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GCM analysis of local influences on ice core δ signals

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Abstract. A high resolution GCM is used to examine the effect of changes in local surface climate parameters on the ice sheets that can influence the interpretation of the isotopic signal of the ice from deep cores. The model suggests that the 10°C difference between the LGM surface temperature deduced from borehole thermometry and that deduced from the water isotope analysis to a great extent may be due to a modification of the precipitation seasonality in central Greenland. For central East Antarctica, the model tends to suggest a weak opposite bias.

Introduction

The linear relationship between annual values of the deuterium or ^{18}O concentrations in polar snow and annual mean temperature at the precipitation site, extremely well obeyed over Greenland [*Johnsen et al.*, 1989] and Antarctica [*Lorius and Merlivat*, 1977], has long been used to reconstruct paleotemperatures from polar ice cores. This approach, in which the present-day spatial isotope vs. surface temperature relationship $\delta = aT_s + b$ defined over a certain region is assumed to hold in time throughout the region, is now being challenged, particularly for Greenland [e.g., *Johnsen et al.*, 1995]. As a result, it is now widely accepted that the Modern - Last Glacial Maximum (LGM, 21 kyr ago) cooling exceeded 20°C in central Greenland rather than ≈10°C as previously thought. *Rommelaere* [1997] has shown that similar borehole thermometry results for Vostok [*Salamatin et al.*, 1997] are subject to large uncertainties because of signal diffusion due to low precipitation rates. Consequently, it is unclear whether the isotopic record at Vostok gives correct information on LGM surface temperatures.

Following *Jouzel et al.* [1997], the various factors possibly influencing the temporal δ vs. T_s relationship are, broadly speaking, the origin, seasonality and intermittency of precipitation as well as the microphysical mechanisms leading to their formation and the difference between cloud and surface temperatures. In this letter, we aim to extract information concerning local influences (e.g., seasonality and intermittency of precipitation) on the δ vs. T_s temporal slope directly from AGCMs experiments.

Methodology

We use the LMDz (where 'LMD' stands for Laboratoire de Météorologie Dynamique, Paris, and 'z' stands for zoom) stretched-grid AGCM to study the precipitation temperature in Greenland and Antarctica for the present-day and LGM conditions at a spatial resolution of 100 km. The model has been specifically adapted for the polar regions by *Krinner et al.* [1997], who demonstrated that the model succeeds well in reproducing the present-day Antarctic climate and in particular the hydrological cycle [*Krinner and Genthon*, 1997]. Separate present-day simulations were carried out over 5 years for the two ice sheets. For the LGM simulations, *CLIMAP* [1981] boundary conditions (sea-surface conditions etc.) were prescribed except for the topography. For Antarctica, a 5-year LGM simulation was run using the *Peltier* [1994] paleo-topography, whereas a 3-year Greenland LGM simulation was run using the output of an ice sheet model [*Ritz et al.*, 1997] for the topography of Greenland. The rationale for using this alternative paleo-topography for Greenland is that the *Peltier* [1994] paleo-topography is much higher in central Greenland than the present-day one, and is thus in contradiction to the much lower central Greenland accumulation rates inferred for the LGM [e.g. *Dahl-Jensen et al.*, 1993]. The simulations, especially the Greenland LGM simulation, are quite short. However, this does not seem to be a problem in the present context, because the results discussed in this letter are also valid for each single year of the simulation. In other words, the model is in steady state.

Simulated present-day and LGM climate of the two ice sheets is quite realistic [*Krinner*, 1997]. Throughout this letter, the different simulations will be referred to as $AA0$, $AA21$, $GL0$, and $GL21$, where the first letters stand for the focused region (GreenLand or AntArctica), and the number indicates the period in kyr b.p.

For each grid point, the *precipitation temperature*

$$T_{pr} = \frac{\int \sum_k (T(k,t)pr(k,t))\,dt}{\int \sum_k (pr(k,t))\,dt} \qquad (1)$$

is calculated, where $T(k,t)$ is the instantaneous temperature at the model layer k and $pr(k,t)$ is the instantaneous rate of precipitation (in mm/day) originating from that level. The temporal integration is carried out at each time step over the whole duration of the model run. T_{pr} is thus the precipitation-weighted temperature of the model layers where the precipitation forms. It represents the local component of the water isotope signal δ.

Classically, the δ thermometer is set up by regressing the present-day precipitation (oxygen or hydrogen) isotope ratio against the measured surface temperature T_s, both in time (seasonal cycle) and in space. This yields an affine global relationship $\delta = aT_s + b$ both for $\delta^{18}O$ and for δD with correlation coefficients close to 0.9 [*Johnsen et al.*, 1989; *Lorius and Merlivat*, 1977].

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We can proceed in an analogous way by regressing T_{pr}, our proxy for δ, against T_s for each ice sheet. The present-day T_s vs. T_{pr} regression can then be applied to the simulated LGM precipitation temperature T_{pr}, yielding a precipitation-temperature-deduced surface temperature $\tilde{T}_s(T_{pr})$, which has to be compared to the simulated surface paleo-temperature. If the regression between T_s and T_{pr}, calculated for the present-day model runs, still holds for the LGM model runs, then $\tilde{T}_s(T_{pr})$ will be close to the simulated surface temperature T_s. Note that this is analogous to using the present-day δ vs. T_s regression to deduce paleo-surface temperatures from the isotopic signal drawn from the ice cores. We can then address the following questions: 1) How does the climate change between the LGM and present-day influence the (local or global) relationship between δ and the surface temperature? 2) Can the apparent discrepancy between the oxygen isotope temperature reconstructions and the borehole temperature measurements at Summit be explained by these changes? 3) If so, what is the relative importance of the changes in inversion strength, precipitation seasonality etc.? 4) Are the simulated changes in the relationship between T_{pr} and T_s similar for Antarctica and Greenland?

Results

The present-day T_s vs. T_{pr} relationships over the two ice sheets exhibit quite high correlation coefficients (r^2= 0.85 and 0.90). As a consequence, the present-day error $\Delta_{pr} = \tilde{T}_s(T_{pr}) - T_s$, i.e., the difference between the surface temperature estimate $\tilde{T}_s(T_{pr})$, obtained from the present-day T_{pr} by using the present-day T_s vs. T_{pr} regression, and the simulated present-day surface temperature T_s, is generally very small over both ice sheets (not shown).

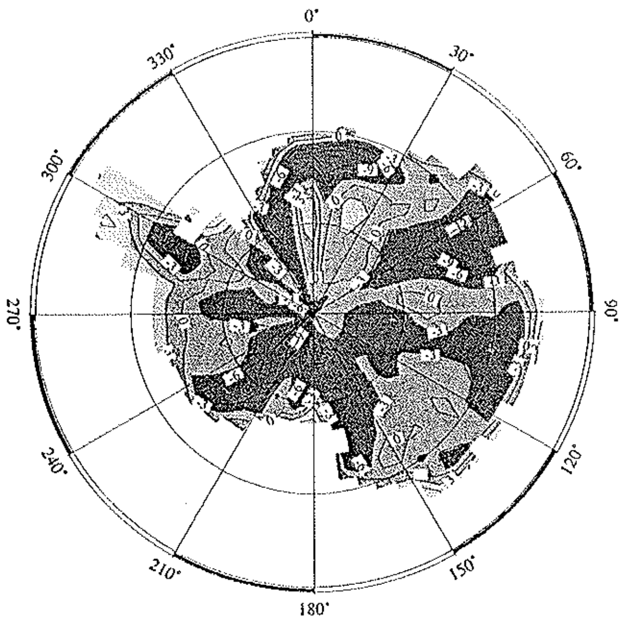

Figure 1. The LGM surface temperature estimation error Δ_{pr} (in °C) over Antarctica The contour interval is 3°C. Dark gray shading indicates values below −3°C, medium gray from −3°C to +3°C, and light gray shading indicates values above +3°C.

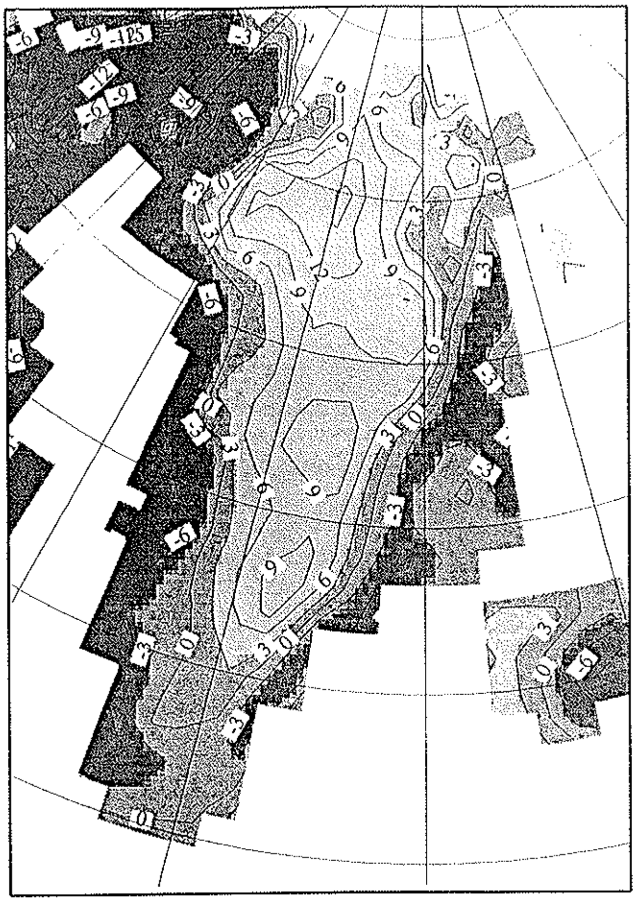

Figure 2. The LGM surface temperature estimation error Δ_{pr} (in °C) over Greenland. The contour and shading intervals are the same as in figure 1.

Figure 1 displays the LGM error $\Delta_{pr} = \tilde{T}_s(T_{pr}) - T_s$ over Antarctica, where $\tilde{T}_s(T_{pr})$ was calculated using the present-day Antarctic regression applied to the LGM T_{pr}. Δ_{pr} tends to be slightly negative in the largest part of East Antarctica. This means that $\tilde{T}_s(T_{pr})$ tends to be somewhat lower than T_s. However, the signal is somewhat spatially inhomogeneous over the East Antarctic Plateau area.

Figure 2 displays Δ_{pr} over Greenland for the *GL*21 simulation. A clear significant signal is visible; its strength increases with altitude (i.e., decreases with temperature). The error attains its maximum in the northern part of Greenland and a secondary maximum of about 10°C in the Summit area. This means that at Summit, $\tilde{T}_s(T_{pr})$ overestimates the simulated surface temperature by that amount.

Discussion and Conclusion

Following *Jouzel et al.* [1997], one can imagine several causes for a change in the T_s vs. T_{pr} relationship:

• The precipitation primarily forms near the warmest tropospheric layer [*Bromwich*, 1988]. Both in reality [*Phillpot and Zillman*, 1970] and in the LMDz GCM, the inversion strength over the ice sheets is well linearly correlated to the surface temperature. If this relationship changes, then the T_s vs. T_{pr} relationship is also likely to change. We thus regress the present-day T_{pr} against the annual mean inversion tem-

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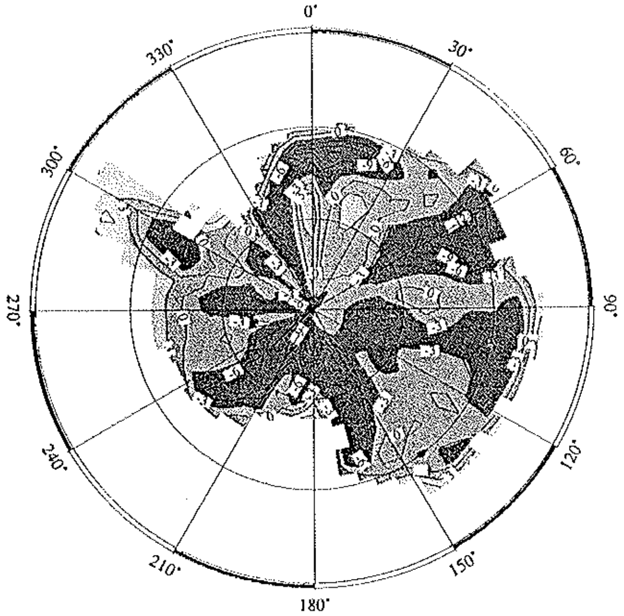

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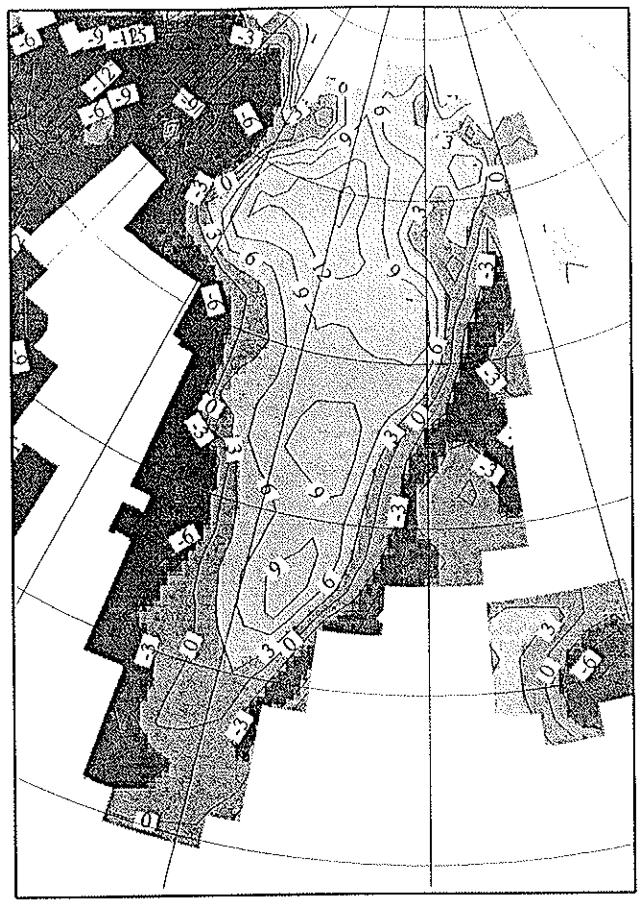

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• A second possibility is a change in the precipitation seasonality. If, for example, precipitation decreases relatively more strongly in winter than in summer in a colder climate, the isotopic signal is likely to carry a warm (summer) bias, leading to an overestimation of the paleo-surface temperatures. In order to estimate the effect of the change in the precipitation seasonality, one can proceed in the same way as above by using the precipitation-weighted inversion temperature

$$T_{inv,pr} = \frac{\sum \overline{T}_{inv}\,\overline{pr}}{\sum \overline{pr}} \qquad (2)$$

instead of the annual mean inversion temperature. $T_{inv,pr}$ is an intermediate diagnostic variable between T_{inv} and T_{pr}. For calculating $T_{inv,pr}$, the monthly precipitation quantities $\overline{pr}$ and inversion layer temperatures $\overline{T}_{inv}$ are used, thus isolating the effect of the seasonality of the precipitation, but not taking into account short-term variability. As above, regressing T_s against $T_{inv,pr}$ for the present and using this regression for calculating an estimated paleo-surface temperature $\tilde{T}_s(T_{inv,pr})$ and an estimation error $\Delta_{inv,pr} = \tilde{T}_s(T_{inv,pr}) - T_s$ allows to evaluate the importance of this effect.

• A third possibility is a change in the short-term correlation (on a time scale of the order of a few days) between temperature and precipitation. High precipitation events on the ice sheets are mostly due to intrusions of warm air which cools down during its ascent and thus loses its moisture-holding capacity. Strong precipitation events are thus generally characterized by a positive deviation of the atmospheric temperature compared to the longer-term (e.g., monthly) mean temperature. A change in this short-term precipitation - temperature covariance can cause the residual error which is not explained by the two effects discussed above.

Table 1 displays the simulated mean surface temperature T_s for the higher parts of both ice sheets (above 3000m a.s.l.) and for the present-day and the LGM. Additionally, it shows the errors Δ_{pr}, Δ_{inv}, and $\Delta_{inv,pr}$ discussed above. As the regressions were calculated from the present-day conditions, the errors for the present-day estimations are generally low (mostly below 1°C). Note that the correlation coefficients r^2 of the regressions discussed in this section are all above 0.85, and mostly above 0.90.

The LGM error Δ_{inv} is quite low because for the higher parts of the ice sheets, the linear T_s vs. T_{inv} relationship, when calculated for the whole ice sheet, does not change dramatically [*Krinner*, 1997].

Using the precipitation-weighted inversion temperature $T_{inv,pr}$ yields an error $\Delta_{inv,pr}$ of about 7°C in the estimation of the LGM surface temperature in the Summit area. Thus, the largest part (about 70%) of the total Δ_{pr} of about 10°C seems to be caused by a change in the precipitation seasonality. Indeed, the LMDz GCM simulates an important change in the central Greenland precipitation seasonality. In accordance with observations reported by *Shuman et al.* [1995], the model does not simulate a clear seasonality of present-day precipitation in the higher regions of the Greenland ice sheet. During the LGM, however, the model simulates a clear summer precipitation maximum on the Greenland plateau areas, with very low accumulation rates in winter. This is due to the fact that in the LGM simulation, the winter cyclones advect much less humidity to Greenland than today because of the southward shift of the sea ice margin which inhibits evaporation over the North Atlantic. Thus, according to the model, the annual-mean isotopic composition of precipitation, which has little seasonal bias at the present, would carry a strong warm (summer) signature during the LGM. On the East Antarctic Plateau, however, the annual cycle does not change that drastically; the model simulates the precipitation maximum in autumn and lower accumulation rates in summer for both climatic periods. The isotopical seasonal bias is thus essentially unaffected in that region.

Table 1. T_s and the regression errors Δ_{pr}, $\Delta_{inv,pr}$, and Δ_{inv} (in °C, explanations see text) on the higher parts of Greenland and Antarctica (mean values for all grid points above 3000m a.s.l.)

run	T_s	Δ_{pr}	$\Delta_{inv,pr}$	Δ_{inv}
AA0	-54.9	0.8	0.2	-0.2
AA21	-60.7	1.9	0.8	1.4
GL0	-35.8	1.3	0.9	0.4
GL21	-51.8	9.8	6.9	0.8

Some experimental support for the simulated change in the annual precipitation cycle over central Greenland comes from work by *Raynaud et al.* [1997], who state that the magnitude of close-off porosity changes observed ice from the GRIP core could partially be explained by changes of the annual cycle of snow deposition.

Note that the role of seasonality for explaining GRIP and GISP2 isotopic data, pinpointed by *Steig et al.* [1995], may be model dependent. For example, it does not appear to be supported by the NASA/GISS model experiments [*Charles et al.*, 1994]. On the other hand, many of the models participating in the Paleoclimate Modeling Intercomparison Project (PMIP, *Joussaume and Taylor* [1995]) do show similar changes in the Greenland precipitation seasonality [*Krinner*, 1997]. Furthermore, the results depend on the imposed sea surface boundary conditions. Simulations with dramatically reduced LGM winter sea ice extents, as suggested by *Weinelt et al.* [1996], do not yield the same results.

As a whole, the LMDz model provides, principally through changes in precipitation seasonality, a plausible explanation for the ≈10°C $\delta^{18}O$ − borehole thermometry conflict.

We obviously keep in mind that our study deals with only part of the factors which influence the δ vs. T_s relationship. In particular, we do not account for possible changes in the origin of the precipitation, the importance of which has been illustrated by *Boyle* [1997]. One way to separate the role of, for example, seasonality and source conditions would be to incorporate the precipitation temperature diagnostic in isotopic GCMs and to run them for different LGM oceanic temperatures. This can be done with existing isotopic GCMs and hopefully in the near future with an isotopic version of the LMDz model we have used in this study.

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References

Boyle, E.A., Cool tropical temperatures shift the global $\delta^{18}O$ T relationship, *Geophys. Res. Lett., 24,* 273 - 276, 1997.

Charles, C., D. Rind, J. Jouzel, R. Koster and R. Fairbanks, Glacial-interglacial changes in moisture sources for Greenland: Influences on the ice core record of climate, *Science, 263,* 508-511, 1994.

CLIMAP, Seasonal reconstruction of the earth's surface at the last glacial maximum, *Geological Society of America, Map chart series MC-36,* 1981.

Dahl-Jensen, D., S.J. Johnsen, C.U. Hammer, H.B. Clauen, and J. Jouzel, Past accumulation rates derived from observed annual layers in the GRIP ice core from Summit, Central Greenland, *Ice in the Climate System*, NATO ASI Series, I 12, ed. W.R. Peltier, 1993.

Johnsen, S.J., W. Dansgaard, and J.W. White, The origin of Arctic precipitation under present and glacial conditions, *Tellus, 41,* 452-469, 1989.

Johnsen, S.J., D. Dahl-Jensen, W. Dansgaard and N.S. Gundestrup, Greenland temperatures derived from GRIP bore hole temperature and ice core isotope profiles, *Tellus, 47b,* 624 - 629, 1995.

Joussaume, S. and Taylor, K. E., Status of the Paleoclimate Modeling Intercomparison Project (PMIP), in *Proceedings of the first international AMIP scientific conference*, 425-430, 1995.

Jouzel, J., R.B. Alley, K.M. Cuffey, W. Dansgaard, P. Grootes, G. Hoffmann, S.J. Johnsen, R.D. Koster, D. Peel, C.A. Shumann, M. Stievanard, and J. White, Validity of the temperature reconstruction from water isotopes in ice cores, *J. Geophys. Res.*, in press.

Jouzel, J., C. Lorius, J.R. Petit, C. Genthon, N.I. Barkov, V.M. Kotlyakov, and V.M. Petrov, Vostok ice core: a continuous isotope temperature record over the last climatic cycle (160,000 years), *Nature, 329,* 402-408, 1987.

Lorius, C., and L. Merlivat, Distribution of mean surface stable isotope values in East Antarctica. Observed changes with depth in a coastal area, in *Isotopes and impurities in snow and ice.* Proceedings of the Grenoble Symposium Aug./Sep. 1975, edited by IAHS, Vienna, 1977.

Krinner, G., C. Genthon, L. Li, and P. Le Van, Studies of the Antarctic climate with a stretched-grid GCM, *J. Geophys. Res., 102,* 13731-13745, 1997.

Krinner, G., and C. Genthon, The Antarctic surface mass balance in a stretched-grid general circulation model, *Ann. Glaciol.*, in press, 1997.

Krinner, G., *Simulations du climat des calottes de glace*, Ph.D. thesis, Université Grenoble 1, France, 1997 (partially in French).

Peltier, W.R., Ice age paleotopography, *Science, 265,* 195-201, 1994.

Phillpot, H.R., and J.W. Zillman, The surface temperature inversion over the Antarctic continent, *J. Geophys. Res., 75,* 4161-4169, 1970.

Raynaud, D., J. Chappellaz, C. Ritz, and P. Martinerie, Air content along the GRIP core: a record of surface climatic parameters and elevation in central Greenland, *J. Geophys. Res.,* in press.

Rommelaere, V., *Trois problèmes inverses en Glaciologie*, Ph.D. thesis, Université Grenoble 1, 1997 (in French).

Salamatin, A.N., V.Y. Lipenkov, N.I. Barkov, J. Jouzel, J.R. Petit, and D. Raynaud, Ice core age dating and paleothermometer calibration on the basis of isotopes and temperature profiles from deep boreholes at Vostok station (East Antarctica), *J. Geophys. Res.,* in press.

Shuman, C.A., R.B. Alley, S. Anandakrishnan, J.W.C White, P.M. Grootes, and C.R. Stearns, Temperature and accumulation at the Greenland Summit: Comparison of high-resolution isotope profiles and satellite passive microwave brightness temperature trends, *J. Geophys. Res., 100,* 9165-9177, 1995.

Steig, E.J., P.M. Grootes, and M. Stuiver, Seasonal precipitation timing and ice core records, *Science, 266,* 1885-1886, 1994.

Weinelt, M., M. Sarnthein, U. Pflaumann, H. Schultz, S. Jung, and H. Erlenkeuser, Ice-free nordic seas during the last glacial maximum? Potential sites of deepwater formation, *Paleoclimates, 1,* 283-309, 1996.

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