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# Borehole versus isotope temperatures on Greenland: Seasonality does matter

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**Abstract.** New simulation results obtained with the Hamburg Atmosphere General Circulation Model ECHAM-4 under maximum glacial boundary (LGM) conditions confirm the paleotemperatures on Greenland determined by borehole thermometry. The disagreement between $\delta^{18}O$ isotope based temperatures and the borehole temperatures of the LGM is not only reproduced by the model, but the simulation results provide a plausible explanation: Paleotemperatures inferred from $\delta^{18}O$ measurements in ice cores are biased by a substantially increased seasonality of precipitation over Greenland during the LGM. During the glacial winter a much more zonal circulation prevents the effective transport of moisture to the Greenland ice sheet, and therefore reduces the contribution of isotopically strongly depleted winter snow to the annual mean isotope signal.

## Introduction

Since several decades stable water isotopes ($H_2{}^{18}O$, HDO) have been shown to provide a valuable tool for paleoclimate studies [*Dansgaard*, 1964; *Jouzel et al.*, 1987]. To determine past surface temperatures it has been generally assumed that the observed present day spatial relationship between surface temperature ($T_S$) and the isotopic composition of precipitation (usually given as $\delta^{18}O$ or $\delta D$) can be used as an analogue of the temporal $T_S$-$\delta^{18}O$-relation. However, recent isotope independent measurements of paleotemperatures on Greenland by borehole thermometry [*Jouzel*, 1999, and references herein] indicate that the temperature difference at Summit, Central Greenland, between the last glacial maximum (LGM) and present day was in the range of $-23\pm2$ K, twice as large as estimated from $\delta^{18}O$ data using the classical approach. Several hypotheses have been proposed to reconcile this discrepancy and a detailed overview of these hypotheses has been given by Jouzel et al. [1997].

Here, we report the results of a new study, where we have tested all but one of these hypotheses using an atmospheric general circulation model (AGCM) which explicitly models two stable water isotopes ($H_2{}^{18}O$, HDO) in the hydrological cycle. Such an AGCM allows an independent simulation of both quantities $\delta^{18}O$ and $T_S$ [e.g. *Hoffmann et al.*, 1998; *Cole et al.*, 1999]. Hence possible changes of the isotope-temperature-relation in time and space can be explored by using different boundary conditions for AGCM model experiments.

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## Model Experiments

Our results are based on isotope modeling using the Hamburg AGCM ECHAM-4 [*Roeckner et al.*, 1996] with both $H_2{}^{18}O$ and HDO explicitly built into the water cycle of the AGCM [*Hoffmann et al.*, 1998]. All experiments reported here were performed in 3.75° x 3.75° model resolution, each of them running for 10 years with seasonally varying constant boundary conditions. The model includes diagnostic code for tagging water vapor from different source regions. The control experiment was integrated under present-day climate boundary conditions. For the LGM simulation CLIMAP boundary conditions (sea surface temperatures, solar insulation, glacial atmospheric $CO_2$) were prescribed except for the Greenland topography. In agreement with new results of Cuffey and Clow [1997] the glacial Greenland topography change proposed by Peltier [1994] was lowered by three-quarters, yielding an absolute glacial rise at Summit of +200 m compared to present. Additionally, we assumed a slight glacial enrichment ($\delta^{18}O$: +1.5‰, $\delta D$: +12‰) of the heavy water isotopes in the oceans to correct for the isotopically lighter water locked up in glacial ice sheets.

Fourteen different evaporation areas of the water vapor were defined for tagging. Over land, each continent was selected as a distinct source region. For the ocean, annual mean sea surface temperatures (SST) were chosen to define the different evaporation regions of the Polar Seas (SST≤10°C) the Northern Atlantic and Northern Pacific (10°C<SST≤25°C) and the Tropical Atlantic and Tropical Pacific (SST>25°C), respectively. Thus, the ocean source regions of the control experiment and the LGM simulation differed in their geographical position but had the same mean SST range.

In addition to the control experiment and the LGM simulation, we performed two other LGM sensitivity experiments: In the first one we used the Peltier [1994] topography change to evaluate the influence of a higher Greenland ice sheet. In the second sensitivity experiment we investigated the influence of cooler tropical SST during the LGM. Several authors have claimed that the CLIMAP SST reconstruction is too warm for tropical regions. Thus, for the second sensitivity study, we assumed that between 30°S and 30°N SST were at least 5° cooler than present-day SST, but kept the CLIMAP SST if they prescribed an even stronger cooling. Northwards (southwards) of 45°N (45°S) the standard CLIMAP SST were prescribed with a linear transition zone between 30° and 45°.

## Results & Discussion

*Mean state for the present and the LGM climate:* Modeled 10-year-mean values of $T_S$ (-29.4°C), precipitation (22.6cm/y) and $\delta^{18}O$ (-29.5‰) in the grid box enclosing the Summit area are close to present in-situ observations and measurements on ice cores (Table 1). In order to compare mean model values in a consistent way with field data, the modeled $T_S$ and precipi-

**Table 1.** Comparison of In-Situ Measurements and Ice Core Data to Modeled Values for the Present Climate and the Last Glacial Maximum (LGM)

| Climate | Data | $T_S$ (°C) | Prec. (cm/y) | $\delta^{18}O$ (‰) |
|---|---|---|---|---|
| present | Observations | -32 | 23 | -34.8 |
| | Control Experiment | -29.4 ± 1.2 | 22.6 ± 4.3 | -29.5 ± 0.7 |
| LGM | GRIP/GISP2 Estimates | -50 to -55 | 5.5 to 7 | -41 to -43 |
| | LGM Experiment | -52.9 ± 1.3 | 4.5 ± 0.9 | -33.2 ± 1.9 |
| | Sensitivity Study (Peltier topography) | -59.2 ± 1.0 | 2.9 ± 0.7 | -36.7 ± 2.0 |
| $\Delta_{LGM}$ | GRIP/GISP2 Estimates | -18 to -23 | -16 to -18 | -6 to -8 |
| | LGM - Control Exp. | -23.5 ± 2.7 | -18.6 ± 5.2 | -3.7 ± 2.6 |

The ice core data was compiled from *Cuffey and Clow* [1997], *Grootes et al.* [1993], *Johnsen et al.* [1992], *Shuman et al.* [1996].

tation are calculated as standard arithmetic means while the modeled mean $\delta^{18}O$ value is precipitation-weighted

$$\delta^{18}O = \Sigma_i (\delta^{18}O_i \cdot pr_i) / \Sigma_i pr_i$$

based on monthly mean values $\delta^{18}O_i$ and precipitation $pr_i$. The slightly lower model values of $T_S$ and $\delta^{18}O$ as compared to the observations can be explained by model resolution, since the grid box enclosing the Summit area is 500 m lower than the true Summit location. Corresponding ECHAM-4 simulations with a finer spatial grid are in better agreement with the observations. In the LGM experiment $T_S$ (-53°C) and precipitation (4.5cm/y) are also close to the estimates derived from borehole thermometry and ice core data, although the precipitation amount is slightly underestimated. However the mean $\delta^{18}O$ value (-33.2‰) is significantly higher than the ice core data (-41‰ to -43‰) which can partly be explained again by model resolution. Nevertheless the modeled $\delta^{18}O$ anomaly $\Delta_{LGM}$ of the LGM minus the present climate is about 3‰ less than observed as well (Table 1). This shortcoming in the LGM experiment is not fully understood, since the height difference (LGM to present) in the simulation (+200m) is even slightly larger than the estimates of Cuffey and Clow [1997]. It is also obvious from Table 1 that the higher glacial elevation of the Greenland ice sheet proposed by Peltier [1994] results in even lower model values of $T_S$ and precipitation which deviate from the ice core data.

*The seasonal cycle:* In the control experiment, $T_S$ shows a clear seasonal cycle with a minimum of -41±3°C in January and maximum of -14±2°C in July (Fig. 1) which agrees well with observations [*Shuman et al.*, 1996]. Parallel to $T_S$, there is also a strong seasonal amplitude of the modeled $\delta^{18}O$ signal (11.3±4.4‰) which is confirmed by many studies on ice cores [e.g. *Johnsen et al.*, 1989]. In contrast to $T_S$ and $\delta^{18}O$, the modeled precipitation for the present-day climate does not show such a strong seasonal cycle. However the higher simulated values in late summer/early autumn and the small minimum in late winter/early spring have also been reported before [*Bromwich et al.*, 1993]. Under LGM boundary conditions the shape of the seasonal cycle of $T_S$ and $\delta^{18}O$ is almost unchanged. In contrast, the seasonal cycle of precipitation is considerably affected: Modeled LGM winters are very dry with monthly precipitation of less than 1mm/month. Analyses of the geopotential height at 500hPa show that such extremely dry glacial winters are caused by a flow of air masses from more northerly directions compared to the present climate. The advected air masses are substantially colder and dryer, and thus responsible for the aridity and stronger cooling over Greenland in LGM winters as compared to LGM summers.

*Modeled temperature-isotope relations:* The simulated modern spatial isotope-temperature-slope (0.58±0.07, $r^2$=0.77 ±0.08) is close to the observations (0.67±0.02‰/°C) [*Johnsen et al.*, 1989]. For the LGM simulation the spatial slope (0.38 ±0.10‰/°C) is significantly lower and its variance $r^2$ (0.39 ±0.18) larger than for the control experiment (Plate 1, top). For determining the temporal $\delta^{18}O$-$T_S$-relation for the Summit area we correct the LGM $\delta^{18}O$ values for the changed isotope values of the ocean source and then calculate for each combination of the ten control and ten LGM simulation years the temporal slope as $m = \Delta_{LGM}\delta^{18}O / \Delta_{LGM}T_S$. The mean value of the grid box enclosing Summit (0.23±0.08‰/°C) is about 60% smaller than the modeled modern spatial slope, similar to the relationship based on the borehole thermometry measurements. Thus, the observed discrepancy between borehole and isotope temperatures is clearly reproduced in our simulations.

Since the $\delta^{18}O$ signal is temperature dependent but only archived during precipitation events, the isotopic composition is not so much related to the annual mean surface temperature $T_S$ but rather to a precipitation-weighted temperature $T_{S,pr}$

$$T_{S,pr} = \Sigma_i (T_{S,i} \cdot pr_i) / \Sigma_i pr_i$$

where $T_{S,i}$ and $pr_i$ are the temperature and precipitation amount, respectively, at time i [e.g. *Steig et al.*, 1994]. For a yearly uniform distribution of precipitation events the $\delta^{18}O$-

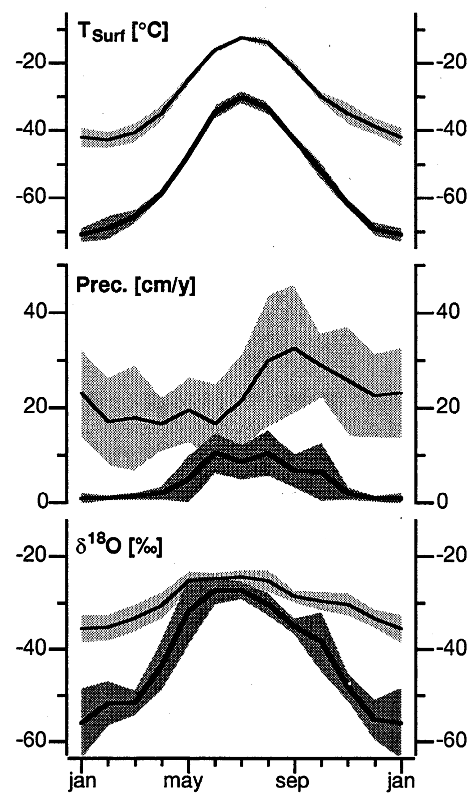


**Figure 1.** Modeled seasonal cycle (solid line) of $T_S$, precipitation and $\delta^{18}O$ and its standard deviation 1σ (gray area) in the grid box enclosing Summit for the present (light gray) and LGM climate (dark gray). For clarity reasons, January values are drawn twice.

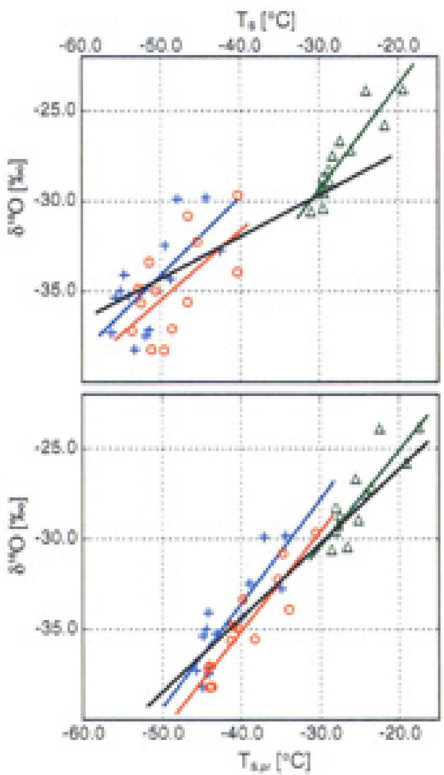


**Plate 1.** Top: Modeled spatial $T_S$-$\delta^{18}$O-relation on Greenland for the present climate (green triangles) and the LGM climate prescribing CLIMAP SST (red circles) or cooler tropical SST (blue crosses). The temporal relation (LGM-present) for the grid box enclosing Summit is drawn in black. Bottom: The same spatial and temporal relations, but for the precipitation-weighted temperature $T_{S,pr}$.

temperature-relation will be quite similar for $T_S$ and $T_{S,pr}$. On the other hand, a strong seasonal cycle of precipitation with less snowfall during winter than during summer will shift $T_{S,pr}$ to warmer temperatures than $T_S$ and thus alter the $\delta^{18}$O-temperature-relation. To quantify this effect for our model results we re-calculate the spatial and temporal slopes for $T_{S,pr}$ using monthly mean values of $T_{S,i}$ and $pr_i$. As expected the spatial slope for the control experiment is similar for $T_S$ and $T_{S,pr}$ (Plate 1, bottom). The spatial LGM slope (0.55±0.06 ‰/°C, $r^2$= 0.80±0.08) computed with $T_{S,pr}$ is now close to the modern value (0.53±0.08‰/°C, $r^2$=0.72±0.16), despite significant lower mean temperatures during the LGM. Due to the warmer LGM $T_{S,pr}$ values, the temporal slope (0.41±0.11 ‰/°C) for the grid point enclosing Summit is now close to both spatial relations, too. Thus, we see in our model results a dominant effect of the changed glacial precipitation cycle explaining the simulated isotope-temperature-relations.

In addition, we have analyzed our simulation results with respect to several other hypotheses proposed for explaining the discrepancy between the temporal and spatial isotope-temperature-relation on Summit.

*Origin of precipitation:* A substantial moisture source change during the LGM could result in an isotopic signal, which is independent of local temperature changes on Greenland [*Charles et al.*, 1994]. The modeled isotopic signatures of the most important source regions for the present climate show variations in the range of -20‰ to -48‰. However a major change of the heterogeneous collection of moisture sources does not occur in the LGM simulation (Table 2). Our findings agree with previous GISS AGCM experiments [*Charles et al.*, 1994].

*Cool tropical SST:* Boyle [1997] proposed that cooler glacial tropical SST might explain the difference in temporal vs. spatial $\delta^{18}$O-$T_S$-slope. Cooling of the initial source of water vapor transported to Greenland shifts the spatial isotope-temperature-relation towards colder temperatures. We calculated the spatial and temporal temperature-isotope-relations on Summit for our second LGM sensitivity experiment with cooler tropical SST. As clearly seen in Plate 1, the hypothesis of Boyle [1997] is correct. Cooler SST shift the glacial temperature-isotope-relation on Greenland, but this effect is small. The seasonality of precipitation is similar to the CLIMAP LGM simulation and the effect of the changed seasonality is dominating the isotope-temperature-slopes.

*Difference in cloud versus surface temperatures:* The temperature directly imprinted in the isotope signal is not the surface temperature but the temperature during formation of precipitation, i.e. the cloud temperature. A shift in the relation between cloud and surface temperatures under a glacial climate could explain the difference between modern spatial and temporal $\delta^{18}$O-$T_S$-relation [*Krinner et al.*, 1997]. We assume as a first guess that most of the precipitation is formed near the warmest tropospheric layer [*Krinner et al.*, 1997], and define the inversion temperature $T_{inv}$ as the temperature of the warmest model layer in the troposhere. The mean inversion strength $T_S$-$T_{inv}$ over Greenland in the LGM simulation is 6.3° larger than in the control experiment. However, the strongest changes are found during the winter season when no precipitation is formed in the LGM simulation. The precipitation-weighted inversion strength $T_{S,pr}$-$T_{inv,pr}$ changes only by 4.2° between present and LGM climate. If we use the estimated inversion temperatures, the temporal slopes become slightly steeper (for $T_{inv}$: 0.32‰/°C, for $T_{inv,pr}$: 0.61‰/°C) but this inversion effect is much smaller than the seasonality effect. These findings agree with results performed with the LMDz model [*Krinner et al.*, 1997].

## Conclusions

To our knowledge, the present ECHAM-4 results are the first isotope AGCM simulations, which clearly reproduce the borehole versus isotope temperature discrepancy. They also suggest that a change in seasonal cycle of precipitation is the

**Table 2.** Relative Contribution (in %) and Mean $\delta^{18}$O Value (in ‰) of Different Vapor Source Regions to the Modeled Precipitation at Summit, Greenland

| Region | Present Prec. (%) | Present $\delta^{18}$O (‰) | LGM Prec. (%) | LGM $\delta^{18}$O (‰) |
|---|---|---|---|---|
| Polar Seas | 15.2 | -19.8 | 12.4 | -20.2 |
| Northern Pacific | 7.9 | -41.1 | 9.2 | -41.0 |
| Northern Atlantic | 27.8 | -26.7 | 26.1 | -25.6 |
| Tropical Pacific | 9.6 | -46.6 | 12.2 | -48.4 |
| Tropical Atlantic | 13.9 | -31.6 | 6.4 | -30.8 |
| North America | 15.3 | -24.9 | 18.0 | -26.5 |
| Eurasia | 6.1 | -31.5 | 11.0 | -32.3 |
| rest | 4.9 | - | 4.7 | - |

most plausible explanation for the disagreement: The extremely dry winters during the LGM lead to a systematic bias of isotope estimated annual mean surface temperatures towards summer values. A change in the inversion strength and/or cooler tropical SST might have altered the temporal isotope-temperature relation, too, but the impact of these effects is much smaller.

How reliable are these new model results? Older isotope AGCM simulations under full LGM conditions did not show a notable change in the seasonality of precipitation [*Charles et al.*, 1995]. However those simulations were not able to clearly reproduce the discrepancy between borehole and isotope temperatures either [*Jouzel et al.*, 1997]. To the contrary, a majority of the AGCMs participating in the PMIP project (8 out of 13) strongly support our findings of a changed seasonality of precipitation under LGM conditions [*Krinner*, 1997]. Similar results are found in two further AGCM studies (no isotopes included) [*Fawcett et al.*, 1997; *Krinner et al.*, 1997].

Clearly, there might also be other (polar) regions and/or past climates where the use of isotope temperatures is affected by a change in the seasonality of precipitation. There is no a priori guarantee that any modern isotope-temperature-relation is appropriate for calculating past temporal temperature variations. Isotope modeling with AGCMs has clearly demonstrated its utility as a tool with which one can infer changes in isotope-temperature-relations for different paleoclimates.

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