IN-47
045 006

Water isotopes in precipitation:

Data/model comparison for present-day and past climates

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Variations of HDO and $H_2{}^{18}O$ concentrations are observed in precipitation both on a geographical and on a temporal basis. These variations, resulting from successive isotopic fractionation processes at each phase change of water during its atmospheric cycle, are well documented through the IAEA/WMO network and other sources. Isotope concentrations are, in middle and high latitudes, linearly related to the annual mean temperature at the precipitation site. Paleoclimatologists have used this relationship to infer paleotemperatures from isotope paleodata extractable from ice cores, deep groundwater and other such sources. For this application to be valid, however, the spatial relationship must also hold in time at a given location as the location undergoes a series of climatic changes. Progress in water isotope modeling aimed at examining and evaluating this assumption has been recently reviewed (Jouzel et al., 1997) with a focus on polar regions and, more specifically, on Greenland. This article was largely based on the results obtained using the isotopic version of the NASA/GISS Atmospheric General Circulation Model (AGCM) fitted with isotope tracer diagnostics. We extend this review in comparing the results of two different isotopic AGCMs (NASA/GISS and ECHAM) and in examining, with a more global perspective, the validity of the above assumption, i.e. the equivalence of the spatial and temporal isotope-temperature relationship. We also examine recent progress made in modeling the relationship between the conditions prevailing in moisture source regions for precipitation and the deuterium-excess of that precipitation.

1. Introduction

The validity of using the isotopic composition of paleowater (e.g., from deep ice cores or lake sediments) for inferring paleotemperatures has been reviewed in recent articles principally focusing on Greenland ice core data (Jouzel et al., 1997 and 1998). In the standard reconstruction approach, the linear spatial relationship ($\delta = a\ T_S + b$) between the surface temperature, T_S, and the isotopic content of the precipitation (δD or $\delta^{18}O$ expressed in per mill with respect to V.SMOW, the Vienna Standard Mean Ocean Water (Craig, 1961)) is assumed to hold at a single geographic point as it experiences a sequence of climatic changes over time. The isotope paleothermometer thus employs a so-called "modern analogue method" similar to that adopted in most other paleoclimate reconstruction. However, the assumption that the present-day spatial slope ("a" in the equation above) serves as a reliable surrogate for the more relevant "temporal" slope is now being challenged, particularly for Greenland where independent long term (glacial-interglacial) estimates of temporal slopes appear considerably lower than the observed present-day spatial slopes.

In the present review article, we extend the discussions of Jouzel et al. (1997, 1998) by examining the performances of two different isotopic GCMs (NASA/GISS New York and ECHAM Hamburg) and by accounting for several recent experiments (Armengaud et al., 1998, Hoffmann et al, 1998 and in press; Werner et al., 1998; Cole et al., in press). These isotopic GCMs are particularly useful since they allow, by simulating different climatic periods, a direct comparison between spatial and temporal δ/T_S relationships. Results from present-day, LGM (Last Glacial Maximum at 21 kyr BP), mid-Holocene (6 kyr BP) and 2 * CO_2 experiments will be discussed. We also consider some aspects linked with the climatic information contained in the deuterium excess parameter, $d = \delta D - 8 * \delta^{18}O$.

2. Estimates of Temporal δ/T_S Relationships.

The present-day spatial $\delta^{18}O/T_S$ relationship is well documented worldwide, with slopes ranging from ~ 1.1 ‰/°C in high-latitude areas to virtually zero in tropical regions, where $\delta^{18}O$ is more strongly correlated to the amount of precipitation. Unfortunately (Rozanski et al., 1992), the temporal $\delta^{18}O/T_S$ slope cannot be similarly well-characterized due to a paucity of relevant data estimating the temporal slope requires records of both surface temperature and isotope concentration in precipitation that span a long period of time at the same site. Still, some estimates are possible. Through the IAEA/WMO network initiated in 1961, about three decades of data are available for a number of sites, most of them situated in Europe. Using these data, Rozanski et al. (1992) noted that decadal-scale changes of $\delta^{18}O$ content in precipitation over Europe closely follow decadal-scale changes of surface air temperature, thus confirming that isotope records convey information on past temperature changes. The average $\delta^{18}O/T_S$ temporal relationship inferred from these data was about 0.6‰/°C, which is close to the spatial slope observed for European stations in the IAEA/WMO network (IAEA, 1992).

In polar regions (Jouzel et al., 1997), five different approaches, relevant to a wide range of time scales, have been employed to estimate past temperatures for independent validation of the isotope thermometer: (1) use of temperatures recorded in the vicinity of the isotope sampling site (if not at the site itself) to extend the comparison over the period of instrumental observations; (2) use of Automatic Weather Stations (AWS) and satellite microwave brightness temperatures in conjunction with high resolution isotope profiles (this approach was employed at the GISP2 site for 1987 - 1990); (3) analysis of the changing percentage of melt layers in ice cores to give estimates of shifts in summer temperatures during various periods of the Holocene; (4) estimation of paleotemperatures over a wide range of timescales (centuries to tens of millennia) from bore hole temperature profiles; and (5) analysis of the temperature-dependent change in snow accumulation at a given location. Various studies employing these approaches provide compelling evidence that the temporal slope in Greenland tends to be lower than the present-day spatial slope (see Jouzel et al., 1997, and references therein). The temporal slope appears consistently closer to the spatial slope during recent times, i.e. during the Holocene period, than it does for the glacial/interglacial timescale. For this latter timescale, the difference between the spatial and temporal slopes can reach a factor of two or more, at least over Greenland, as independently shown by Cuffey et al. (1995), Johnsen et al. (1995) and recently by Dahl-Jensen et.al. from paleothermometry information retrieved from the GISP2 and GRIP cores, respectively. They estimated a value of ~0.3 ‰/°C for the LGM/present-day temporal slope, whereas the observed present-day spatial slope over Greenland is 0.67 ‰/°C which has been recently confirmed in applying an inverse Monte-Carlo method (Dahl-Jensen et al., 1998). Unfortunately, the same method is not applicable for the comparable short and rapid Dansgaard-Oeschger events since the temperature signal they most probably imprinted on the ice is already diffused out. For such rapid events, no direct temperature reconstruction from bore hole temperatures is available but there is some hope to estimate temperature changes in applying the paleothermometry method recently developed by Severinghaus et al. (1998).

In mid-latitudes, the temporal slopes for the glacial/interglacial timescale can be estimated from paleo-groundwater, with the noble gas content of such groundwater providing the necessary temperature data. Such a study conducted in the Great Hungarian Plain led to an estimate of the of 0.59 ‰/°C, which is similar to the present-day spatial slope in Europe. The glacial/interglacial temporal slope obtained from some other aquifers, however, is lower, with values of around 0.3 - 0.4 ‰/°C in England and Germany (see Rozanski et al., 1992). The information contained in North American paleogroundwater is more ambiguous. They may be either isotopically lighter or heavier than Holocene groundwater, depending on the area considered (Phillips et al., 1986; Stute et al., 1992; Plummer, 1993; Dutton, 1995), though the estimated Last Glacial Maximum temperatures are consistently cooler.

3. Use of Simple Rayleigh-Type Models

Different types of models have been developed to understand the water isotope cycle. Rayleigh-type distillation models are the simplest ones. They are useful because they include the main physical controls over the global distributions of δD and $\delta^{18}O$ in precipitation yet are simple enough for comprehensive analyses and efficient first-order sensitivity studies. Briefly, a Rayleigh distillation model (Dansgaard, 1964) computes the isotopic content of an idealized, isolated air parcel traveling from an oceanic source towards a region where condensation and finally precipitation takes place. Condensate forms in isotopic equilibrium with the surrounding vapor and is removed immediately from the parcel. Under this framework, the isotope content of the precipitation is a unique function of the initial masses of isotope and water vapor within the air parcel, of the water vapor mass remaining when the precipitation forms, and of the assumed temperature-dependent fractionation coefficients. The water masses can themselves be characterized in terms of ambient temperatures and vapor pressures.

Rayleigh-type distillation models successfully reproduce the main characteristics of the global water isotope cycle, in particular the observed seasonal and spatial variations, the observed relationships with local temperature, and the strong link between δD and $\delta^{18}O$ (Craig, 1961; Dansgaard, 1964; Friedman et al., 1964). These models work particularly well in middle and high latitudes where precipitation generation is not dominated by large convective systems. Their ability to simulate the present-day temperature/isotope relationships correctly in these regions was, in fact, a major justification for the assumed equivalence between temporal and spatial δ/T_S slopes. Enhancements of Rayleigh-type models include the estimation of initial isotope concentrations in vapor from sea surface conditions (Merlivat and Jouzel, 1979) and a treatment of kinetic fractionation processes during snow formation (Jouzel and Merlivat, 1984; Ciais and Jouzel, 1994). They show how sea surface temperature, T_W, and the temperature of formation of the precipitation, T_C, combine to influence the isotopic content of a precipitation. For example, Aristarain et al. (1986) showed that for Antarctic snow $\Delta\delta^{18}O$ can be equated, over a large range of snow formation temperature, to $1.1\ \Delta T_C - 0.55\ \Delta T_W$, where Δ represents a difference between two climates.

This equation implies that the constancy of the evaporative source is a prerequisite for using the spatial δ/T_S slope as a surrogate for the temporal slope. Indeed, an equal, simultaneous change of the temperatures in the source region and at the precipitation site would result in a temporal slope that is lower than the spatial slope by a factor of about 2 (Aristarain et al., 1986; Boyle, 1997). The possible effects of source temperature variation on the temporal slope have been considered in various studies (Siegenthaler and Matter, 1983; Grootes, 1993, Jouzel et al., 1997). Boyle (1997) recently suggested that the discrepancy between central Greenland bore hole temperature and the isotopic composition of LGM ice can be explained by cooler tropical temperatures during the LGM. This author assumes that the spatial $\delta^{18}O/T_S$ slope is time-invariant and that the intercept varies with tropical temperatures and global isotope composition. The assumption that tropical ocean temperatures were 5°C cooler than they are at present (see Boyle, 1997 and references herein) leads to an apparent temporal $\delta^{18}O/T_S$ slope of 0.37 ‰/°C, which is close to that derived from bore hole paleothermometry.

In the real and very complex world, however, many other explanations for the difference between the temporal and spatial slopes are viable, and simple Rayleigh-type models cannot address all of these explanations. In Antarctica, for example, observations and simple models agree only with respect to the temperature of precipitation formation, which is roughly the temperature just above the inversion layer (Robin, 1977), a temperature much warmer than the surface temperature. A change in the strength of the inversion layer between climates, a change difficult to predict with a simple Rayleigh-type model, can have a significant impact on the temporal slope. In addition, a simple Rayleigh-type model cannot properly account for the complexity of dynamical and microphysical processes leading to the formation of individual precipitation events, or for the changes in ocean surface characteristics, in surface topography and in atmospheric circulation associated with important climatic changes. In the light of these deficiencies, the physics of water isotope fractionations have been incorporated into atmospheric GCMs, as discussed in the next section.

4. Use of Isotopic GCMs for Present-Day Climate

An isotopic GCM is essentially an atmospheric GCM fitted with special tracer diagnostics that follow HDO and $H_2{}^{18}O$ tracers through every stage of the water cycle. Equilibrium and kinetic fractionation processes are accounted for at every change of phase (surface evaporation, atmospheric condensation, and reevaporation of precipitation). Joussaume et al. (1984) pioneered the approach, simulating global fields of isotope concentration for present-day January climate using a low-resolution version of the GCM of the Laboratoire de Météorologie Dynamique (LMD/Paris). Jouzel et al. (1987) generated a full annual cycle of isotope fields with the 8° * 10° NASA Goddard Institute for Space Studies (GISS) GCM and studied the sensitivity of the model results to various parameterizations (Jouzel et al., 1991). Both the LMD and the GISS isotopic GCMs have since been run at higher spatial resolutions (Joussaume and Jouzel, 1993; Charles et al., 1994, 1995, Andersen, 1997). Water isotopes have more recently been incorporated into two different versions of the ECHAM GCM (Hoffmann and Heimann, 1993, 1997, Hoffmann et al., 1998a ; Werner et al., 1998), and into the GENESIS GCM, which was developed at the National Center for Atmospheric Research in the U.S. (Matthieu et al., submitted). We will discuss here in some more detail the results of the GISS (8° * 10° version) and the ECHAM (T42 resolution corresponding to 2.8°*2.8°) model.

As discussed in Jouzel et al. (1987), Joussaume and Jouzel (1993) and Hoffmann et al. (1998a), these models reproduce well the main characteristics of water isotope distributions in present-day precipitation. The models realistically simulate the decrease of $\delta^{18}O$ in higher latitudes, the lack of a latitudinal gradient in the tropics, and the land-sea contrast in isotope concentrations, among other features. Although the different models use quite different physical parameterizations and numerical schemes to describe the global climate, obviously they capture the main characteristics of the global cycle of the water isotopes. We illustrate the striking correspondence between the GISS and ECHAM models by comparing the simulated zonal means of $\delta^{18}O$ (see Figure 1a). Only at high northern latitudes and in particular over the Arctic ocean does the GISS model predict a precipitation up to 8‰ more depleted than the ECHAM model. This difference can only partly be attributed to the 4 to 6°C colder temperatures simulated there by the GISS model and, therefore, is not completely understood.

Here Figure 1 :

Figure 1: Zonal mean of (a) $\delta^{18}O$ in precipitation (control run), (b) the seasonal gradient in ‰ /°C (i.e. the slope of the $\delta^{18}O$ /Temperature relation for the mean 12 months of the control run), (c) the difference of the annual $\delta^{18}O$ between the LGM (corrected for a supposed 1.6 ‰ enrichment of the glacial ocean) and the Control simulation and (d) the temporal (LGM-Control) gradient.

At the regional level, a recent simulation performed by Werner et al. (1998) has clearly shown that the quality of an isotopic GCM in terms of its ability to simulate correctly the observed isotopic distribution can be excellent when using a high resolution. A prerequisite for this is a good simulation of the region's climate characteristics. This is illustrated in Figure 2, in which the observed and simulated $\delta^{18}O$-T_S distributions over Greenland are compared.

Here Figure 2.

Figure 2 : Spatial linear relationship between δ and surface temperature on the Greenland ice sheet (Dansgaard slope), a) observations, b) model results (from Werner et al., 1998).

In addition to the long-term temporal $\delta^{18}O$-T_S relationship (see section 5), one can define a modern seasonal isotope-temperature gradient. In Figure 1b, using the 12 long-term monthly means of $\delta^{18}O$ and T_s, we calculated for each grid a $\delta^{18}O$-T_S slope. This 'seasonal' slope is typically about a factor of 2 weaker than the spatial slope in the respective regions. This is mainly due to seasonal changes in the characteristics of the vapor source that strongly diminish the seasonal relationship compared to the spatial or the long-term (interannual) temporal gradient (Siegenthaler and Matter, 1983 ; Aristarain et al.1986). Over land, both models simulate realistically a seasonal gradient between 0.2 and 0.6 ‰/°C, with higher values in the interior of the continents (see the maxima at 50°N in Fig. 1b where the northern hemisphere's land masses are largest). Continental re-evaporation seems to amplify the isotope response to local temperatures. Because the isotopic composition of water remains unchanged in both models during re-evaporation, a strong contribution of recycled water to local rainfall (usually during summer, see Koster et al., 1993) enriches isotopically the subsequent precipitation and thus amplifies the seasonal amplitude of the water isotopes. However, Figure 1b demonstrates differences between the models, too. The ECHAM produces systematically higher seasonal gradients most probably due to differences in the model's land surface schemes and/or the seasonal vapor transport.

Long simulations (10 years or more), using observed sea surface temperatures from the recent period as boundary conditions, have recently been analyzed (Hoffmann et al., 1998a ; Cole et al., in press). These simulations capture the weak correlation between the isotopic signal and the temperature but differ with respect to the observed anticorrelation with precipitation amount, which seems overestimated by Cole et al. (submitted). These authors have indeed identified the changes in the amount of precipitation and in the contributions of local and nearby sources as the most important determinants of simulated interannual isotopic changes. As previously examined by Cole (1993), the simulation of Hoffmann et al. (1998 a) demonstrates that the strongest interannual climate anomaly, the El Nino Southern Oscillation, imprints a strong signal on water isotopes. This makes the water isotopes a good candidate for long term reconstruction of the ENSO phenomenon given that suitable archives, which conserve the isotopes in seasonal resolution, can be found.

5. Simulated present-day / glacial temporal isotope/temperature relationships

Although certain model weaknesses in Figures 1a and 1b can be identified, the present-day performances of the models are adequate enough to justify simulations of isotope behavior in alternative climates with the goal of improving interpretations of isotope paleodata. The application of the isotopic GCM toward this goal is straightforward. In a simulation of the present-day climate, the ambient environmental conditions (e.g., temperatures) in a region of interest and the isotope concentrations in precipitation or vapor there are carefully noted. Additional climates are then simulated (through modification of solar forcing, surface boundary conditions, etc.) in separate numerical experiments, and the same variables are recorded again for each climate. The first climate modeled for this analysis (other than the present-day climate) is that of the Last Glacial Maximum (Joussaume and Jouzel, 1993 ; Jouzel et al., 1994 ; Charles et al., 1994, 1995; Hoffmann et al., 1997). The LGM is particularly relevant for several reasons: (a) the glacial climate is very different from the current climate; (b) the LGM boundary conditions are adequately known (CLIMAP, 1981); and (c) isotope paleodata are available for this period in both polar and temperate regions, allowing partial validation of model results. Comparisons with available paleodata (Joussaume and Jouzel, 1993, Jouzel et al., 1994, Hoffmann et al., 1997) suggest that GCMs reproduce LGM isotope concentrations reasonably well.

Figure 3 shows the $\delta^{18}O$ anomalies simulated by the GISS and the ECHAM model for the LGM period. Although the basic features of both simulations are the same (a strong isotopic depletion in high latitudes due to the continental ice masses, the larger sea ice extent and the southward shift of the oceanic polar front as prescribed by the CLIMAP data set), some interesting differences can be stated, presumably caused by slightly different imposed boundary conditions. For example, the GISS simulation used the icesheet reconstruction devised by (1976, 1981) while the ECHAM simulation used a more recent one devised by Tushingham and Peltier (1991). In this new reconstruction, the Laurentide icesheet is prescribed about 1 km lower than before and the remaining ice mass is distributed on West Antarctica and Greenland, which one thus assumed to be considerably higher than before. In these regions, the modification of the orography in the new reconstruction produces a stronger rainout of air masses and, consequently, more depleted precipitation. A comparison with the observed isotope signal of the corresponding ice cores from Central Greenland and West Antarctica leads us to the conclusion that the isotope models do not support the new ice sheet reconstruction. In fact, this reconstruction was already criticized for other reasons (Edwards, 1995).

Here Figure 3

Figure 3: Difference of annual mean $\delta^{18}O$ in precipitation between LGM and Control for the GISS and the ECHAM.

A further difference between the two LGM runs is seen in the generally higher δ-values produced by the ECHAM model in the tropics and subtropics (about 1‰ in the zonal mean, see Fig.1c). The ECHAM model is generally more sensitive than the GISS model (as was already suggested by its larger seasonal gradients) to the prescribed SST changes in the tropics. In the ECHAM simulation, the monsoon circulation over both Africa and Asia is strongly diminished during the LGM, and the reduction of precipitation produces considerably higher δ-values than in the control run (up to 3.5‰) due to the isotopic 'amount effect'. In the tropical and subtropical Pacific both models react similarly to the imposed SST changes. Reduced tropical (-2°C) and slightly increased subtropical (1-2°C) temperatures weaken the moisture convergence in the tropics and therefore reduce the precipitation in the ITCZ, enhancing the precipitation in the subtropics. The GISS and the ECHAM models agree in their corresponding isotope changes, each producing less depleted precipitation in the region of the ITCZ and more depleted precipitation in the subtropics (see Fig.3). Since the two LGM simulations differ regionally, much significance should be given to the calculated temporal δ/T_S slopes (see Figure 1d and 4 as well as Table 1) for the regional calibration of the paleothermometer. The slopes vary over the continents between 0.4 and 0.8 ‰/°C, except over East Antarctica, wich shows gradients higher than 1‰/°C. Again, the models tend to simulate higher temporal gradients in the interior of the continents (see Figure 4). As for the seasonal gradient, the water isotopes in the ECHAM model are more sensitive to temperature changes over the continental interior.

Here Figure 4

Figure 4: Slope characterizing the temporal $\delta^{18}O$ /T_S relationship derived from LGM-Control differences (‰/°C) for the GISS and the ECHAM models. The temperature and the $\delta^{18}O$ field have been smoothed prior to the calculation of the slope. Points are only shown with $T_{Ann} < 15°C$ and $\Delta_{LGM\text{-}Control}T < -3°C$.

Although the differences between the temporal and spatial slopes are high in some regions (e.g., a 50% difference in West Antarctica), the overall similarity between the slopes led Jouzel et al. (1994) to suggest that spatial slopes seem on average to be adequate surrogates for temporal slopes. The relative differences over the ice sheets were of order 30% or less (see Table 1). Jouzel et al. (1994) also note that, in mid- and high northern latitudes, the GISS GCM often simulates temporal slopes that are a bit lower than the spatial slopes. They did not, however, infer any conclusion from this bias because it was not seen in all regions (see discussion in Jouzel et al., 1997). This is consistent with the ECHAM results and there is, indeed, no robust tendency for lower temporal than spatial slopes in the model experiments discussed here (In fact, in Antarctica, both models generate temporal slopes that are slightly higher than the spatial, see Table 1). Moreover, some regions are strongly affected by atmospheric circulation changes (e.g., in Asia, the strength of the monsoon is reduced during the LGM). Although, for modern conditions, the isotopes are clearly influenced by the temperature effect, circulation changes lead to a high noise level in the relation between the isotopes and local temperatures (see the regional σ values in Table 1).

A thorough interpretation of the water isotopes in such regions, therefore, should take into account local circulation changes as well.

Regions	Spatial Gradient		LGM-Control Gradient	
	m	**r**	**m**	**s**
Global				
Obs.	0.58	0.9		
GISS	0.59	0.96	0.51	0.34
ECHAM	0.58	0.97	0.46	0.3
N.Amer.				
Obs.	0.56	0.88		
GISS	0.56	0.96	0.43	0.13
ECHAM	0.44	0.87	0.58	0.2
Greenl.				
Obs	0.51	0.84		
Obs.Ice	0.67	0.94		
GISS	0.51	0.86	0.43	0.06
ECHAM	0.51	0.95	0.49	0.08
Europe				
Obs.	0.48	0.73		
GISS	0.55	0.92	0.36	0.05
ECHAM	0.44	0.80	0.31	0.07
Asia				
Obs.	0.47	0.8		
GISS	0.49	0.94	0.44	0.1

ECHAM	0.35	0.9	0.43	0.47
WAnt.				
Obs.				
GISS	0.61	0.93	1.02	0.34
ECHAM	0.7	0.98	0.77	0.15
EAnt.				
Obs.				
GISS	1.0	0.83	1.25	0.42
ECHAM	0.77	0.96	0.88	0.2

Table1: List of simulated and observed spatial and temporal gradients for the 8°x10° GISS model and the ECHAM3 T42 model. m denotes a gradient in ‰/°C, r the corresponding correlation, σ the spatial standard deviation of the corresponding quantity. For all calculations only grid points (or stations) are considered with an annual mean temperature < 15°C. For the temporal gradient (LGM-Today), the analysis has been further limited to grid points with a temperature change of at least -3°C. The observations are from the IAEA/GNIP network (IAEA, 1992) and, for the Greenland ice sheet, from Johnsen et al. (1989) and Hoffmann et al. (1998b).

6. Simulations for Warmer Climates

The advantage of using more than two climates to define the temporal slope and the interest of examining climates similar to those that may someday prevail due to an increase in atmospheric greenhouse gases motivated a recent 2*CO_2 isotopic experiment with the GISS 8° * 10° isotopic model (See Hansen et al., 1984 for a discussion of an analogous simulation with the non-isotopic version of the GCM). The simulated climate in the 2*CO_2 simulation is about 4° C warmer on average than that in the present-day simulation, with higher temperature increases at higher latitudes. As a result, the 2*CO_2 simulation produced isotopically heavier precipitation at high latitudes. Somewhat surprisingly, though, the simulation also produced slightly lighter precipitation in some mid-latitude areas and consistently lighter precipitation in tropical and equatorial regions (Figure 5). In these latter areas, $\delta^{18}O$ in precipitation is decreased by up to 3‰ in association with the aforementioned precipitation 'amount effect'. Some type of "compensation effect" (decreases in tropical $\delta^{18}O$ making up for increases in high latitude $\delta^{18}O$) may be in evidence here.

Here Figure 5

Figure 5: Simulated change of $\delta^{18}O$ in precipitation (2 * CO_2 minus present-day) for the 8° * 10° version of the GISS model.

Spatial and temporal slopes from all three simulated climates are compared in Figure 6. The top part of the figure shows the spatial $\delta^{18}O/T_s$ relationships simulated over Greenland for present-day, LGM, and 2*CO_2 conditions; the spatial slopes are 0.51, 0.76, and 0.73 ‰/°C, respectively. These three slopes are each higher than the temporal slopes computed with the three climates, which range from 0.23 to 0.49 ‰/°C over Greenland, as shown in the lower part of the figure (each plot in the lower part represents a single Greenland grid cell.). These results therefore add some support to the hypothesis that the temporal slope might be generally lower than the measured present-day spatial slope over Greenland. Similar results are seen over East Antarctica, contradicting the aforementioned results obtained with the present-day and LGM simulations alone. Similar results were also produced over some (but not all) sections of northern hemisphere continents.

Here Figure 6

Figure 6: Scatter plots showing the simulated spatial $\delta^{18}O/T_s$ relationships across Greenland for the present-day, LGM, and 2 * CO_2 climates (top 3 plots) and the temporal $\delta^{18}O/T_s$ relationships at each Greenland grid cell (lower 8 plots).

Compared to the strong climatic changes (relative to present-day) associated with 2 * CO_2 conditions or LGM conditions, climatic changes associated with the period around 6 ky BP are fairly small. Nevertheless, a relatively large amount of paleo archives is available bearing information about the isotopic composition of precipitation during this so called mid-holocene optimum. We performed, therefore, AGCM simulations using boundary conditions on 6 kyr BP, which simply amounted to a change in the control simulations' assumed solar insolation. A stronger (weaker) summer (winter) insolation at low northern latitudes was already mentioned as a possible reason for an intensified ocean-land temperature contrast and, consequently, for a stronger summer monsoon circulation (Prell and Kutzbach, 1987).

The response (Figure 7) to this 'weak' forcing is indeed more ambiguous than the previously discussed response to LGM forcing discussed before. Globally, the $\delta^{18}O$ values produced by the GISS model for the mid-holocene optimum never deviate more than 0.7‰ from the control run values. In regions where the temperature effect dominates, the strongest response of the water isotopes is over the central United States and Canada, where an enrichment of $\delta^{18}O$ in precipitation by about 0.4-0.6‰ parallels a warming there of between 0.5 and 1.0 °C. In low latitudes, the amplification of the hydrological cycle (mainly the African monsoon and in the Amazon basin) leads to slightly lower $\delta^{18}O$ due to the amount effect, with a maximum change of -0.5‰. On the other hand, in the ECHAM model the isotopic response is spatially very noisy, in particular in regions of sparse rainfall. In East Antarctica, for example, no regionally consistent response can be defined. The mid-holocene optimum $\delta^{18}O$ values differ from those of the control run by up to ±3‰. The stronger depletion of nearly all tropical and subtropical continental rainfall ranges from -0.5 to -2‰ caused again by the strengthening of the monsoon in Africa and Southeast Asia.

Here Figure 7

Figure 7: Difference of the annual mean $\delta^{18}O$ in precipitation between the Holocene optimum (6kyr BP) and the present-day climate, as simulated by the GISS and the ECHAM models.

Although the mid-holocene optimum results of both models show some similarities, (e.g., the intensification of the hydrological cycle in low latitudes and the resulting increased isotopic depletion in the precipitation there), the weak change in the forcing produces an isotopic response that is more ambiguous and spatially less coherent than induced by the LGM boundary conditions. Furthermore, circulation effects are strongly influencing the water isotopes. The ECHAM model, for example, simulated a warming in Siberia of about 0.5 to 1.5 °C. Nevertheless, only over eastern Siberia does the model calculate positive isotope anomalies (see Figure 7b). It is certainly necessary to further investigate the deviations from a linear isotope-temperature relationship in such regions where otherwise the temperature effect controls the isotopic composition.

7. Deuterium-Excess simulations

Additional information on the water cycle can be obtained from a combination of the two stable water isotopes, deuterium and oxygen 18 through the 'deuterium excess', d. This parameter was defined by Dansgaard (1964) as the deviation from the Meteoric Water Line (Craig, 1961): $d = \delta D - 8\ \delta^{18}O$. The deuterium excess mainly reflects the kinetic fractionation occurring during non-equilibrium processes such as evaporation above the ocean surface (Merlivat and Jouzel, 1979), evaporation of liquid precipitation under the cloud base (Stewart, 1975), and snow formation (Jouzel and Merlivat, 1984). Above the ocean, the deuterium excess in vapor depends on surface parameters. As shown by simple models (Merlivat and Jouzel, 1979 ; Johnsen et al., 1989), the excess in vapor above the ocean surface increases with increasing ocean surface temperature (by about +0.35‰ per °C) and with decreasing relative humidity (by about -0.43‰ per %).

The first deuterium-excess simulation covering a full seasonal cycle was performed with the GISS model (Jouzel et al., 1987). The results were satisfactory as far as global mean annual distribution is concerned. No attempt was made then, however, to compare simulated and observed seasonal distributions of excess, which have very well defined features. This detailed analysis has been performed for the ECHAM simulation of Hoffmann et al., 1998. Globally, the simulated deuterium-excess agrees fairly well with observations, showing a maximum in the interior of Asia and minima in cold marine regions. Over Greenland, the model fails to show the observed seasonality of the excess, but the overall quality of the model is illustrated by the comparison of the simulated and observed deuterium excess distributions (annual mean) over Antarctica (Figure 8). One should note here that simulating this second order parameter correctly is very difficult.

Simulated distributions are very sensitive not only to the parameterization of kinetic effects but also to the transport scheme used in the GCM (Jouzel et al., 1991) and possibly to model resolution, as shown by a comparison between results obtained with the 8 * 10 and 4 * 5 GISS isotopic GCMs (unpublished). This difficulty in modeling deuterium-excess is even seen in the more recent high resolution simulation of Werner et al. (1998), which produces much too large seasonal cycles and unrealistic negative values.

Here Figure 8 : Figure 7 of your JGR paper

Figure 8 : Deuterium excess d versus in precipitation over Antarctica versus the corresponding δD values simulated by the ECHAM3 T42 GCM (Hoffmann et al., 1998a) with observations from Petit et al. (1991)

Simple Rayleigh type models suggest that the information regarding source conditions is at least partly preserved over the air mass trajectory (Johnsen et al., 1989 ; Petit et al., 1991 ; Ciais et al., 1991). This is of interest for paleoclimatologists as it offers the possibility of deriving information about climatic changes in moisture source areas from isotope paleodata (Jouzel et al., 1982 ; Dansgaard et al., 1989 ; Vimeux et al., submitted). Indeed, some recent results obtained with the GISS isotopic GCM support the idea that deuterium excess values contain information on meteorological conditions at distant evaporative sources (Armengaud et al., 1998). As part of this study, a simple isotopic model was initialized with GCM-derived distributions of water isotopes in the vapor, an initialization that is generally not performed correctly by simple isotopic models (Jouzel and Koster, 1996). Using this combined approach, Delmotte et al. (submitted) recently examined how information about source regions can be derived from the seasonal distribution of the deuterium excess (see also Ciais et al., 1995). All these experiments support the idea that variations of the deuterium excess contain information that cannot be derived from either δD or $\delta^{18}O$ alone.

8. Discussion and Conclusion

A growing body of empirical evidence suggests that long-term temporal slopes in polar regions are consistently lower than spatial slopes, particularly for glacial-interglacial changes. This evidence led Jouzel et al. (1997) to examine the influence of various climatic features on the temporal slope. As they point out, simple isotopic models suggest one possibly important factor, namely a simultaneous and parallel change in condensation and evaporative source temperatures between climates (e.g., cool tropics during the LGM, as suggested by Boyle, 1997). Another potentially important factor involves the seasonality of precipitation (Robin, 1983; Steig et al., 1994). If this seasonality varies greatly between climates -- if, for example, a region receives most of its rainfall during summer in one climate and during winter in another -- the relevance of the locally derived spatial isotope/temperature relationship would be severely compromised. The experiments recently performed by Krinner et al. (1997) from experiments using a GCM in which a diagnostic allows access to the mean temperature of snow formation weighted by the amount of precipitation, are interesting in this respect. These authors have shown that the factor of 2 observed for Greenland between spatial and temporal slopes could be explained in their model experiments by changes in local climate parameters (largely by seasonality) in contrast to explanations rather referring to changes in vapour source conditions (lower tropical SSTs during the LGM, see Boyle, 1997). Those same parameters have practically no influence on the glacial-interglacial Antarctic isotope signal, suggesting that the classical use of the spatial slope as a surrogate of the temporal slope could be more appropriate for Antarctic ice cores than for Greenland ice cores. We should note, however, that Charles et al. (1994) found almost no effect of seasonality change on glacial/interglacial isotope differences in Greenland.

Jouzel et al. (1997) cite data gleaned from isotopic GCM simulations (Jouzel et al., 1994; Charles et al., 1994, 1995) to address the relative importance of these factors. They offer several possible explanations for why temporal slopes are lower than spatial slopes over Greenland and for why this discrepancy appears especially large at the glacial/interglacial time scale; these explanations involve, for example, changes in moisture origin, precipitation seasonality, and the strength of the inversion layer. Despite the many difficulties faced in calibrating the isotope paleothermometer, which are mostly related to the unknown quantitative effects of the aforementioned environmental and sampling factors, Jouzel et al. (1997) concluded that the use of a (calibrated) isotope paleothermometer appears justified. The comparison with the ECHAM results presented here (see also Hoffmann et al, 1998b and Hoffmann et al., in press), further supports this conclusion and places this problem in a wider perspective.

In addition to the critically important issue of infering long term (mainly glacial-interglacial) local temperature changes from paleoarchives, we have briefly examined other isotope issues that benefit from the GCM modeling approach. First, as illustrated through the comparison of the ECHAM and GISS model results with data, improved isotopic GCMs (mainly Hoffmann et al., 1998 and Werner et al., 1998 for the ECHAM model but also Andersen, 1997 for the LMD model and Mathieu et al., submitted for the NCAR model can reproduce the main features of the present-day climate's water isotope distributions. The capacity of these isotopic GCMs to reproduce at least part of the short term variability observed in isotopic composition (Hoffmann et al., 1998a : Cole et al., submitted) and to relate it to climate parameters (e.g. temperature and precipitation amount) is now demonstrated with long simulations performed using observed SSTs. Taking advantage of the tagging of source areas implemented in the GISS model, Cole et al. (submitted) stressed the potential impacts of advective processes and of associated changes in the origin of precipitation on the isotope signal. This is important in view of the growing interest of studies dealing with short-term climatic changes. In this respect, Cole et al.(submitted) note that this style of short-term variability differs markedly from that associated, for example, with the cooling during the last ice age. They suggest a continuum of controls on the isotopic content of precipitation in which smaller, advective temperature changes tend to correlate weakly or not at all with the isotopic signal, wheras periods of global temperature change are likely to generate an isotopic signal more consistent with the standard paleotemperature relationship.

Isotopic modeling also seems very promising for addressing the relationship between the deuterium excess of precipitation and climatic parameters, principally at the evaporative source, though we note that model-simulated excess values are generally less accurate than model-simulated δD or $\delta^{18}O$. The relationship established with simple Rayleigh type models (Merlivat and Jouzel, 1979 ; Johnsen et al., 1989) is confirmed by GCM studies (Armengaud et al., 1998 ; Delmotte et al., submitted) which suggest that information about the characteristics of source regions can be extracted from isotope paleo data.

Finally, we would like to point out that isotopic (atmospheric) GCMs are now beeing used to interpret oceanic oxygen 18 data recovered from the analysis of fossil carbonate. Measurements performed on benthic and planktic foraminifera and on corals allow the inference of seawater $\delta^{18}O$ changes provided that temperature changes can be independently estimated (and provided that certain species-dependent effects are taken into account). The $\delta^{18}O$ contents of deep seawater, obtained from benthic foraminifera, allow estimates of the change in global ice volume, whereas the interpretation of surface data (planktic foraminifera and corals) is more complex.

The $\delta^{18}O$ of sea surface water is, in addition, affected by evaporation and precipitation fluxes at the air-sea interface, as well by continental runoff in coastal areas and by sea ice formation and iceberg discharge in polar regions. All of these processes also affect sea surface salinity (SSS), and there is, as a result, a strong relationship between SSS and $\delta^{18}O$, which can be used to reconstruct paleosalinities. The interpretation of the paleo-oceanic data assumes that the well documented present-day $\delta^{18}O$/SSS relationships hold in time throughout the region, i.e. that the spatial and temporal slopes are similar (Duplessy et al., 1991). To assess the validity of this assumption (see discussion in Rohling and Bigg, 1998) various modeling approaches are now being developed that either look at the ocean surface using a very simple 2-box model (Juillet et al., 1997) or involve the incorporation of water isotope cycles into a 3D oceanic model (Schmidt, 1998; Delaygue et al., in preparation). One long term objective of this modeling effort (Schmidt, personal communication) is the full coupling of the atmospheric and oceanic isotopic models. This represents a new and exciting challenge for our scientific community.

Figure captions

Figure 1: Zonal mean of (a) $\delta^{18}O$ in precipitation (control run), (b) the seasonal gradient in ‰ /°C (i.e. the slope of the $\delta^{18}O$ /Temperature relation for the mean 12 months of the control run), (c) the difference of the annual $\delta^{18}O$ between the LGM (corrected for a supposed 1.6 ‰ enrichment of the glacial ocean) and the Control simulation and (d) the temporal (LGM-Control) gradient.

Figure 2: Spatial linear relationship between δ and surface temperature on the Greenland ice sheet (Dansgaard slope), a) observations, b) model results (from Werner et al., 1998).

Figure 3: Difference of annual mean $\delta^{18}O$ in precipitation between LGM and Control for the GISS and the ECHAM.

Figure 4: Slope characterizing the temporal $\delta^{18}O$ /T_S relationship derived from LGM-Control differences (‰/°C) for the GISS and the ECHAM models. The temperature and the $\delta^{18}O$ field have been smoothed prior to the calculation of the slope. Points are only shown with $T_{Ann} < 15°C$ and $\Delta_{LGM\text{-}Control}T < -3°C$.

Figure 5: Simulated change of $\delta^{18}O$ in precipitation (2 * CO_2 minus present-day) for the 8° * 10° version of the GISS model.

Figure 6: Scatter plots showing the simulated spatial $\delta^{18}O/T_S$ relationships across Greenland for the present-day, LGM, and 2 * CO_2 climates (top 3 plots) and the temporal $\delta^{18}O/T_S$ relationships at each Greenland grid cell (lower 8 plots).

Figure 7: Difference of the annual mean $\delta^{18}O$ in precipitation between the Holocene optimum (6kyr BP) and the present-day climate, as simulated by the GISS and the ECHAM models.

Figure 8: Deuterium excess d versus in precipitation over Antarctica versus the corresponding δD values simulated by the ECHAM3 T42 GCM (Hoffmann et al., 1998a) with observations from Petit et al. (1991)

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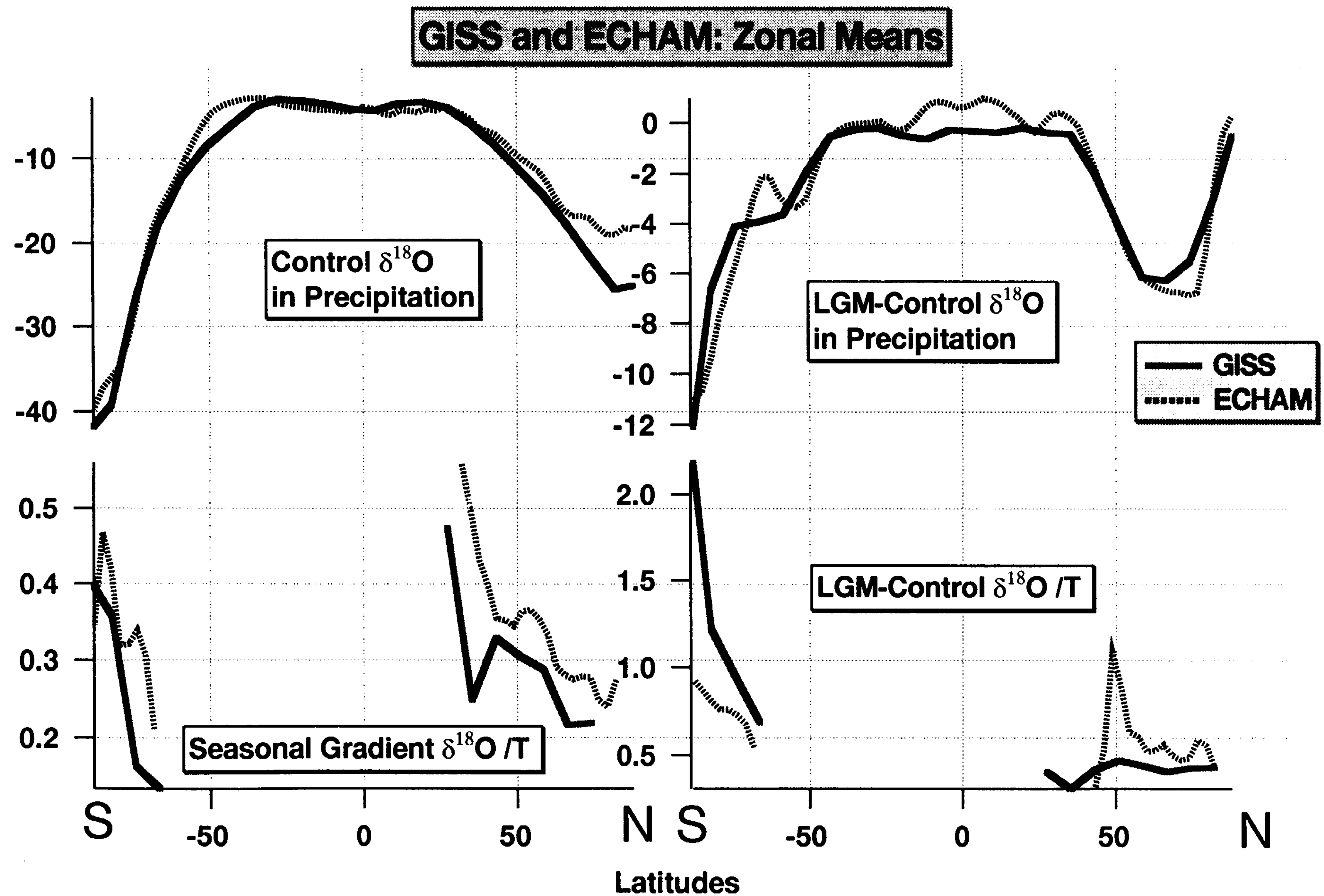


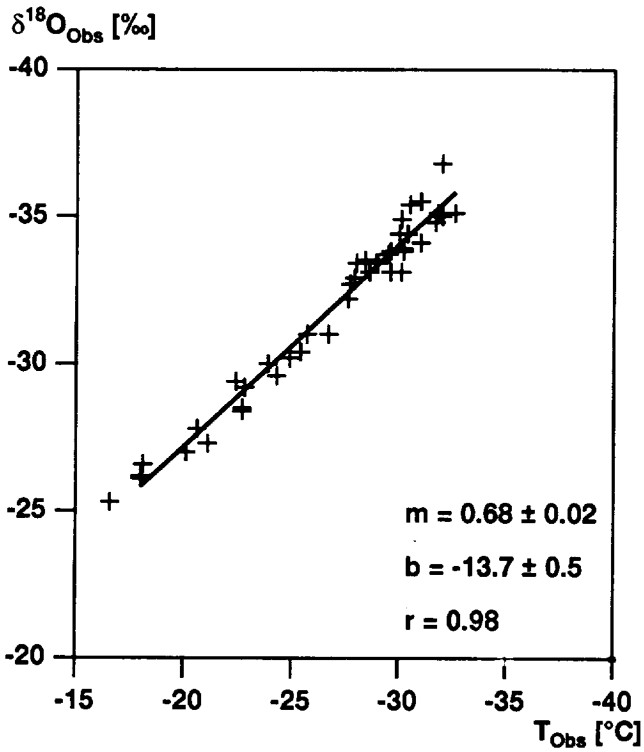
GISS and ECHAM: Zonal Means
Control δ18O in Precipitation
LGM-Control δ18O in Precipitation
GISS
ECHAM
Seasonal Gradient δ18O /T
LGM-Control δ18O /T
-10
-20
-30
-40
0
-2
-4
-6
-8
-10
-12
0.5
0.4
0.3
0.2
2.0
1.5
1.0
0.5
S
-50
0
50
N
Latitudes

Fig 7

SPATIAL TEMPERATURE - $\delta^{18}O$ - RELATION

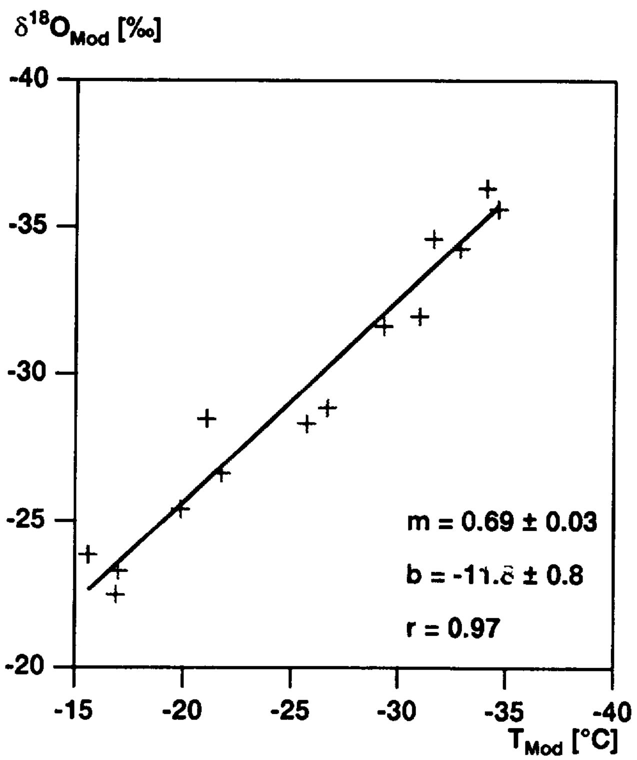
(1) DAANSGAARD (1973): $\delta^{18}O = 0.62\ T - 15.25$ ‰

(2) JOHNSEN (1989): $\delta^{18}O = (0.67 \pm 0.02)\ T - (13.7 \pm 0.5)$ ‰



(3) OBSERVATIONS:

$\delta^{18}O = (0.68 \pm 0.02)\ T - (13.7 \pm 0.5)$ ‰



(4) ECHAM-4 (T42):

$\delta^{18}O = (0.69 \pm 0.03)\ T - (11.8 \pm 0.8)$ ‰

Fig 3

GISS and ECHAM: LGM-180 Anomalies

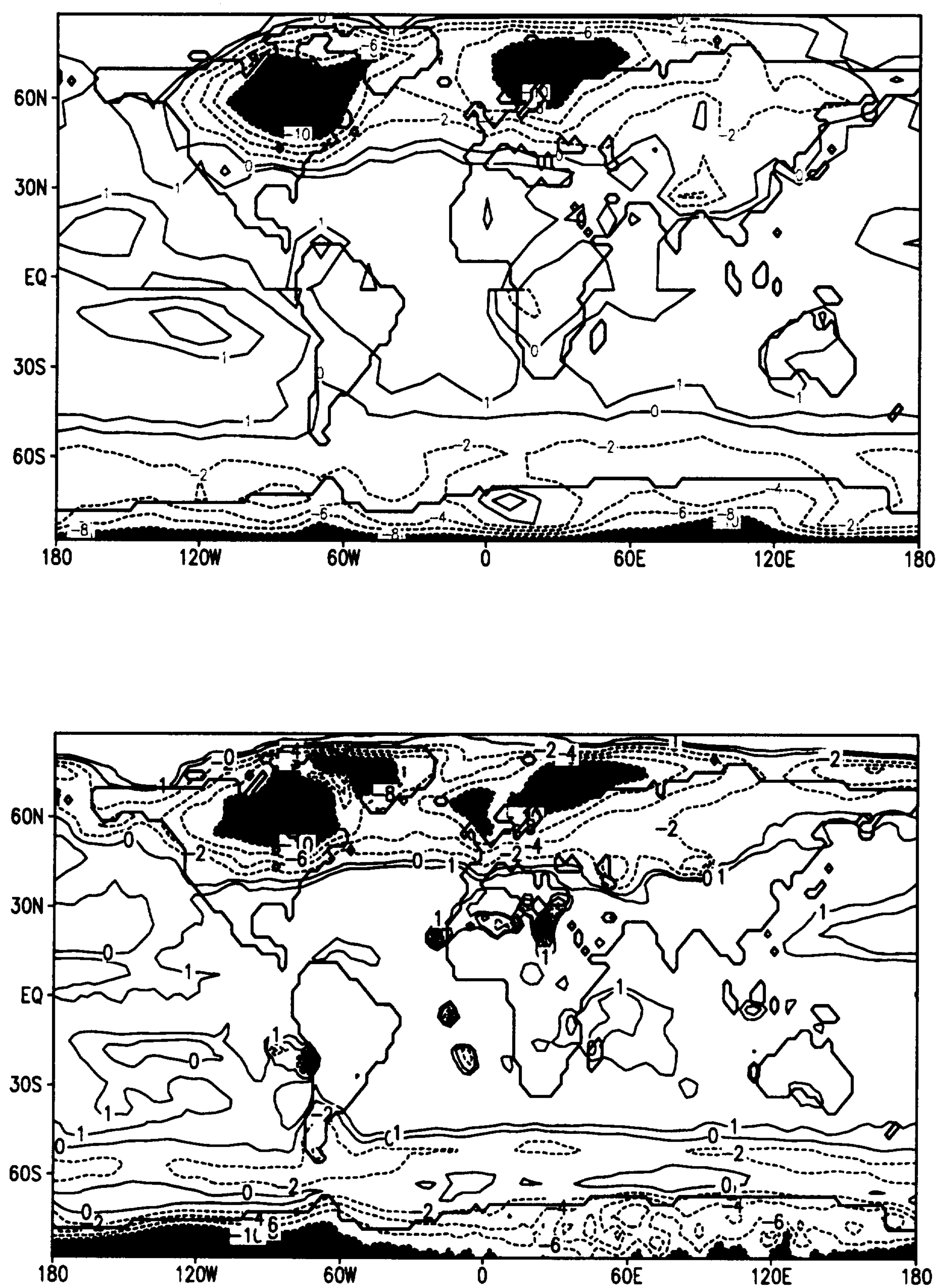


Fig 4

GISS and ECHAM: LGM-Control 18O/T Gradient

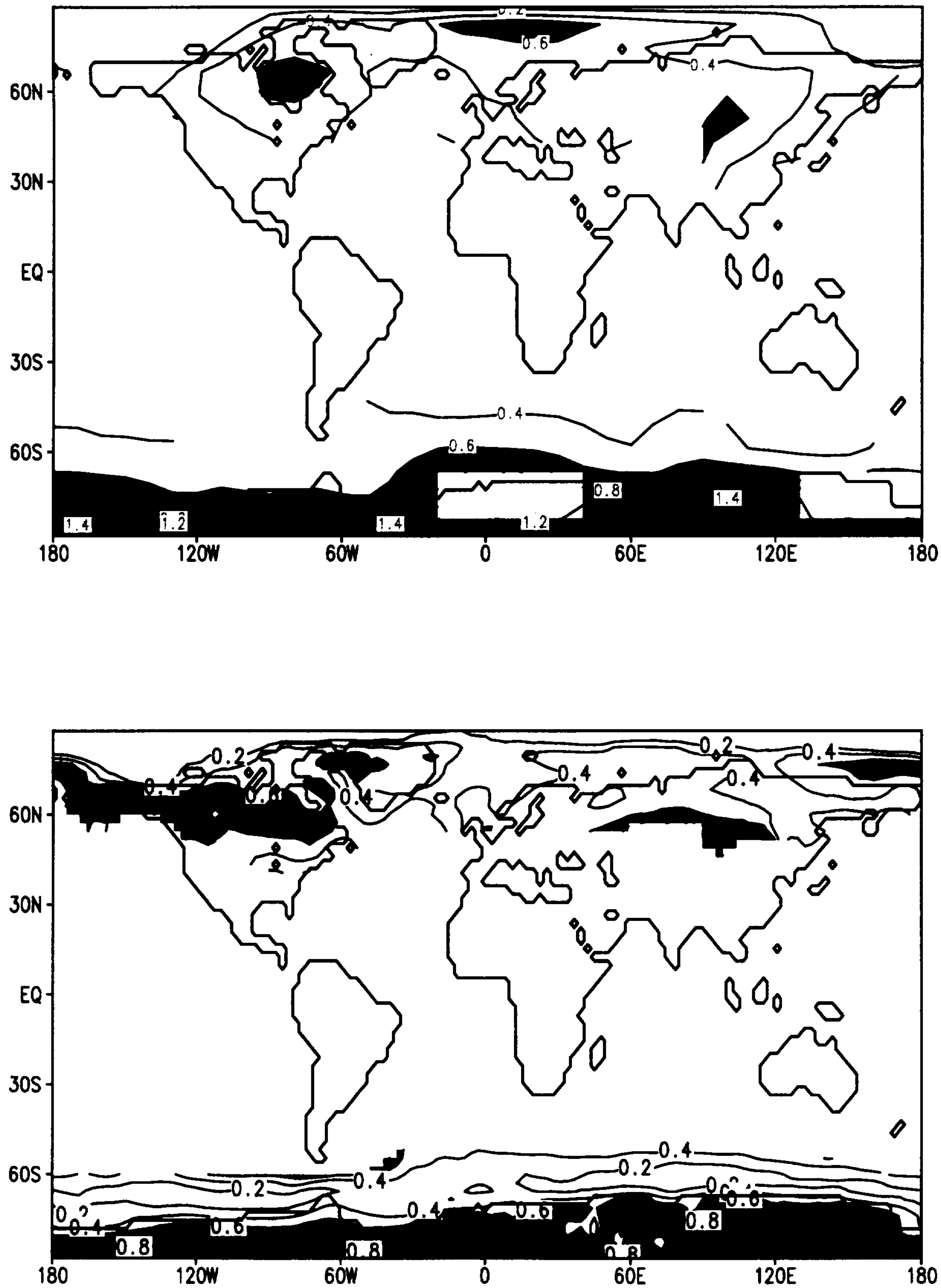


Fig 5

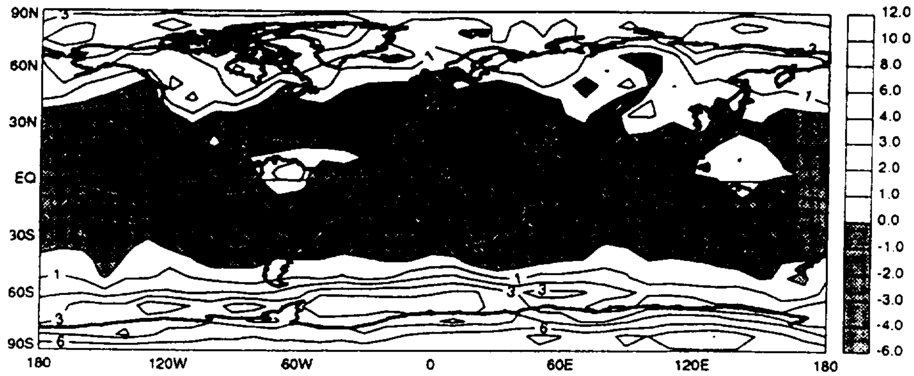

90N
60N
30N
EQ
30S
60S
90S
180
120W
60W
0
60E
120E
180
12.0
10.0
8.0
6.0
4.0
3.0
2.0
1.0
0.0
-1.0
-2.0
-3.0
-4.0
-6.0

Fig 6

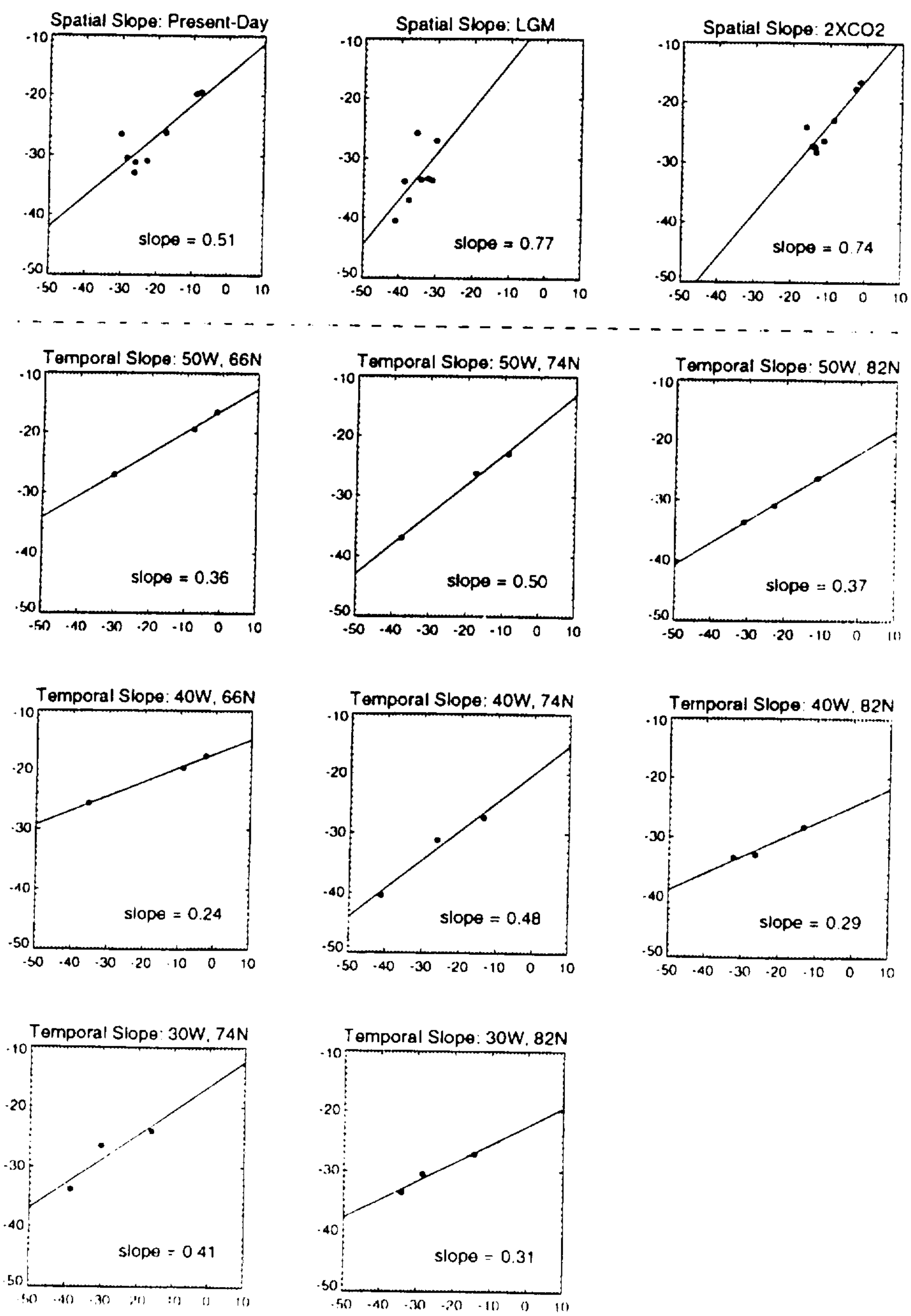

Spatial Slope: Present-Day
slope = 0.51
Spatial Slope: LGM
slope = 0.77
Spatial Slope: 2XCO2
slope = 0.74
Temporal Slope: 50W, 66N
slope = 0.36
Temporal Slope: 50W, 74N
slope = 0.50
Temporal Slope: 50W, 82N
slope = 0.37
Temporal Slope: 40W, 66N
slope = 0.24
Temporal Slope: 40W, 74N
slope = 0.48
Temporal Slope: 40W, 82N
slope = 0.29
Temporal Slope: 30W, 74N
slope = 0.41
Temporal Slope: 30W, 82N
slope = 0.31

Fig 7

GISS and ECHAM: Mid Hol.-180 Anomalies

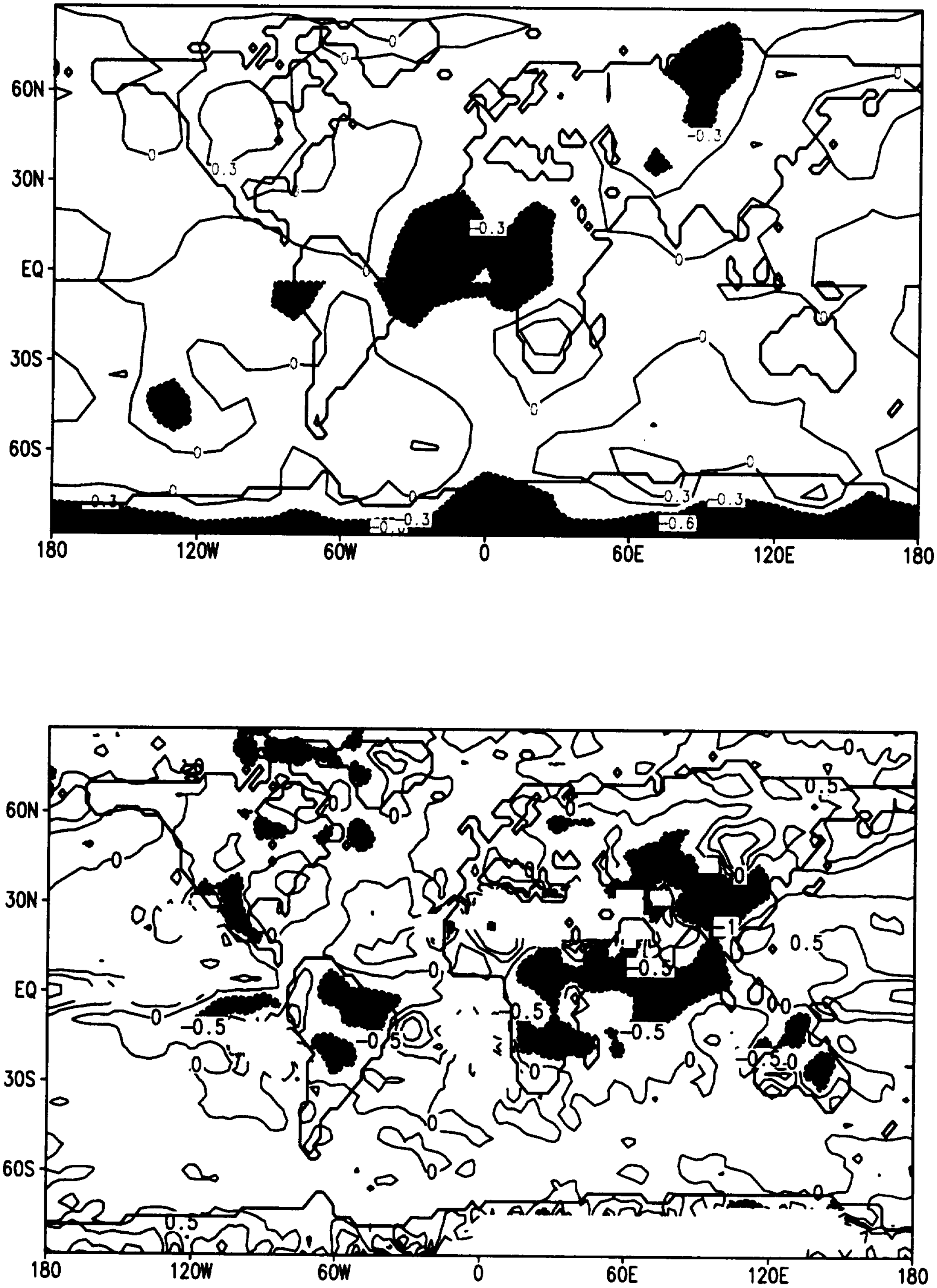
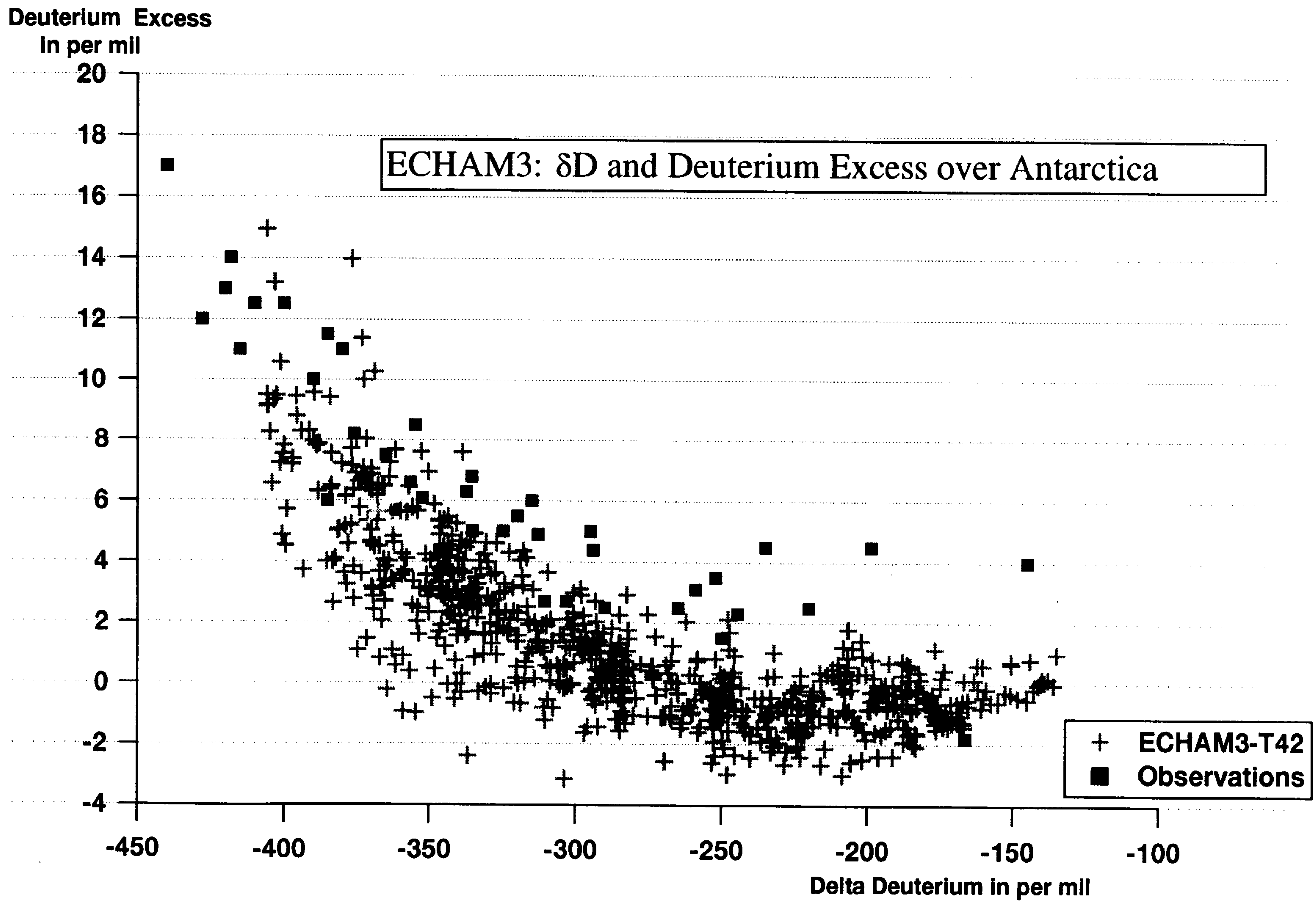


Fig 8



Deuterium Excess in per mil
ECHAM3: δD and Deuterium Excess over Antarctica
20
18
16
14
12
10
8
6
4
2
0
-2
-4
-450
-400
-350
-300
-250
-200
-150
-100
Delta Deuterium in per mil
ECHAM3-T42
Observations