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Tropical cooling and the isotopic composition of precipitation in general circulation model simulations of the ice age climate

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Abstract We test the climate effects of changes in the tropical ocean by imposing three different patterns of tropical SSTs in ice age general circulation model simulations that include water source tracers and water isotope tracers. The continental air temperature and hydrological cycle response in these simulations is substantial and should be directly comparable to the paleoclimatic record. With tropical cooling imposed, there is a strong temperature response in mid- to high-latitudes resulting from changes in sea ice and disturbance of the planetary waves; the results suggest that tropical/subtropical ocean cooling leads to significant dynamical and radiative feedbacks that might amplify ice age cycles. The isotopes in precipitation generally follow the temperature response at higher latitudes, but regional $\delta^{18}O$/air temperature scaling factors differ greatly among the experiments. In low-latitudes, continental surface temperatures decrease congruently with the adjacent SSTs in the cooling experiments. Assuming CLIMAP SSTs, $^{18}O/^{16}O$ ratios in low-latitude precipitation show no change from modern values. However, the experiments with additional cooling of SSTs produce much lower tropical continental $\delta^{18}O$ values, and these low values result primarily from an enhanced recycling of continental moisture (as marine evaporation is reduced). The water isotopes are especially sensitive to continental aridity, suggesting that they represent an effective tracer of the extent of tropical cooling and drying. Only one of the tropical cooling simulations produces generalized low-latitude aridity. These results demonstrate that the geographic pattern of cooling is most critical for promoting much drier continents, and they underscore the need for accurate reconstructions of SST gradients in the ice age ocean.

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1 Introduction

The behavior of the tropical ocean over glacial cycles has been a subject of controversy for the last thirty years, but recent discoveries and modelling efforts directed toward simulating the ice age Earth have heightened the intensity of the debate. Since their creation, the CLIMAP (1981) reconstructions have remained the standard boundary conditions for atmospheric general circulation model simulations of the ice age climate (e.g., Hansen et al. 1984) and, more recently, they have served as a target for comparison to the results of coupled ocean-atmosphere models (e.g., Ganopolski et al. 1998; Weaver et al. 1998). Some lines of independent evidence from deep sea sediments (e.g., Rostek et al. 1993) seem to support the CLIMAP reconstructions of relatively invariant tropical temperatures over ice age cycles. However, other lines of evidence from coral reef and continental archives conflict with the inference of little tropical temperature change (Guilderson et al. 1994; Clapperton 1993; Stute et al. 1995; Thompson et al. 1995) and apparently demand much cooler tropical sea surface temperatures (as much as 5–6 °C) during the last glacial period. Of course, successful application of modelling approaches for understanding the ice age cycles, and therefore, this potential gauge of the overall sensitivity of the climate system, depend almost entirely on the accuracy of such inferences from the geological record.

Even if a preponderance of evidence now favors generalized tropical cooling during the ice age (Broecker

1995), the mechanisms establishing these cooler conditions and the global climate effects ultimately engendered are not at all obvious. Here we describe the results of atmospheric GCM simulations designed to attack both of these issues. We performed three multi-year simulations of the last ice age climate using different specifications for tropical sea surface temperatures. These three experiments essentially constitute a sensitivity test of ocean dynamical mechanisms for lowering tropical SST. The physical reality of these mechanisms may be assessed, in part, by how well the results conform to the paleoclimatic evidence. However, the main purpose of these experiments is not necessarily to simulate specific aspects of the ice age world; the SST patterns that we impose are perhaps too arbitrary for that goal. Instead, the objective here is to consider several general consequences of significant tropical SST variability over geological time scales.

One would predict substantial changes in many aspects of the global climate system with alteration of tropical SSTs (see e.g., Webb et al. 1997; Hostetler and Mix 1999), but we concentrate here on the hydrological cycle processes most likely to leave their signature in the geological record. In particular, the model we used incorporates both water isotope tracers ($H_2{}^{18}O$ and HDO) and moisture source tracers (Jouzel et al. 1987). Unlike other paleoclimatic variables that rely on empirically derived proxies for their reconstruction, the isotope tracers are observed directly in the geological record and therefore offer the possibility of explicit comparisons between results of the GCM sensitivity experiments and measurements from ice cores (both polar and tropical) and continental aquifers. While it is often difficult to separate the various possible influences on water isotopic variability, especially on a global scale, the moisture source tracers help with this problem; they provide a more complete picture of the changes in the model's hydrological cycle that ultimately produced the isotopic changes.

These two different types of water tracers (isotopic and source) combine to suggest some possible interpretations for the ice age ice core record, particularly in low latitudes. They also make well-defined predictions for other low-latitude regions that serve to test the validity of the lowered SSTs imposed, as well as the model's response.

2 Model experiments

The water tracers described here were coupled to a version of the $4 \times 5°$ resolution, nine vertical layer GISS II GCM (fully described in Hansen et al. 1983) that uses fixed monthly SSTs. The isotope tracer component of this model applies the appropriate fractionation equations at every change of phase for water and keeps track of the total mass of various isotopes in all potential reservoirs (Jouzel et al. 1987). Though there are variety of more recent models that incorporate water isotopes in analogous ways (e.g., Cole et al. 1999), we believe that this version of the GISS model is sufficient for addressing many first order questions of isotope paleoclimatology. For example, the model run with modern boundary conditions produces a reasonably accurate reflection of the observed distribution and seasonal cycles of $\delta^{18}O$ and δD in low- to mid-latitudes, with a performance similar to that in the lower resolution $8 \times 10°$ model (Jouzel et al. 1987). Two deficiencies of the $4 \times 5°$ resolution isotope model are that mean annual central Greenland precipitation is isotopically heavier than observed, despite a well-defined seasonal cycle (the model predicts an annual average $\delta^{18}O$ of $-30‰$, and $-34‰$ is observed), while central Antarctic precipitation is isotopically lighter than observed ($-60‰$ predicted versus $-55‰$ observed for $\delta^{18}O$). These problems could either arise from a variety of parametrized isotope effects (Jouzel et al. 1991) or from inaccuracies in the basic hydrological cycle in the model (Rind 1984). With the limitations in mind, we will comment briefly on the isotopic behavior over the poles, but the bulk of the discussion will be confined to lower latitude regions.

Water molecules evaporating from a specified set of grid boxes can also be followed as tracers until they precipitate, and this tagging procedure allows determination of the sources of local precipitation in the model (Koster et al. 1986). In the experiments here, we define moisture source tracers on the basis of temperature. For example, water molecules evaporating from oceanic grid boxes with a surface temperature of 24–27 °C are followed as one tracer; molecules evaporating from boxes of 21–24 °C are followed as another tracer, and so on. We also distinguish between water evaporating from continental or oceanic gridboxes.

The standard ice age simulation (designated CLIM) employs CLIMAP sea surface temperatures, as well as other CLIMAP boundary conditions (ice sheet height, sea-ice extent, topography, etc.). The model also uses the orbital parameters and CO_2 concentrations existing 18 000 years ago. The model does not include interactive vegetation and therefore does not capture the possible land surface feedbacks that other model results suggest could be important for low-latitude continents during the ice age (e.g., Hostetler and Mix 1999). Consideration of such feedbacks is an obvious step beyond the initial analysis of low-latitude hydrological cycle changes presented here.

The ice age experiments were integrated for five years, and the results shown here are the average of the last three years. Standard deviations for many of the model fields in the control run (CON) were presented and discussed in Rind (1988). Long integrations necessary for establishing meaningful standard deviations for all of the moisture source tracers were not performed (for reasons of expense), but standard deviations for the isotope tracers may be found in Cole et al. (1999). In general, we reserve the designation "significant" for changes that exceed twice the standard deviation of the modern control run.

For one experiment (denoted WP for the "warm pool" simulation), we arbitrarily lowered the CLIMAP-estimated temperature of ocean "warm pool" regions by 5 °C. To avoid unrealistically sharp discontinuities, we altered CLIMAP's SST estimates with a tapered cooling: estimates above 27 °C were lowered by 5 °C; estimates which fell below 19 °C were unmodified; any estimates falling between 27 and 19 °C were lowered according to a linear interpolation between the 5 °C maximum and 0 °C minimum cooling. This pattern of cooling virtually eliminates the warm pool regions and effectively minimizes zonal gradients along the equator. Although the rationale for this experiment was to isolate the influence of changing the warmest SSTs, the combination of some geochemical observations (corals and alkenones) prior to the Holocene (Beck et al. 1997; Bard et al. 1997) could actually yield this type of pattern. Thus, this experiment could have some basis in reality, even though we did not design it with particular observations in mind. This experiment might be viewed as a permanent and pervasive equatorial cold state that could conceivably arise through, for example, substantial cooling of the tropical thermocline and attendant feedbacks (Bush and Philander 1998).

For the other experiment (designated MHT for "modern heat transport" simulation), we specified a sea surface temperature field predicted from previous experiments by Webb et al. (1997), where modern model ocean heat transports were imposed in an otherwise "normal" ice age simulation. Such model-derived ocean transports in an ice age climate lead to greater oceanic heat divergence from the tropics and subtropics than assumed by CLIMAP, and as a result, tropical and subtropical SST's were also substantially lower in this case (as indicated by the surface air temperatures shown in Fig. 1). In some respects, this experiment might be viewed as describing the thermohaline "on" condition of a typical Dansgaard-Oeschger cycle (Webb et al. 1997).

The pattern of cooling imposed in MHT was less arbitrary, in the sense that it arose from a specific physical process in the model. However, comparing WP and MHT, the main difference for the low- to mid-latitudes is the stronger cooling of both the Northern and Southern Hemisphere subtropics in MHT. The northern North Atlantic represents an exception: SSTs in this region for MHT are

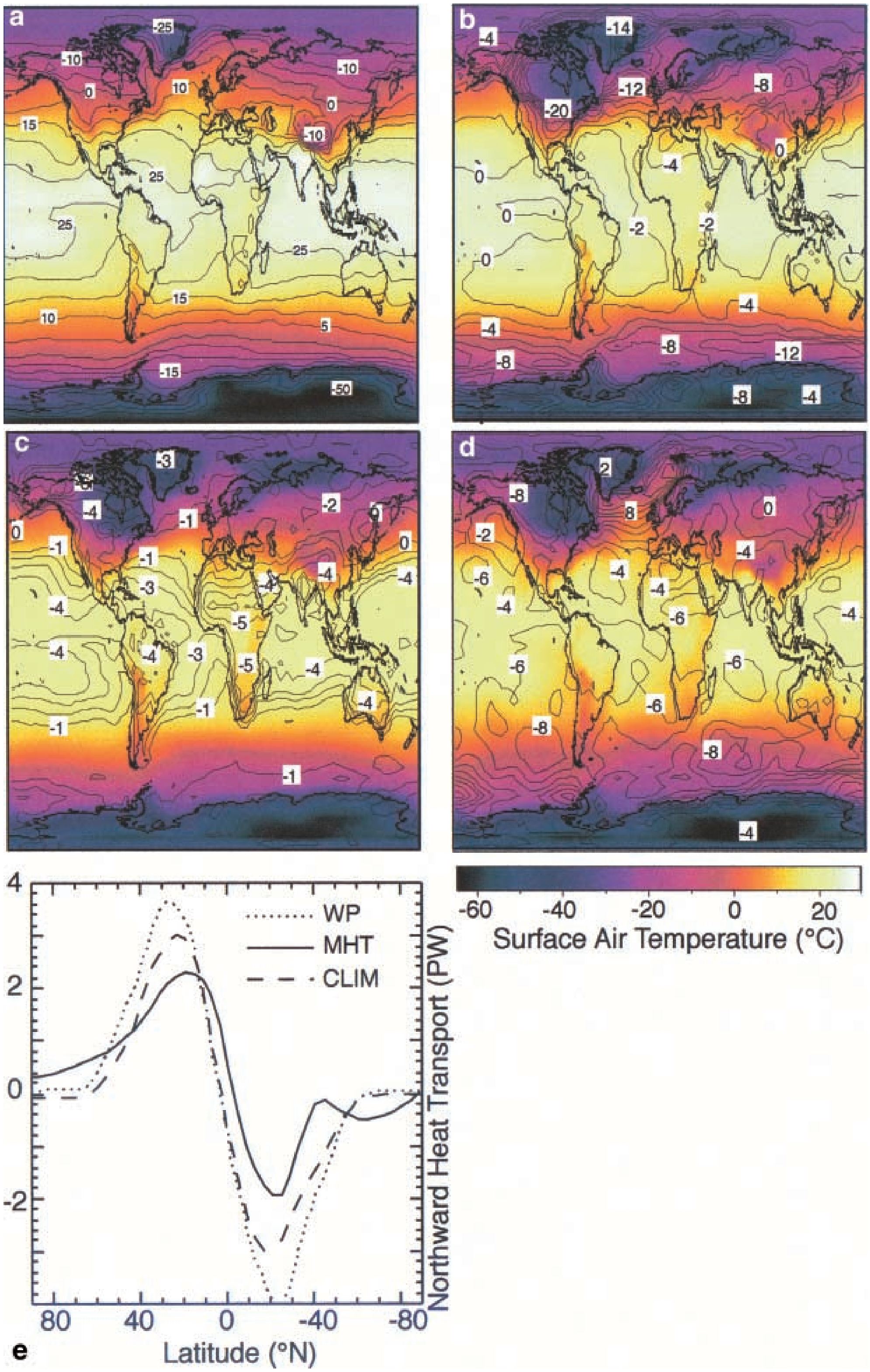


Fig. 1a–e Mean annual surface air temperature for **a** CON experiment; **b** CLIM experiment; **c** WP experiment; **d** MHT experiment; **e** implied ocean heat transports for the three glacial simulations; **f** same as **a**, but expressed as the difference of CLIM minus CON; **g** WP minus CLIM; **h** MHT minus CLIM

significantly warmer (8–10 °C) than CLIMAP ice age estimates, because there is a large convergence of oceanic heat in this region in the modern simulation. But there are other notable differences as well. For example, in MHT, the tongue of cool water in the equatorial eastern Pacific is more pronounced than in WP, and the Southern Ocean adjacent to Antarctica is slightly warmer. As we will demonstrate, some aspects of the hydrological cycle are particularly sensitive to the various SST differences. However, the inferences that can be drawn from these hydrological cycle changes are general enough that they need not depend on the specific SST pattern imposed.

3 Results

3.1 Radiation

With respect to the modern control (designated CON), global mean annual surface air temperature decreased by 5.5 °C in CLIM, by 7° in WP and by 8.5 °C in MHT. Hansen et al. (1984) analyzed the radiation balance for ice age simulations similar to CLIM and showed that a significant component of the cooling in these GISS standard ice age simulations involves water vapor feedback. Not surprisingly, in our experiments, lower tropical SSTs further reduced the greenhouse capacity of the atmosphere; the total water vapor in the atmosphere decreased by an additional (relative to CLIM) 24% in WP and by an additional 33% in MHT. Such enormous reductions were partially responsible for a greater temperature response over the continents. For example, mean annual temperatures over the Laurentide ice sheet in both experiments decreased by as much as 6 °C (relative to CLIM), while temperatures throughout South America and Africa decreased by up to 9 °C relative to CON (Fig. 1). MHT predicts a slightly more pronounced and widespread cooling of the continents than does WP, consistent with the difference in atmospheric water vapor (though other dynamical factors contribute to the cooling over Northern Hemisphere ice sheets in both MHT and WP). On the other hand, MHT was characterized by reduced sea ice in both the North Atlantic and the southernmost Southern Ocean, and, consequently, the adjacent polar continental areas in MHT were much warmer than in CLIM and WP. Thus, the continental temperature response in MHT is quite latitudinally heterogeneous.

MHT was nearly in radiative equilibrium with respect to the control run (by construction), while WP and CLIM were not: the difference in radiation balance, with respect to the modern control, was $+3$ W/m^2 for WP and -3 W/m^2 for CLIM. Thus, left to its own devices, the model would never produce the SST fields imposed in these two experiments. In principle, the imbalances imply that the SSTs would increase in WP and would decrease in CLIM if they were allowed to adjust to equilibrium. In practice, however, the significance of these imbalances is difficult to assess because of the many compensatory factors that ultimately determine equilibrium surface temperatures. For example, one might conclude from the radiative balance that we arbitrarily imposed an unrealistically severe cooling in WP. Yet the SSTs in MHT are in fact slightly cooler than in WP throughout the tropics and subtropics, on a zonal average basis. This counter-intuitive radiative response in MHT results from the fact that widespread increases in total cloud cover led to higher albedo and less of a low latitude radiative surplus (Table 1). Nearly the opposite effect occurred in WP, where tropical cloud albedo decreased significantly. Thus the nature of the radiative equilibrium reached in the MHT experiment may be somewhat fortuitous, because it hinges on the validity of the tropical cloud cover response, a subject of obvious uncertainty in climate modelling.

Table 1 Comparison of selected mean annual diagnostics

Variable	CON			CLIM			WP			MHT		
	Global	NH	Land	Global	NH	Land	Global	NH	Land	Global	NH	Land
Surface temperature (°C)	13.3	14.2		8.95	8.72	1.1	6.34	5.69	−3.19	4.69	5.43	−3.88
Albedo (%)	29.2	28.5	33.8	31.4	31.2	37.4	31.3	31.4	39.1	33.1	32.3	38.7
Net radiation surface (w/m^2)	126.8	125.6	103.1	119.7	118.3	91.3	117.1	115.6	87.7	113.4	113.4	87.7
Net radiation planet (w/m^2)	9.4			6.2			12.8			9.4		
Cloud cover (%)	49.3	48.0	48.1	48.7	47.0	48.0	45.7	44.4	50.7	50.0	48.1	50.3
Tropical lapse rate (deg/km)	5.2			6.2			6.6			6.5		
Precipitation (mm/day)	3.2	3.3	2.7	3.0	3.0	2.2	2.7	2.6	2.4	2.6	2.9	2.1
Atmosphere water vapor (mm)	24	25	19	21	21	15	16	16	11	14	15	11
Ocean ice cover (%)	4.9	4.3		8.7	4.0		8.7	4.0		9.7	2.9	
Snow cover (%)	11.9	13.5		20.7	23.1	37.0	21.3	24.0	38.2	22.5	23.2	39.2
Snow depth (cm)			22.9			29.0			32.8			31.2
Net ice 55–47°N (mm/day)						−0.4			−0.2			−0.2

3.2 Dynamics

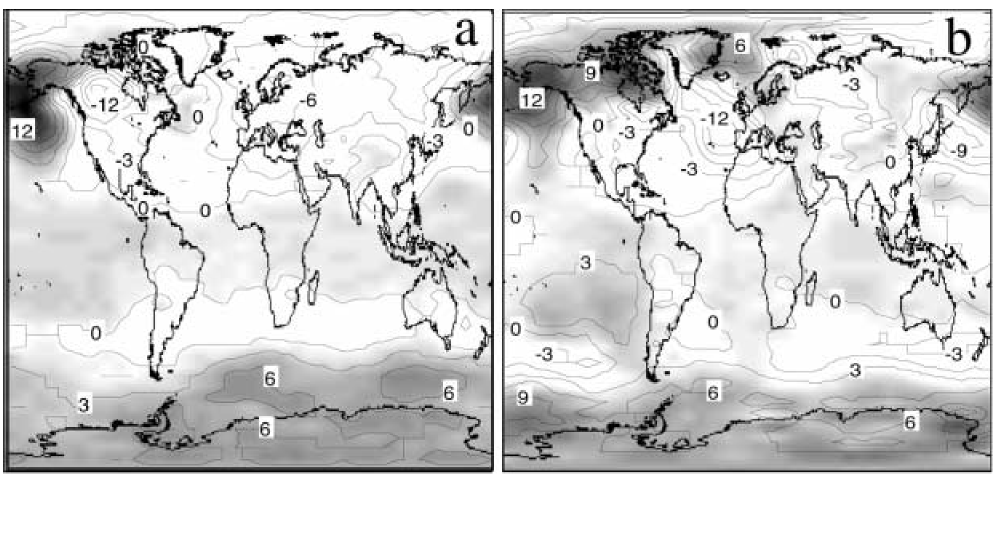
Rind (1986, 1987) presented a thorough analysis of the dynamics of the ice age atmosphere in the GISS model using CLIMAP boundary conditions (i.e., our CLIM experiment). However, the imposition of dramatically different SSTs in WP and MHT leads to several dynamical effects not present in CLIM. In high latitudes, colder N Pacific subtropical surface temperatures weakened the Aleutian low (Fig. 2). The effect was substantial in both MHT and WP. These surface changes partially account for the temperature response over North America. The pattern of surface pressure and temperature anomalies in many ways resembles the standing wave mode generated during El Nino/Southern Oscillation (ENSO) cycles, the Pacific North American mode (Wallace and Gutzler 1981), with both tropical cooling experiments showing mid-latitude anomaly patterns akin to the ENSO cold-phase. Decreased continental temperatures, increased snow cover, and shifting regions of convection over the ocean also led to altered axes of monsoonal flow in both tropical cooling experiments. This change is manifested clearly in the different moisture source tracers (described later), even though the monsoonal surface wind changes are not highly significant in a statistical sense (Fig. 3).

The latitudinal SST gradient imposed in WP was much weaker than in CLIM, and, since lapse rates did not change dramatically (aside from the effect of having less water vapor in the atmosphere), the mid-latitude westerly winds at 200 mb were also much weaker in this cooling experiment (not shown). Furthermore, the zonal and meridional SST gradients in MHT and WP were definitely not equivalent, despite the fact that SSTs were uniformly lower in the tropics in both these experiments. The different SST gradients had important consequences for the intensity and latitudinal extent of the Hadley Cell, as well as the operation of the Walker circulation. Plots of the stream function for boreal winter (Fig. 4) show that the Hadley cell weakened considerably in WP, because the tendency for low-latitude convection decreased (and therefore, so did the latent heat release aloft). Surface trade winds also weakened significantly (Fig. 3), while the vertical shear in the zonal wind along the equator increased. In MHT, the western Pacific warm pool cooling imposed was not quite as strong as in WP, and penetrating convection occurred in that region (this phenomenon is most apparent as an increase in high-cloud amount, but it is also apparent from the convergence of surface winds along the equator). Also, the vertically integrated stream function for MHT (Fig. 4) was characterized by higher absolute values relative to CLIM, with significantly intensified tropical easterlies. This pattern is consistent with the stronger latitudinal temperature gradient in MHT. Therefore, the mere condition of cooler tropical SSTs does not by itself produce exactly the same large-scale dynamical effects in all cases: as with the hydrological cycle, the dynamics of the model are most sensitive to SST gradients in the tropics (Rind 1986, 1998). These dynamical issues are treated in more detail in a separate report (Rind 1998).

3.3 Hydrological cycle

The hydrological cycle is weaker globally with cooler tropical SSTs. Over the ocean, the response in precipitation and evaporation is predictable from the SST change, in general, precipitation (and evaporation) decreases significantly in areas experiencing strong surface temperature reductions, and it increases slightly in regions where surface temperature changes the least. Although this general sense of the hydrological cycle variability is the same in both lowered tropical SST experiments, there were significant regional differences between WP and MHT because of their unique patterns of SST anomalies. In WP, the tropical convergence zones are characterized by intense reductions in both precipitation and evaporation, and, with the exception of the N Pacific, the subtropics are characterized by small precipitation increases. Also, in WP, the colder SSTs led to a more severe precipitation decrease in the

Fig. 2a, b DJF sea level pressure difference (mbar): **a** WP minus CLIM and **b** MHT minus CLIM



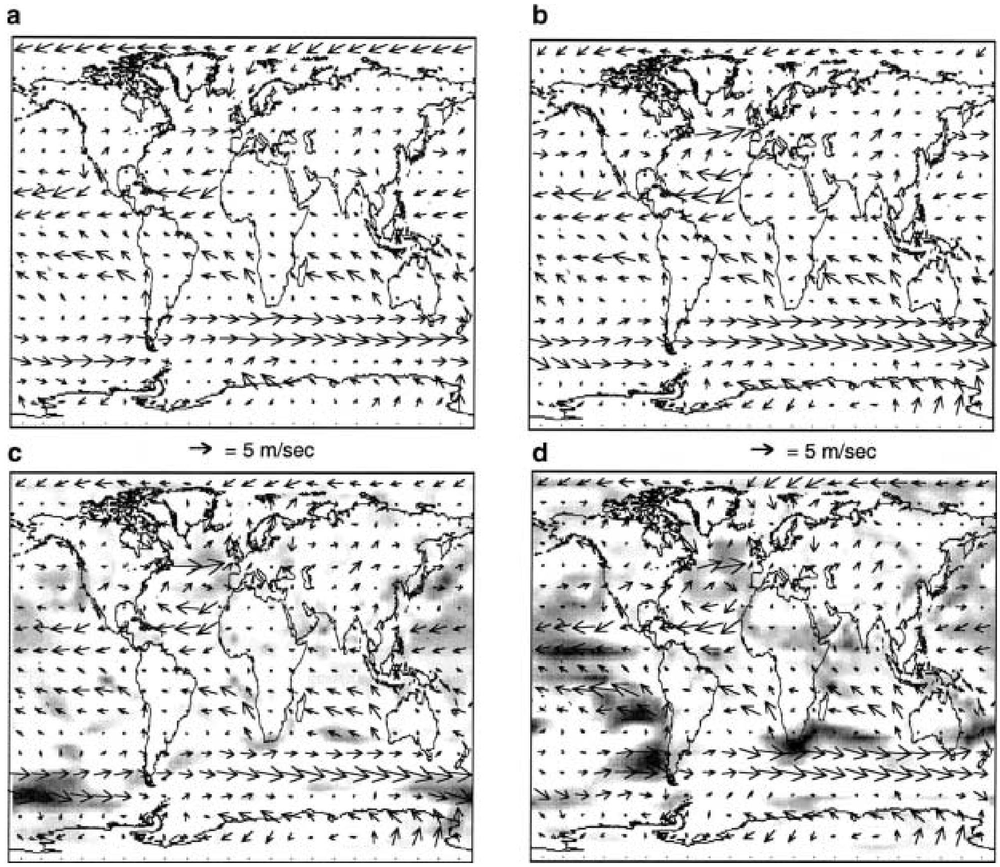


Fig. 3a–d Average July surface winds for **a** CON; **b** CLIM; **c** WP; **d** MHT. *Shaded regions* in **c**, **d** represent areas where wind speeds are higher than in CLIM

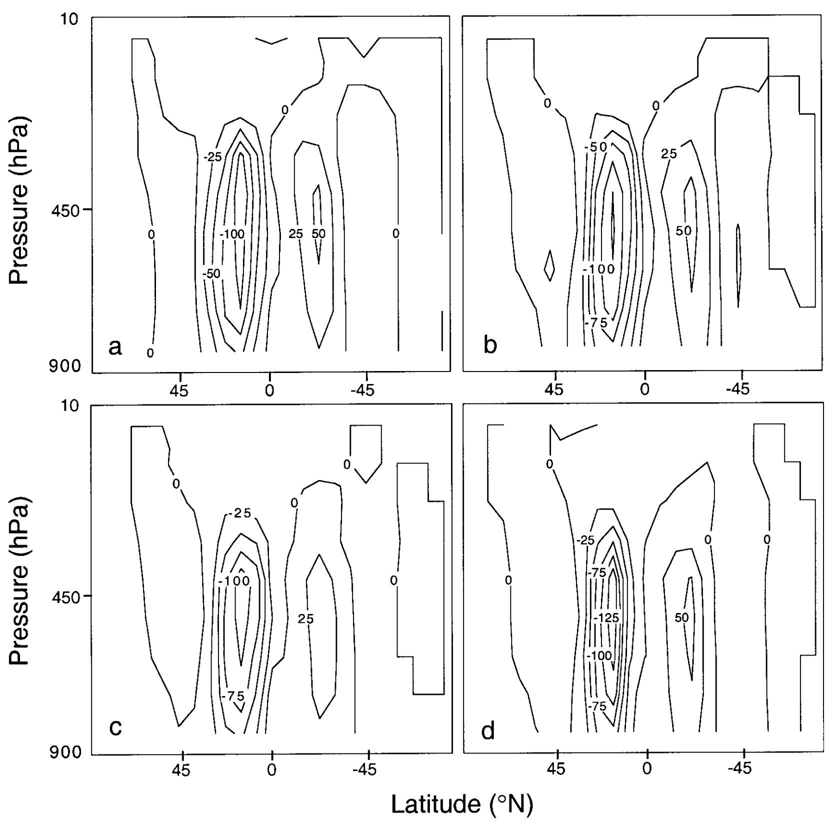


Fig. 4a–d DJF stream function (10^9 kg s^{-1}) for **a** CON; **b** CLIM; **c** WP; **d** MHT

western Pacific than in the eastern Pacific. In MHT, on the other hand, the precipitation reduction in the tropical convergence zones and in the subtropics is particularly intense over the eastern Pacific and the southern Indian Ocean while other tropical regions with more minor cooling, W equatorial Pacific, the northern Indian Ocean and the Caribbean, all show increased rainfall (Fig. 5).

Over the continents, the hydrological cycle response for the mid- to high-latitudes is broadly similar in both lowered SST experiments: despite lower rates of precipitation, snow cover and snow depth increase on all high-latitude continental areas (Table 1). This increase is a direct consequence of lower temperatures inhibiting melting, especially in summer. Although the GISS II model has difficulty maintaining an ice sheet in CLIM (see Rind et al. 1989 for further discussion of this point), cooling of the tropics leads to enough of a temperature decrease over high-latitude continents that the mass balance of ice becomes much less negative; the imbalance produced in CLIM is reduced by more than 50% in both WP and MHT, even at low elevation margins of the ice sheets and even with a much warmer North Atlantic in MHT.

Over low-latitude continents, there are more pronounced differences between the two cooling experiments. In WP, the subtropical regions remain significant centers for evaporation (because there was comparatively little reduction of modern subtropical SSTs.) The moisture evaporated in the subtropical regions is advected equatorward, and this moisture source, along with reduced trade wind speeds (Fig. 3) helps keep the continental areas wet. In MHT, both the subtropics and the tropics cooled, and therefore, there were no large organized centers for marine evaporation besides the central equatorial Pacific and the N Atlantic regions. As a result, the continental regions of the tropics became more arid in MHT, much drier than in WP, despite the fact that the "warm pools" in MHT were not quite as uniformly cool. Tropical East Africa in MHT is a clear example of this drying effect stemming from the intense cooling of the southern subtropical Indian Ocean. Moisture source tracers evaporated from subtropical grid boxes highlight these results more explicitly (Fig. 6).

Further insight to the terrestrial hydrological cycle comes from calculating the relative influence of continentally derived moisture. The distribution of the continental source tracer shows that, in both experiments, the percentage of recycled continental moisture fueling precipitation over land increases dramatically when tropical SSTs are lower (Fig. 7). The magnitude of this recycling effect is far greater in WP, because continental

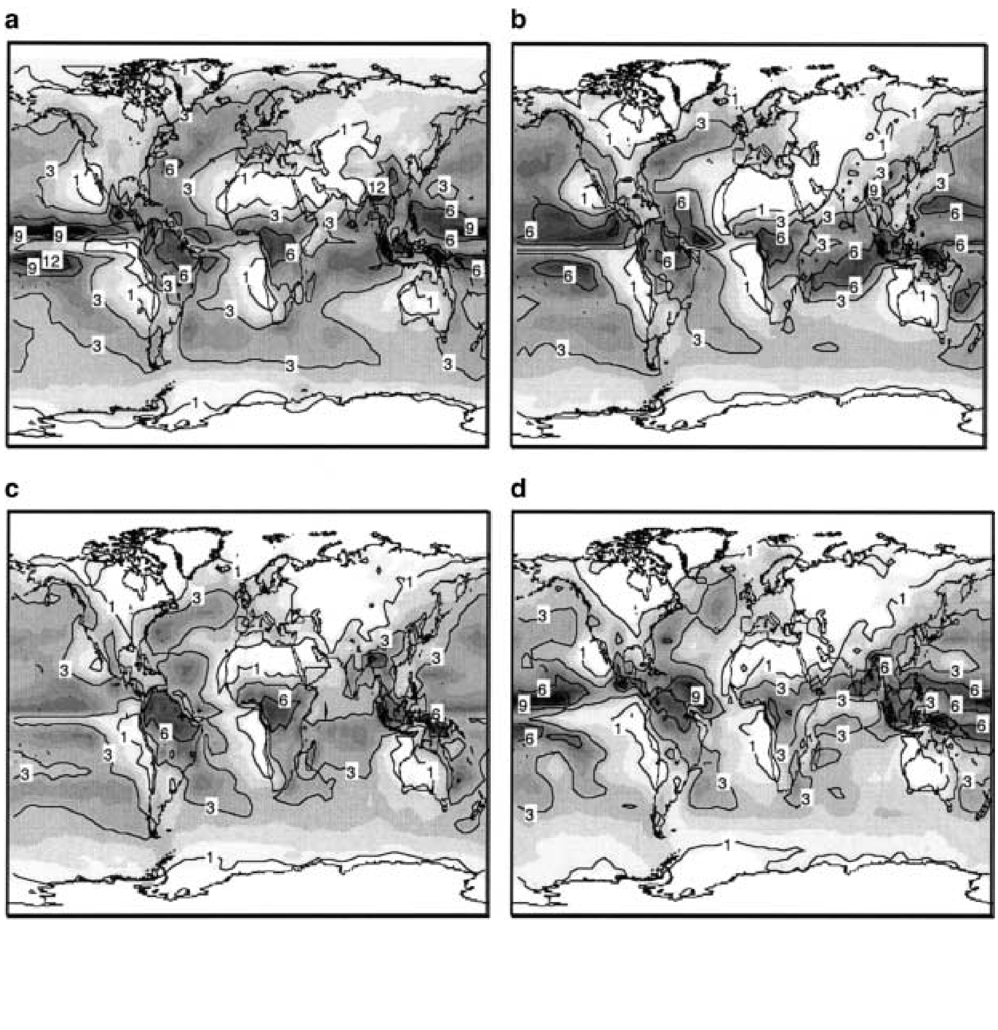


Fig. 5a–d Mean annual precipitation (mmday^{-1}) in **a** CON; **b** CLIM; **c** WP; **d** MHT

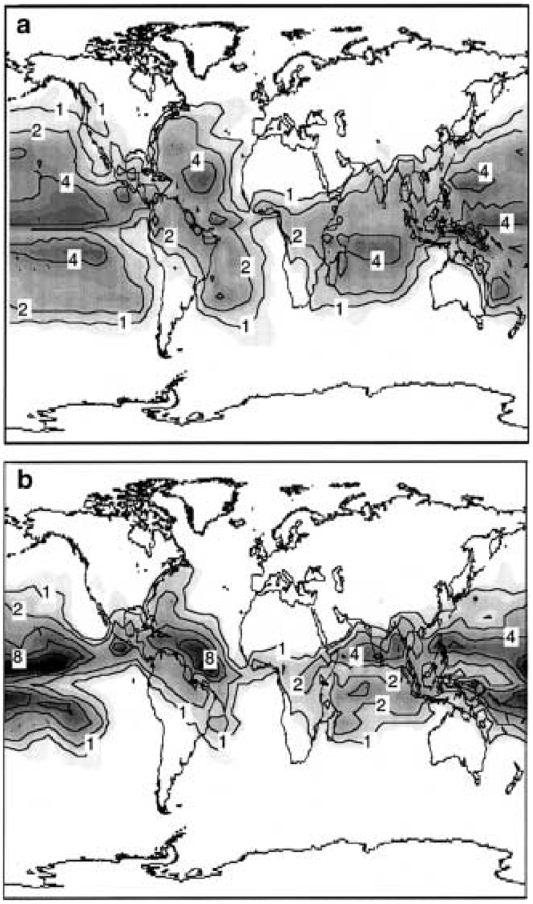


Fig. 6a, b Amount of local precipitation (mmday^{-1}) that originally evaporated from gridboxes with a temperature greater than 15 °C in **a** WP and **b** MHT

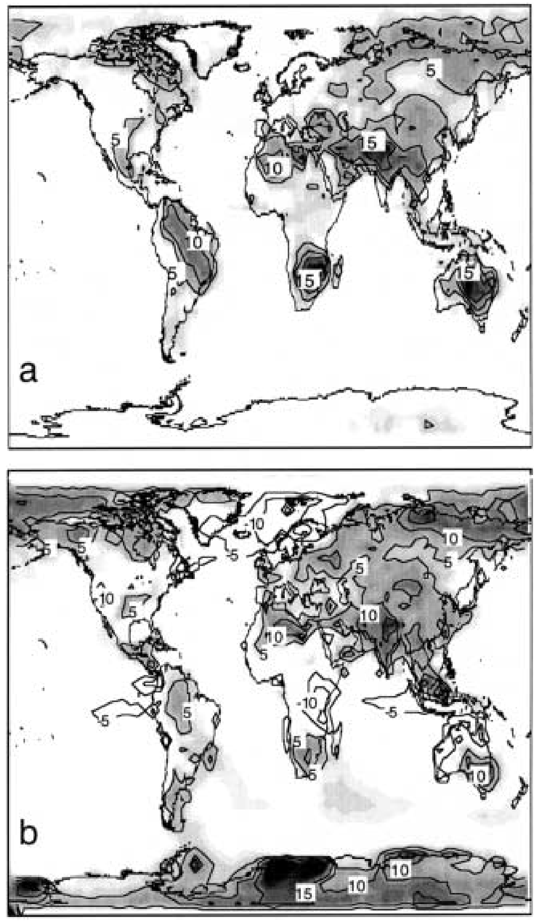


Fig. 7a, b Change in the percentage (%) of precipitation originally evaporated from continental gridboxes. **a** Difference between WP-CLIM and **b** difference between MHT-CLIM. Positive values indicate greater continental recycling of moisture

precipitation is not nearly as strongly reduced by the lower tropical SSTs (in WP, ground wetness actually increases slightly, relative to CON, throughout much of the tropics). By contrast, in MHT, the soil moisture content is reduced enough that the relative influence of continental evapotranspiration is limited, in spite of the decreased marine moisture supply.

3.4 Isotope tracers

Significant and unique changes in the isotopic composition of precipitation occur in the two tropical cooling experiments (Fig. 8). The differences in $\delta^{18}O$ result from a geographically complicated mixture of temperature changes, precipitation changes, and moisture source changes.

In some regions, the distinction between these various influences is clear. For example, over North America and over the Tibetan Plateau, the reduction in $\delta^{18}O$ in both cooling experiments (as much as 6‰, relative to CLIM) corresponds with the temperature decrease, with the approximate proportionality expected (0.7‰/°C) if temperature were the main control (Dansgaard 1964).

On the other hand, in the low-latitudes, $\delta^{18}O$ variability generally bears little relationship to surface temperature, and is instead more closely tied to rainfall variability. This is true both geographically and seasonally in the model, a result borne out by modern isotopic observations that also exhibit a weak local air temperature dependence (Dansgaard 1964; IAEA 1995). In fact, modern observations show that any correspondence between $\delta^{18}O$ and temperature in the tropics is usually in the opposite sense to that observed in the mid- to high latitudes, because lower surface temperatures correspond to less intense rainfall, which in turn implies increased $\delta^{18}O$. In our experiments, the $\delta^{18}O$ increase over the eastern Pacific and central Indian Ocean in MHT (relative to CLIM) represents a particularly striking manifestation of this "amount effect".

In still other areas, particularly over the polar ice caps, the modelled $\delta^{18}O$ of precipitation is affected by both local air temperature change and moisture source changes. Simple Rayleigh models for scaling $\delta^{18}O$ variability to temperature, assuming a constant vapor source, are easiest to interpret in regions where the change in source region temperature is similar to the

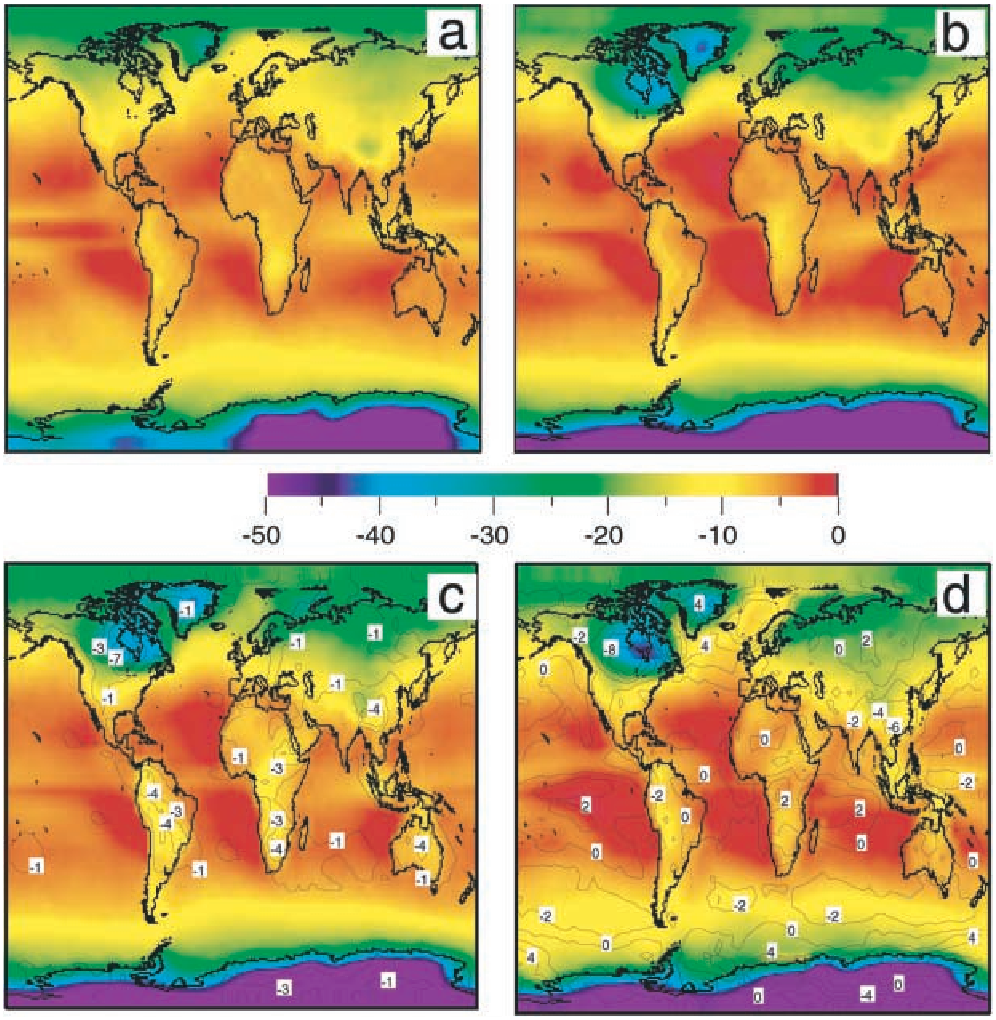


Fig. 8a–d Mean annual $\delta^{18}O$ (in ‰) of precipitation in **a** CON; **b** CLIM; **c** WP; **d** MHT. The contours in **c** and **d** denote the difference of $\delta^{18}O$ values with respect to CLIM. Note the similarities between the contours for $\delta^{18}O$ and the pattern of continental recycling (Fig. 7) over low-latitude continents

change in local temperature (Gat 1988). For example, in a Rayleigh model, as the difference between the source and local temperature becomes greater, more distillation occurs, and therefore, the isotopic composition is driven to lower values. Since we have defined source tracers on the basis of temperature, we can assess the isotopic influence of shifts in either the location or the characteristics of moisture sources by calculating the average temperature of the original sources for precipitation in any gridbox (Fig. 9).

Comparing the results for the Greenland ice sheet as a case study, we find different sensitivities of isotope change to local air temperature change, depending on the pattern of SSTs imposed. In the case of WP, the relationship of $\delta^{18}O$ to temperature change in a spatial sense (calculated by plotting the $\delta^{18}O$ and temperature along a hypothetical moisture source path for the North Atlantic) is characterized by a lower slope than for the modern case (Fig. 10). This somewhat muted isotopic response to temperature in WP results from the fact that the average temperature of the moisture source decreased more than did the local temperature at the site of precipitation (i.e., central Greenland Fig. 9). It should be noted, however, that this effect is fairly small and does not produce differences from modern trajectories that are highly significant from a statistical perspective. On the other hand, in MHT, the polar regions are especially affected by the increased influence of the more proximal vapor source, a shift that resulted from the reduction in sea ice in both hemispheres (relative to CLIM). In fact, the $\delta^{18}O$/temperature slope over Greenland and the North Atlantic in MHT is much steeper than for the modern spatial ratio. Thus the ultimate scaling of $\delta^{18}O$ to local air temperature over Greenland in the model clearly depends on ocean-wide temperature distributions, as opposed to being influenced by SSTs in one specific region.

In low- to mid-latitudes, the moisture source tracer distributions suggests that the single strongest influence on the $\delta^{18}O$ of continental precipitation in both experiments is the degree of continental moisture recycling; the isotopic change is more closely related to this factor than all other variables combined (see patterns of precipitation, or temperature change). The low $\delta^{18}O$ values over land in the tropics are clearly associated with higher percentages of continentally-derived moisture. This

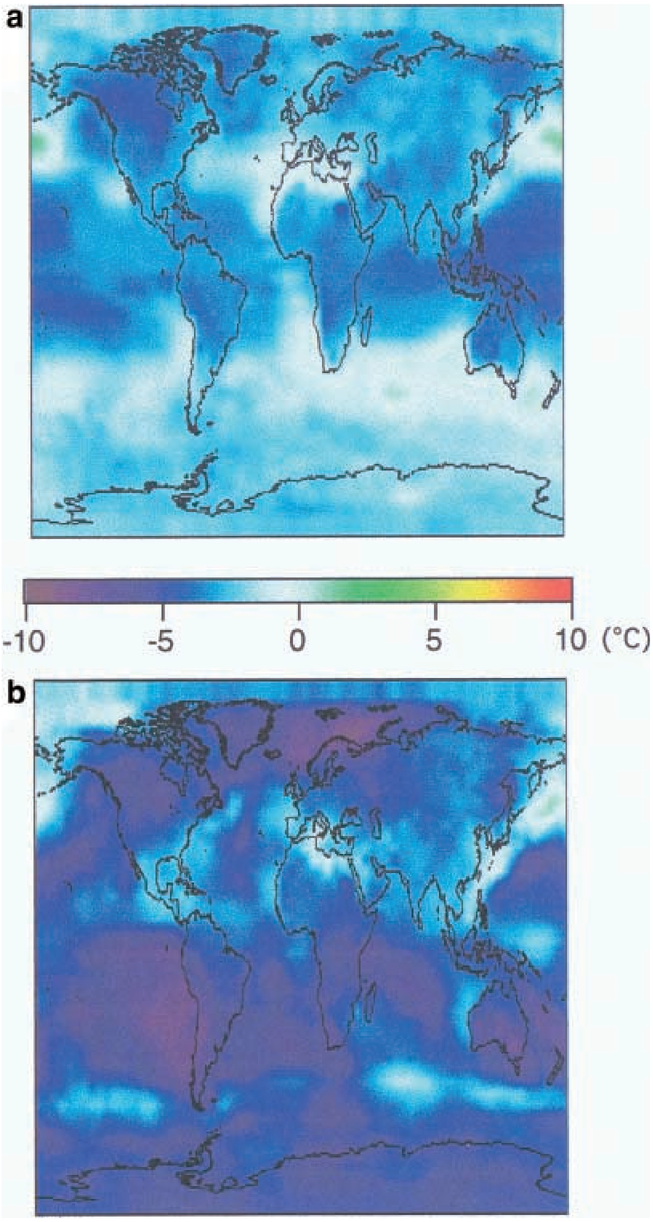


Fig. 9a, b The difference (a WP minus CLIM, and b MHT minus CLIM) in mean average temperature of the evaporative moisture source for local precipitation, calculated as a weighted average temperature. This weighted average is calculated using the specific fraction of local precipitation originating from a set of gridboxes, multiplied by the average temperature of those respective gridboxes

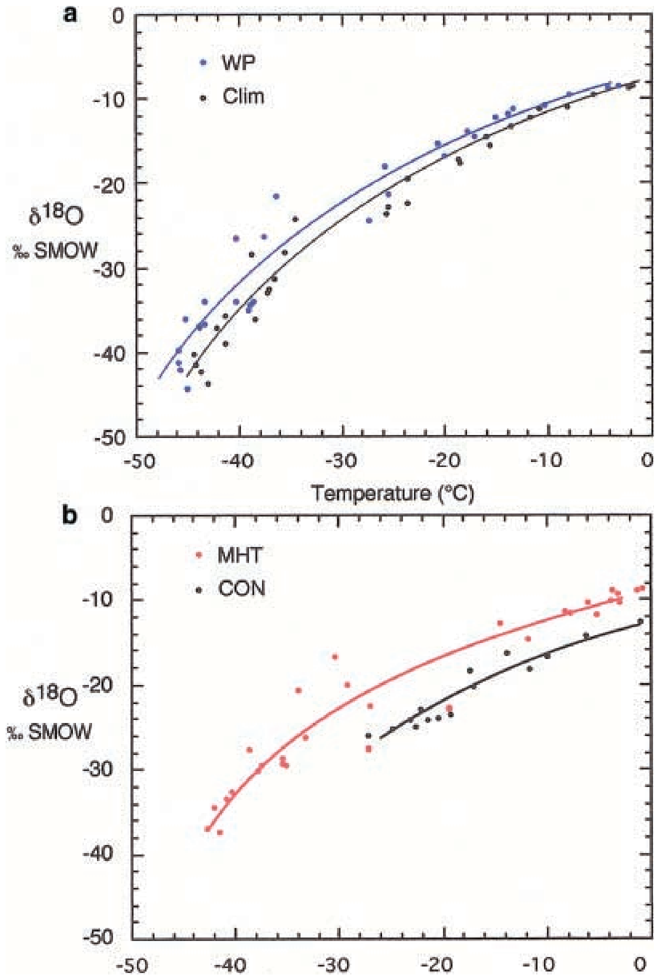


Fig. 10a, b Spatial trajectories of the mean annual δ^{18}O-temperature relationship for the different simulations, using values for Northern Hemisphere gridboxes lying along 50 to 40°W. These meridians were chosen to represent one dominant path for maritime moisture supplying Greenland precipitation. The experiments are split into two panels for clarity. The WP trajectory clearly has the highest δ^{18}O value for any given temperature below 0 °C

relationship implies that the isotopic composition of precipitation over regions such as tropical South America and equatorial Africa depends critically on the fraction of already distilled moisture evaporated from land, relative to direct (less isotopically depleted) marine sources. The relationship between continental recycling and isotopic composition is most clearly expressed in WP, the experiment with wetter tropical continents. In MHT, the continental aridity is extreme enough that the lower δ^{18}O values are not observed in most parts of the tropics, and regions such as equatorial Africa even experience a significant increase in δ^{18}O. It is important to emphasize that the large δ^{18}O differences between the two cooling experiments must be the result of changes in the hydrological cycle, independent of local temperature change, because the surface temperature response over the tropical continents is quite similar.

4 Discussion

The two cooling experiments show that the imposition of different patterns of lower tropical SSTs produces dramatic changes in the hydrological cycle that should be readily identifiable in the paleoclimatic record. In the continental realm, there are several types of geological evidence that can be compared to the predictions of these experiments for the glacial maximum: (1) the mass balance of the Northern Hemisphere ice sheets, which must have been positive to promote ice growth, or at least near neutral to prevent rapid, and unobserved, ice sheet retreat; (2) calibrations of δ^{18}O to temperature in Greenland ice cores (Cuffey et al. 1995; Severinghaus et al. 1998); (3) the lowering of snowlines in tropical glaciers by about 1000 m; (4) the ice core records from the Andes which show an ice age δ^{18}O decrease of about 8‰ (Thompson et al. 1995, 1998); (5) evidence for aridity in low latitude continents (e.g., Street and Grove 1979; Servant et al. 1993; Cross et al. 2000). We will consider each of these issues in turn.

The cooling of the tropical and subtropical Pacific in both the experiments here leads directly to much a cooler North America. In MHT, for example, the June temperature anomaly (relative to CLIM) reaches −9 °C over southern Canada and the Midwestern USA. WP produces a similar, though less extreme cooling pattern over North America (−4––6 °C relative to CLIM). This strong response suggests that, in whatever way it may be achieved, tropical and subtropical cooling could result in powerful dynamical (Cane 1998) and radiative (Webb et al. 1997; Bush and Philander 1998) feedbacks that serve to maintain or perhaps even generate an ice age. It is also significant that substantial cooling of the Laurentide region occurred in the model independently of the North Atlantic SSTs: both MHT and WP achieve similar temperature effects over North America, despite vastly different North Atlantic surface temperatures.

However, in most ice age experiments performed to date including those described here, the GISS model has produced a negative ice sheet mass balance over the Laurentide region (Rind et al. 1989). This inability to sustain a large ice sheet suggests that either the model is somehow flawed, or that the boundary conditions used are inappropriate (and therefore are not consistent with the radiation changes imposed.) Our experiments test the second of these possibilities. It is hard to imagine much colder boundary conditions than those assumed here, even if one envisioned increased aerosol loading of the ice age atmosphere (which we neglected). And, while the ice imbalance is reduced significantly in the tropical cooling experiments, the persistent negative balance makes one or more of the ice sheet parametrizations in the model (surface flux, albedo) seem suspect. In fact, our experiments with different snow and ice albedo formulations (for example, no aging of the ice versus moderate aging, etc.) did not eliminate the imbalance, so even this variable can be eliminated as the fundamental problem. A more accurate representation of surface fluxes over ice seems necessary for addressing the growth of large ice sheets in more detail with the GCM (Rind et al. 1989), and, in this regard, detailed intercomparison with other independent modelling studies of the Laurentide Ice Sheet may prove valuable (e.g., Hall et al. 1996).

The isotopic composition of polar ice has served as one principal means for deducing high-latitude temperature variability, though the exact scaling of isotopic variability to temperature has been subject to debate. Recent isotope/temperature calibrations suggest that applying the modern spatial gradient dδ^{18}O/dtemp (0.7‰/°C) to the ice core isotopic record may underestimate the actual glacial/interglacial temperature change over Greenland (Cuffey et al. 1995; Severinghaus 1998), and it is possible that hydrological cycle variability could contribute to a variable δ^{18}O/temperature scaling. In WP, there was a relatively small isotopic change over Greenland given the large amount of cooling. Thus, these WP results tend to support previous suggestions (Boyle 1997) that lowered tropical SSTs dampen the response of polar δ^{18}O to local temperature change by altering the initial conditions of the moisture source. However, this effect in our model experiments is quite small. Furthermore, the moisture source effect in the model is much more complicated than simply lowering a single evaporative temperature by 5 °C (as would be a typical assumption in a Rayleigh model), because sources other than the subtropical Atlantic contribute to Greenland precipitation. This complexity is highlighted in the case of MHT. In this experiment, the decrease in sea ice produced large positive isotopic changes (with respect to CLIM) without much accompanying temperature change over the ice sheets, because the evaporative moisture source shifted to higher latitudes. These results emphasize the need for accurate three dimensional isotopic models, in which the hydrological cycle is free to respond to climate changes in ways that are not always obvious from simpler models.

Rind and Peteet (1985) first demonstrated that CLIMAP boundary conditions do not produce a large enough cooling in the GISS model to allow for the observed lowering of tropical snowlines in Columbia, East Africa, New Guinea, and Hawaii. They suggested that a universal reduction of CLIMAP SSTs by 2 °C resulted in a better match to the observations, producing a 1000 m lowering of the 0 °C isotherm in the tropics. The results presented here reinforce this basic conclusion. In both cooling experiments, the 0 °C isotherm throughout the tropics descends to a much lower elevation, by about 1500 m, relative to the 4.5 km level of CLIM. Furthermore, unlike other aspects of the climate response, the pattern of SST cooling imposed makes very little difference in this freezing level descent, because zonal temperature gradients are smoothed aloft. Thus both tropical cooling experiments have no trouble reproducing significant tropical snowline lowering, even if precipitation were reduced strongly. In fact, in our experiments, precipitation reductions (such as over East Africa in MHT) suggest that 1500 m freezing level descents do not necessarily overestimate the observed 900–1000 m snowline descent.

While determining the cause of snow line descent may suffer from a problem of non-uniqueness, the 8‰ glacial age decrease in δ^{18}Oice in Peru and Bolivia (Thompson et al. 1995, 1998) may provide a more discriminating test for the magnitude and pattern of tropical SST lowering. The conventional interpretation of these ice core records is that the δ^{18}O change represents a direct response to a large ice age temperature decrease at the site of the ice cap, and that, therefore, this change demands lower tropical SSTs. Our results suggest a slightly different, but nevertheless perfectly complementary mechanism for arriving at the same implication of tropical ocean cooling. In our experiments, there is a strong indirect link between low $\delta^{18}O$ and lower tropical SST: the continental recycling effect, which produces low $\delta^{18}O$ values throughout much of South America and Africa, becomes more important as marine evaporation diminishes and as trade winds decrease. This continental recycling in turn depends on the prevalence and pattern of lower SSTs. Thus, from the GISS model perspective, the Andean ice core changes would be more precisely interpreted as a byproduct of isotopic "preconditioning" from terrestrial moisture sources that vary with the degree of tropical cooling, as opposed to being strictly a direct response to temperature.

An exact comparison between the Andean ice core record and the model isotopic results is not appropriate, because the model's topography cannot resolve the true elevation of the Andes. For example, the grid square incorporating Huascaran ice cap has an elevation of 1500 m, while the actual elevation is over 6000 m. And in reality, there must be some local temperature-dependent isotopic distillation occurring at high elevation over the Andes, because the $\delta^{18}O$ value of deep convective precipitation (at surface temperatures of 28 °C) reaches a minimum of about −10‰ (IAEA), while the recent values in the Huascaran ice core (at −10 °C) are about −17‰ (Thompson et al. 1995). Nevertheless, despite the model's limited resolution, all model results and modern observations suggest that it would be difficult to produce an 8‰ change in $\delta^{18}O$ice by air temperature change alone in the tropics. Our experiments show that a continental recycling effect could explain as much as half of this shift. Since both local temperature change and variable continental recyling are related in the cooling experiments to the same underlying cause, lower tropical SST, a combination of temperature and recycling effects seems likely in reality.

On a broader level, it is important to emphasize that CLIMAP's assumption of a relatively invariant tropical ocean also necessarily implies little or no change in tropical $\delta^{18}O$ from ice age to modern; the model simply cannot reproduce the 8‰ lowering observed over the Andes without both cooling and disturbance of the hydrological cycle. The same consideration applies to other low-to-mid latitude ice core sites such as Tibet, which also show a significant $\delta^{18}O$ice decreases during glacial periods (Thompson et al. 1997). Thus, our results would argue that the tropical ice core isotopic record is an irrefutable chronicle of substantial tropical climate change.

That said, there are still large $\delta^{18}O$ differences between even the two lower SST experiments over the tropical continents. One major discrepancy between the experiments lies in the ground wetness, and, therefore, the isotopic differences imply that the isotopic variability at low altitudes in these regions is highly sensitive to the availability of continental moisture. In this regard, one might ask how the different patterns of continental moisture produced in the experiments compare to the observations of general ice age aridity in the geological record. Compilations of lake level fluctuations (Street and Grove 1979), among other climate proxy variables (Servant et al. 1993), suggest that equatorial South America and Africa were much drier during the last ice age. There are some suggestions of increased ice age wetness in subtropical Northern and Southern Africa (Street and Grove 1979), but other indicators extend ice age aridity throughout the African continent (Sarnthein 1978). Over South America, there is considerable debate over the regional extent of drought, and recent evidence from the Bolivian Altiplano suggest wetter conditions during the ice age over parts of Amazonia (Cross et al. 2000). Of the three simulations presented here, only MHT predicts substantial drying of the tropical continents, and only the equatorial regions at that.

Of course, it is possible that the real low-latitude extent of ice age drying has been overestimated because of the sparse regional coverage and the difficulties in reading the paleoclimatic record of the tropics, including developing a reliable chronology for paleoenvironmental change (Ledru et al. 1998). On the other hand, if the agreement between several different lines of paleoclimatic evidence (e.g., Sarnthein 1978) is proof of widespread aridity, then several implications arise from our model results. First, cooling of the tropics alone is not sufficient to produce the observed aridity; WP, for example, imposed severe "warm pool" cooling, with little apparent effect on tropical ground moisture. Second, the pattern of cooling is a more important control on continental hydrological cycle. In particular, MHT is characterized by stronger zonal SST gradients and much cooler subtropical oceans. The lack of subtropical marine evaporation in MHT leads directly to drying of Africa and eastern South America. Also, any pattern of cooling that serves to strengthen the Walker Circulation creates the tendency for drought in some of the northern parts of South America, and this was the case with MHT. Regardless of these different possibilities, our results highlight the utility of $\delta^{18}O$ in lacustrine archives or aquifers as a quantative tracer of aridity. If widespread drying did occur during the last ice age, it should be readily apparent in the isotopic record. For example, one indirect implication of our results is that an 8‰ ice age decrease in $\delta^{18}O$ over Peru is inconsistent with extreme drought in the Amazon region of South America. By extension, wetter conditions over the Bolivian altiplano (Cross et al. 2000) might be completely compatible with lower $\delta^{18}O$ in the Andean ice cores (Thompson et al. 1995).

Finally, as a means for delineating the important mechanisms of ice age cycles, it is appropriate to consider whether any of the tropical cooling experiments leads to better reconciliation of paleoclimatic evidence on a global scale. By producing colder high-latitude continents, lower tropical snowlines, lower tropical $\delta^{18}O$

signature, and a somewhat muted isotopic reflection of temperature change over Greenland, WP clearly moves the terrestrial response in the proper direction for explaining the paleoclimatic record, all in ways that the CLIMAP SSTs fail. Many of these same apparent "improvements" (with the exception of the isotopic/temperature relationship) also appear in MHT, an experiment which arose from a definitive mechanism of maintaining constant oceanic heat transports and which is atleast in radiative equilibrium. What these experiments have in common is, of course, the substantial cooling of the central portions of the tropical Pacific, Indian and western part of the tropical Atlantic Oceans. Thus, these results reinforce other recent experiments (Hostetler and Mix 1999) that demonstrate the importance of conditions in a few key areas for dictating large-scale climate processes on all time-scales (Cane 1998).

However, the results also illustrate the importance of the ability to specify fields of temperature information throughout the ocean during glacial periods, because many aspects of the hydrological cycle response, low-latitude continental aridity, for example, depended more critically on the gradients of change, rather than on the average magnitude of tropical temperature change (see Rind 1998, for further discussion of the paleoclimatic significance of reconstructing SST gradients). This point should come as no real surprise for anyone familiar with interannual climate variability in the tropics, but it is often overlooked (by necessity) in the discussion of paleoceanographic reconstructions derived from a limited number of discrete core sites. While this requirement of broad geographical coverage poses a great challenge for the paleoceanographic record, it may be the only path toward successful integration of the land and ocean observations, or for understanding mechanisms of tropical oceanic change in general.

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References

Bard E, Rostek F, Sonzogni C (1997) Interhemispheric synchrony of the last deglaciation inferred from alkenon palaeothermometry. Nature 385: 707–710

Beck JW, Recy J, Taylor F, Edwards RL, et al (1997) Abrupt changes in early Holocene tropical sea surface temperature derived from coral records. Nature 385: 705–707

Boyle EA (1997) Cool tropical temperatures shift the global delta O-18-T relationship: an explanation for the ice core delta O-18-borehole thermometry conflict? Geophys Res Lett 24: 273–276

Broecker WS (1995) Ice cores – cooling the tropics. Nature 376: 212–213

Broecker WS, Denton GH (1989) The role of ocean atmosphere reorganizations in glacial cycles, Geochim Cosmochim Acta 53: 2465–2501

Bush ABG, Philander SGH (1998) The role of ocean-atmosphere interactions in tropical cooling during the last glacial maximum. Science 279: 1341–1344

Cane MA (1998) Climate change – a role for the tropical Pacific. Science 282: 59

CLIMAP Project Members (1976) The surface of the ice age Earth. Science 191: 1131–1138

Clapperton CM (1993) Nature of environmental changes in South America at the Last Glacial Maximum. Palaeogeogr Palaeoclim Palaeoeco 101: 189–208

Cole JE, Rind D, Webb RS, Jouzel J, Healy R (1999) Climatic controls on interannual variability of precipitation delta O-18: simulated influence of temperature, precipitation amount, and vapor source region. J Geophys Res 104: 14 223–14 235

Cross SL, Baker PA, Settzer GO, Fritz SC, Dunbar RB (2000) A new estimate of the Holocene low stand level of Lake Titicaca. Holocene 10: 21–32

Cuffey KM, Clow GD, Alley RB, Stuiver M, and others (1995) Large Arctic temperature change at the Wisconsin-Holocene Glacial transition. Science 270: 455–458

Dansgaard W (1964) Stable isotopes in precipitation. Tellus 16: 436–468

Ganopolski A, Rahmstorf S, Petoukhov V, Claussen M (1998) Simulation of modern and glacial climates with a coupled global model of intermediate complexity. Nature 391: 351–356

Guilderson TP, Fairbanks RG, Rubenstone JL (1994) Tropical temperature variations since 20 000 years ago: modulating interhemispheric climate change. Science 263: 663–665

Hall NMJ, Valdes PJ, Dong BW (1996) The maintenance of the last great ice sheets – a UGAMP GCM study. J Clim 9: 1004–1019

Hansen J, Russell G, Rind D, Stone P, Lacis A, Lebedeff S, Ruedy R, Travis L (1983) Efficient three-dimensional global models for climate studies, models I and II. Mon Weather Rev 111: 609–662

Hansen J, Lacis A, Rind D, Russell G, Stone P, Fung I, Ruedy R, Lerner J (1984) Climate sensitivity: analysis of feedback mechanisms. In: Hansen JE, Takahashi T (eds) Climate processes and climate sensitivity. AGU, Washington, D.C., pp 130–163

Hostetler SW, Mix AC (1999) Reassessment of ice-age cooling of the tropical ocean and atmosphere. Nature 399: 673–676

International Atomic Energy Agency (1990) Environmental isotope data numbers 1–8: world survey of isotope concentration in precipitation. IAEA, Vienna

Jouzel J, Koster RD, Suozzo RJ, Russell GL, White JWC, Broecker WS (1991) Simulations Of the HDO and (H2O)-O-18 atmospheric cycles using the Nasa Giss general circulation model-sensitivity experiments for present-day conditions. J Geophys Res 96: 7495–7507

Jouzel J, Koster RD, Suozzo RJ, Russell GL, White JWC, Broecker WS (1987) Simulations of the HDO and H218O atmospheric cycles using the NASA GISS general circulation model: the seasonal cycle for present-day conditions. J Geophys Res 92: 7495–7507

Koster RD, Jouzel J, Suozzo RJ, Russell GL, Broecker WS, Rind D, Eagleson P (1986) Global sources of local precipitation as determined by the NASA/GISS GCM. Geophys Res Lett 43: 121–124

Ledru M-P, Bertaux J, Sifeddine A, Suguio K (1998) Absence of last glacial maximum records in lowland tropical forests. Quat Res 49: 233–237

Rind D (1984) The influence of vegetation on the hydrologic cycle in a global climate model in: Hansen JE, Takahashi T (eds) Climate processes and climate sensitivity. AGU, Washington, D.C., pp 73–91

Rind D (1986) The dynamics of warm and cold climates. J Atmos Sci 43: 3–43

Rind D (1987) Components of the ice age circulation. J Geophys Res 92: 4241–4281

Rind D (1988) Dependence of warm and cold climate depiction on climate model resolution. J Clim 1: 965–997

Rind D (1998) Latitudinal temperature gradients and climate change. J Geophys Res 103: 5943–5971

Rind D, Peteet D (1985) Terrestrial conditions at the Last Glacial Maximum and CLIMAP sea-surface temperature estimates: are they consistent? Quat Res 24: 1–22

Rind D, Peteet D, Kukla G (1989) Can Milankovitch orbital variations initiate the growth of ice sheets in a general circulation model? J Geophys Res 94: 12 851–12 871

Rostek F, Ruhland G, Bassinot FC, Muller PJ, and others (1993) Reconstructing sea surface temperature and salinity using delta-o-18 and alkenone records. Nature 364: 319–321

Sarnthein M (1978) Sand deserts during the glacial maximum and climatic optimum. Nature 272: 43–46

Severinghaus JP, Sowers T, Brook EJ, Alley RB, Bender ML (1998) Timing of abrupt climate change at the end of the Younger Dryas interval from thermally fractionated gases in polar ice. Nature 391: 141–146

Servant M, Maley J, Turcq B, Absy ML, Brenac P, Fournier M, Ledru MP (1993) Tropical forest changes during the late Quaternary in African and South American lowlands. Global Planet Change 7: 25–40

Street FA, Grove AT (1979) Global maps of lake-level fluctuations since 30 000 yr BP. Quat Res 12: 83–118

Stute M, Forster M, Frischkorn H, Serejo A, and others (1995) Cooling of tropical Brazil (5 °C) during the Last Glacial Maximum. Science 269: 379–383

Thompson LG, Mosley Thompson E, Davis ME, Lin PN, and others (1995) Late glacial stage and holocene tropical ice core records from Huascaran, Peru. Science 269: 46–50

Thompson LG, Yao T, Davis ME, Henderson KA, and others (1997) Tropical climate instability: the last glacial cycle from a Qinghai-Tibetan ice core. Science 276: 1821–1825

Thompson LG, Davis ME, Mosley-Thompson E, Sowers TA, Henderson KA, Zagorodnov VS, Lin PN, Mikhalenko VN, Campen RK, Bolzan JF, Cole-Dai J, Francou B (1998) A 25 000-year tropical climate history from Bolivian ice cores. Science 282: 1858–1864

Wallace M, Gutzler DS (1981) Teleconnections in the geopotential height field during the Northern Hemisphere Winter Season. Monthly Weather Review 109: 784–812

Weaver A, Eby M, Fanning AF, Wiebe E (1998) Simulated influence of carbon dioxide, orbital forcing, and ice sheets on the climate of the last glacial maximum. Nature 394: 847–853

Webb RS, Rind DH, Lehman SJ, Healy RI, Sigman D (1997) Influence of ocean heat transport on the climate of the Last Glacial Maximum. Nature 385: 695–699