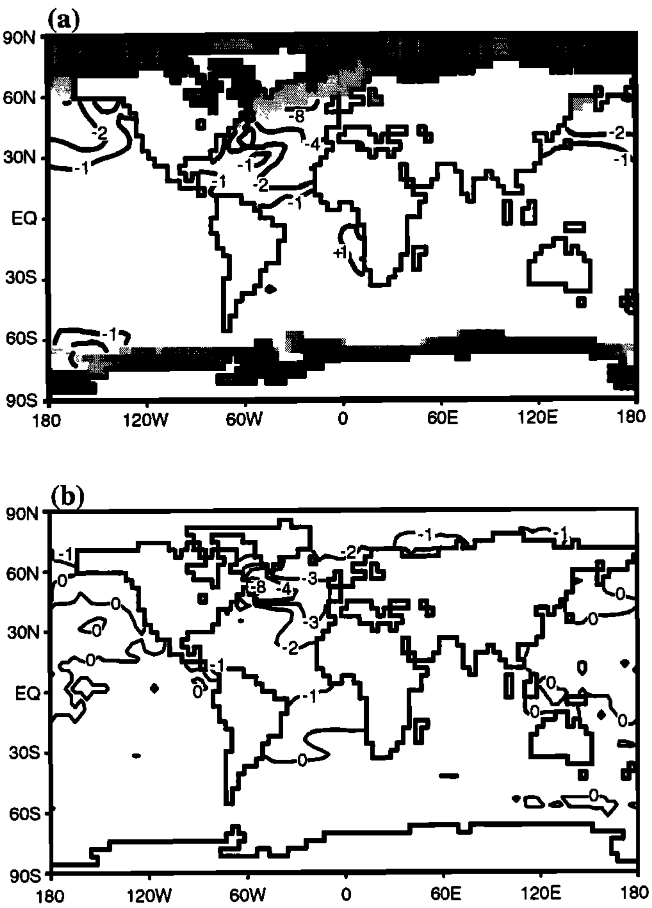

Figure 1. Differences of the boundary conditions between the control and cold climate experiments: (a) mean anomalies of sea surface temperatures (SSTs) and sea ice (temperature contour lines at -1°C, -2°C, -4°C, -8°C, and -16°C; heavily shaded area: sea ice prescribed in both climates at least half of the year, lightly shaded 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‰, and -8‰).

al., 1998]. The ECHAM-3/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 and the OAGCM meltwater experiment to use as boundary conditions for the isotope control and cold climate experiments, respectively (Figure 1a). For the cold climate scenario with an additional changed $\delta^{18}O_{ocean}$ field we used results of a GCM experiment with an OGCM coupled to an atmospheric energy balance model [*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 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. In this experiment a highly idealized $\delta^{18}O$ of seawater was included. 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 pattern that is shown is a combination of the $\delta^{18}O$ of meltwater and $\delta^{18}O$ changes of seawater caused by changes of the ocean circulation. 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 for SSTs and $\delta^{18}O_{ocean}$ we did not change any other boundary condition such as topography, ice shield distribution, or insolation, which were all set to present-day values in the experiments.

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.

3. 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 into two parts: (1) the $\delta^{18}O$ changes caused by colder SSTs alone (Figure 2a), and (2) additional $\delta^{18}O$ changes in precipitation for the colder climate caused by the assumed change in $\delta^{18}O_{ocean}$ (Figure 2b). Colder SSTs alone affect the $\delta^{18}O$ signal over the Atlantic region, Scandinavia, and the western part of Europe (Figure 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 Strait area associated with the prescribed cooling and the increased sea ice cover. This Pacific signal is much weaker than the Atlantic one 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 (Figure 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 precipitation) 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 ef-

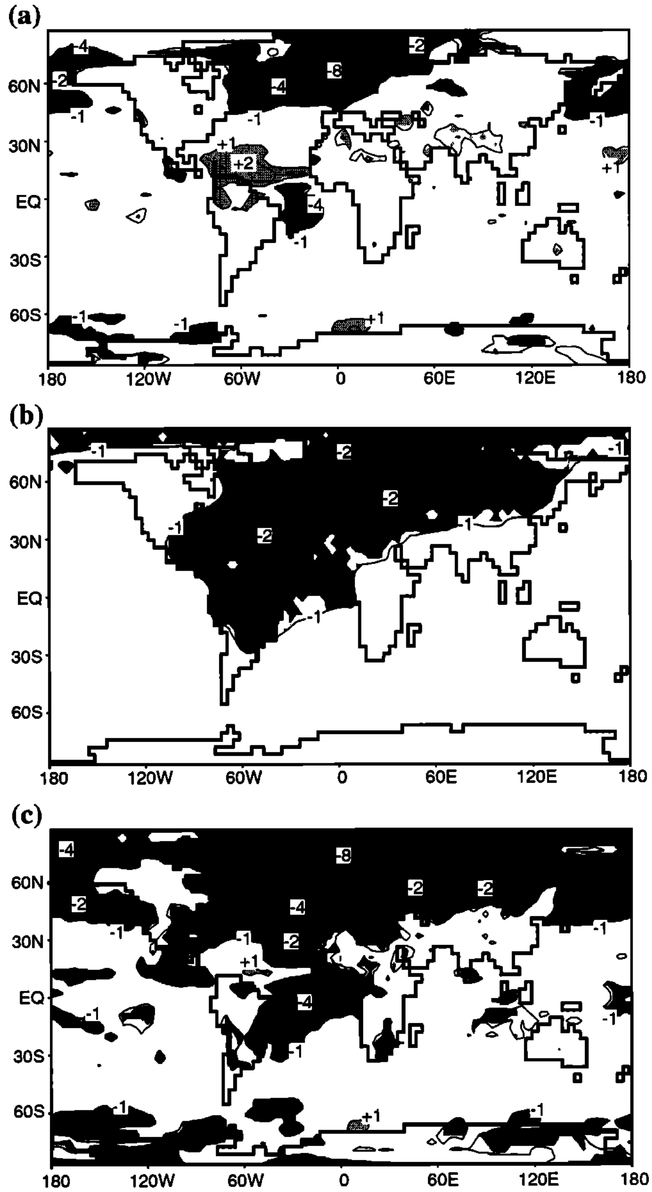

Figure 2. (a) Changes of the $\delta^{18}O$ values in precipitation of the cold climate minus the control climate, if only SSTs are changed for the cold climate. (b) Changes of the $\delta^{18}O$ values in precipitation for the cold climate simulation with both SST and $\delta^{18}O_{ocean}$ changed, minus the cold climate simulation with only SST changed. (c) Changes of the $\delta^{18}O$ values in precipitation of the cold climate minus the control climate, if both SSTs and $\delta^{18}O_{ocean}$ are changed for the cold climate. Contour lines in all three plots are at ±1‰, ±2‰, ±4‰, and ±8‰. Significant $\delta^{18}O$ changes (two-sided u test, 95% level) are denoted by light shading (positive anomalies) and heavy shading (negative anomalies), respectively.

fects of changed SSTs and changed $\delta^{18}O_{ocean}$ (Figure 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. 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 show 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 of 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 meltwater induced lowering of the North Atlantic SSTs (Figure 3a); and (3) a widespread positive anomaly between +1‰ and +2‰ over the high northern latitudes which is produced by the isotopic composition of the meltwater (see Figure 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].

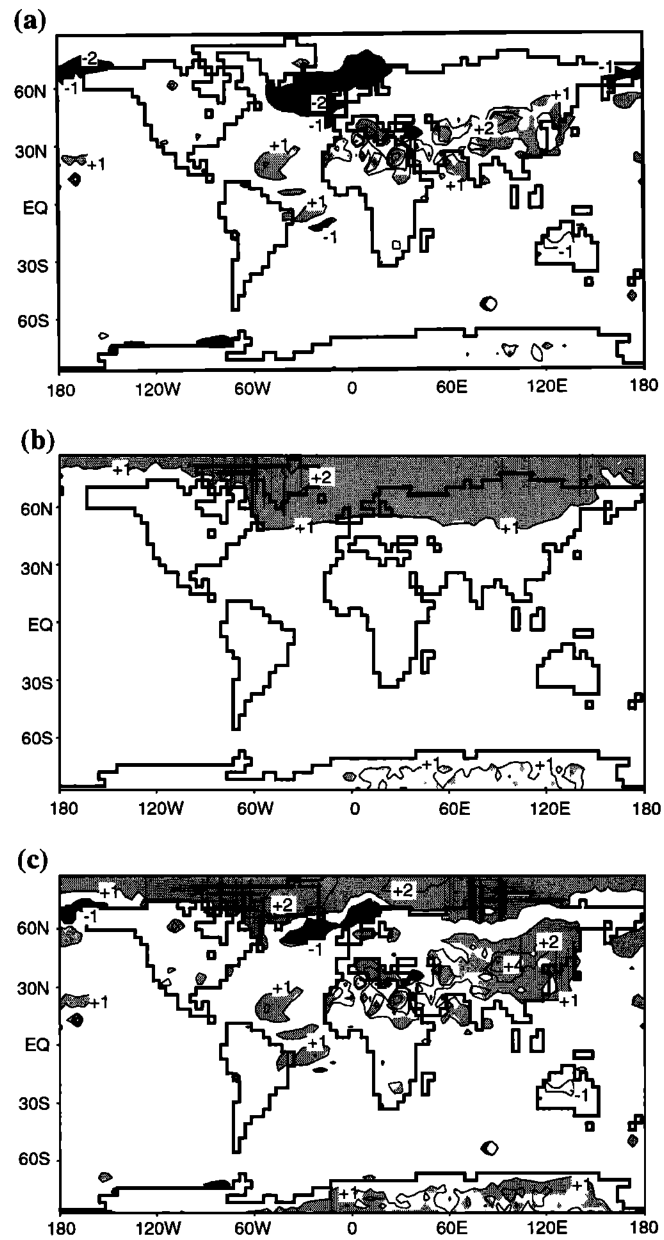

Figure 3. The same as Figure 2, but for the changes of the deuterium excess d (contour lines in all three plots at ±1‰, ±2‰, and ±4‰; shading of significant anomalies as in Figures 2a-2c).

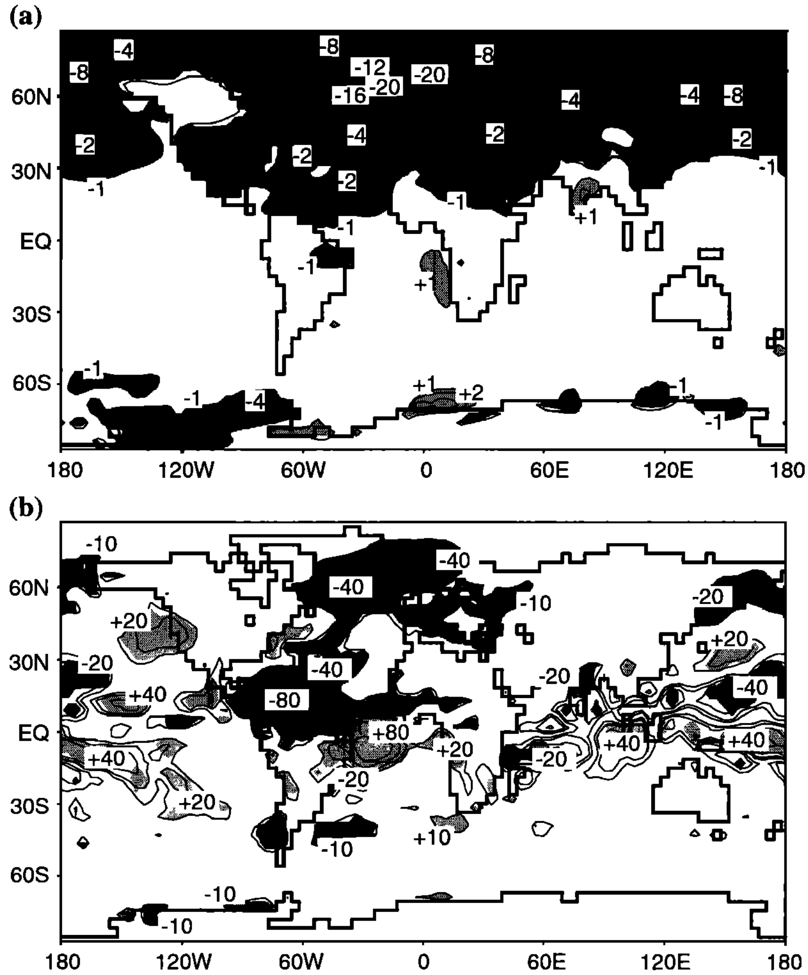

Figure 4. (a) Changes of surface temperature between the control and cold climate experiments. Contour lines are at ±1°C, ±2°C, -4°C, -8°C, -12°C, -16°C, and -20°C. Significant temperature changes (two-sided *u* test, 95% level) are denoted by light shading (positive anomalies) and heavy shading (negative anomalies), respectively. (b) Same as Figure 4a, but for the changed amount of precipitation. Contour lines are at ±10, ±20, -40, -80 mm month^{-1}; shading of significant anomalies as in Figure 4a.

4. Discussion

To understand the modeled isotope anomalies, one has to consider the main influences of the $\delta^{18}O$ signal in precipitation. In the extratropics, under a present-day climate, $\delta^{18}O$ strongly correlates with surface temperatures ("temperature effect", observed mean spatial slope: 0.61‰ °C^{-1} [*International Atomic Energy Agency (IAEA)*, 1992]). However, 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)$^{-1}$ yr^{-1} [*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 (Figure 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 over sea ice. Strong cooling is also seen over the Greenland ice sheet (-8°C to -12°C) and the Bering Strait (-8°C). The latter is directly correlated to a minimum in SSTs, too. The cooling in the northern Pacific is caused by an intensified wintertime 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 and over the Norwegian Sea and some parts of the European continent (Figure 4b). However, an additional feature is that the Intertropical Convergence Zone (ITCZ) over the Atlantic is strongly shifted to the southeast. 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^{-1}) in a band from middle America to the Sahel zone, while south of the equator precipitation amount increases (up to +140 mm month^{-1}). 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 $\delta^{18}O_{ocean}$ (Figure 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 Figure 2c a strong $\delta^{18}O$ anomaly only 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 does not depend 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 (Figure 3b). In fact, the evaporation process reacts nonlinearly on the imposed changes of the water isotopes at the sea surface. Using the global evaporation model of Merlivat and Jouzel [1979], we estimated the change of deuterium excess of the vapor formed in the region of the meltwater Δd:

$$\Delta d = 8\,\frac{1}{\alpha_{18O}}\,\frac{(1-k_{18O})}{(1-k_{18O}h)}\,\Delta\delta^{18}O_{ocean} - \frac{1}{\alpha_D}\,\frac{(1-k_D)}{(1-k_D h)}\,\Delta\delta D_{ocean}$$

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 seawater. In this simple calculation we estimate the average change in the North Atlantic (north of 30°N) as $\Delta\delta^{18}O_{ocean}$=-3‰. The resulting change of the deuterium excess yields about +1.9‰, fairly close to the change of more than +1‰ simulated by the ECHAM in northern high latitudes. These changes are of the same order of magnitude as present-day deuterium excess anomalies caused by a more realistic assumption of the deuterium excess values of seawater d_{ocean}. It is well known that the present d_{ocean} value is not zero (as assumed for all our simulations) but can vary and is mostly negative in evaporative zones. If we assume d_{ocean} values for the northern Atlantic in the range of -3‰ to 1‰, we calculate by the given formula an excess value d of the vapor formed in that region of -2.7‰ to 0.9‰. A possible change of Δd_{ocean} between the control and cold climate simulations (caused by the freshwater input) would result in an additional Δd anomaly in precipitation. However, to calculate these effects more quantitatively, an isotope ocean-atmosphere coupled GCM with more realistic surface fluxes of $\delta^{18}O$ and d should be used.

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 that our simulation captures some of the atmospheric responses to 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 kyr B.P.). It has been shown that several aspects of the 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., *Blunier et al.*, 1997; *Severinghaus et al.*, 1998; *Taylor et al.*, 1997]. Observed or estimated changes of $\delta^{18}O$, d excess, surface temperature, and precipitation amount at 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. However, modeled temperature anomalies are 3 times greater than expected from $\delta^{18}O$ anomalies if the modern (spatial) isotope-temperature gradient of 0.67‰ °C^{-1} [*Johnsen et al.*, 1989] is applied. Apparently, a lower (temporal) gradient has to be assumed for a cooling by a meltwater event. Such a 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]. The latest ECHAM-4 simulations under full glacial boundary conditions are able to reproduce the changed isotope-temperature relation. The deviation from the modern spatial gradient is explained by an increased seasonality of precipitation over Greenland during the LGM [*Werner et al.*, 2000]. Less snowfall during winter season causes a bias of the mean $\delta^{18}O$ values measured in ice cores toward 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.

A rapid decrease of about 3‰ in the deuterium excess during the transition from the YD to the Preboreal has been reported in

Table 1. Ice Core Data from Summit, Greenland, and Sajama, Bolivia, and Corresponding Model Results of the ECHAM-4 Simulations

Location	$\delta^{18}O$, ‰	d, ‰	T_{surf}, °C	Prec., cmy^{-1}
Summit, Greenland				
GRIP/GISP2 ice cores	-5.3	+3		-10 to -12
ECHAM-4	-2.7	+3.2	-11.4	-14.7
	(-1.3)	(+1.0)	(-11.4)	(-14.7)
Sajama, Bolivia				
ice cores C-1, C-2	-5.2			(positive)*
ECHAM-4	-1.6	-0.2	-0.6	+12 5
	(-0 6)	(-0.5)	(-0.6)	(+12.5)

Changes in $\delta^{18}O$ values, deuterium excess d, surface temperature T_{surf} and precipitation amount from the Younger Dryas stadial YD (or 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 parentheses. Ice core data were compiled from *Alley et al* [1993], *Taylor et al.* [1997], and *Thompson et al.* [1998]. GRIP denotes Greenland Ice Core Project, and GISP denotes Greenland Ice Sheet Project.

*Relative higher net accumulation during the cold climate of the DCR was observed.

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. interpreted this change as a redistribution of source areas of Greenland's precipitation toward 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 paleorecords 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 $\delta^{18}O_{prec}$ of 3-4‰ between the Preboreal and the YD [*von Grafenstein et al.*, 1999]. Using the classical $^{18}O/T$ interpretation with a gradient of 0.6‰ $°C^{-1}$ for Europe results in a temperature difference of 5°C to 6.7°C for the YD. 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.

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.*, 1995, 1998]. 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 temperature 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; for example, the grid box of the Sajama ice cap is only 2300 m above sea level (asl) (the ice cores were drilled at 6542 m asl). 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 Figure 2 will definitely have an imprint on the isotope composition of precipitation over the Andes. However, since positive and negative anomalies are located so close together, it is difficult to determine how the $\delta^{18}O$ signal on Sajama would be altered. It seems very likely, though, 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.

The observed deuterium excess anomalies in India and Asia (Figure 3a) might be related to changes in the monsoon intensity during the deglaciation period [e.g., *Zonneveld et al.*, 1997]. However, so far there exist no measurements of the deuterium excess from any water isotope archive in this region for the Younger Dryas or a similar cold climate period. Therefore a validation of the simulated *d* anomalies in the Asian region by observations remains an open question.

5. Conclusions

Clearly, the modeled effects of a meltwater event on the $\delta^{18}O$ signal in precipitation 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 freshwater 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 lead to a better interpretation of paleorecords of fast climatic changes:

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 poleward of 45°N. In general, the depletion of isotopes is related to a cooling of surface temperatures but is enhanced due to the lower surface ocean ^{18}O isotopic composition.

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).

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

A depletion of the isotopic composition of ocean surface waters by a massive meltwater 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.

Changes of the deuterium excess are even more affected by the imposed oceanic isotope anomaly due to the nonlinearity of the evaporation process. Similar to $H_2{}^{18}O$, an interpretation of deuterium excess anomalies as changes in the surface temperatures and/or humidity at the evaporation site, solely, might yield erroneous results. However, more realistic estimates of the deuterium excess anomalies of ocean surface water caused by a meltwater input are needed to evaluate the importance of the reported nonlinearity effect.

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