Sensitivity of the Atlantic Intertropical Convergence Zone to Last Glacial Maximum boundary conditions

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Received 22 April 2003; revised 21 July 2003; accepted 15 September 2003; published 11 December 2003.

[1] Recent paleoproxy records suggest that the mean latitude of the Atlantic Intertropical Convergence Zone (ITCZ) varied synchronously with North Atlantic climate over a range of timescales throughout the Holocene and Last Glacial Maximum. We show that the present-day "meridional mode" of atmosphere-ocean variability in the tropical Atlantic is a potentially useful model for understanding these paleoclimate changes. The tropical Atlantic in a coupled atmospheric general circulation and slab ocean model responds to Last Glacial Maximum conditions with a southward displacement of the ITCZ. This response arises primarily through the land ice sheet that forces increased North Atlantic trades analogous to the forcing on the present-day meridional mode. Changes to sea ice coverage and to ocean heat transport associated with a weakened Atlantic thermohaline circulation also cause a meridional mode response, though through different mechanisms. Our results highlight the potential for tropical Atlantic paleoclimate to be driven from the high latitude influences, in particular, land ice on glacial-interglacial timescales. *INDEX TERMS:* 3344 Meteorology and Atmospheric Dynamics: Paleoclimatology; 3339 Meteorology and Atmospheric Dynamics: Ocean/atmosphere interactions (0312, 4504); 4215 Oceanography: General: Climate and interannual variability (3309); *KEYWORDS:* Atlantic climate, Last Glacial Maximum, ITCZ

Citation: Chiang, J. C. H., M. Biasutti, and D. S. Battisti, Sensitivity of the Atlantic Intertropical Convergence Zone to Last Glacial Maximum boundary conditions, *Paleoceanography*, *18*(4), 1094, doi:10.1029/2003PA000916, 2003.

1. Introduction

[2] Recent sediment core records from the Cariaco Basin region (just off the northern coast of South America, approximately 10.5°N and 65°W) show that over the Holocene and last glacial cycle, the climate of the tropical Atlantic experienced significant changes that appear to be linked to the climate of the North Atlantic over a wide range of timescales, from decadal through glacial-interglacial [*Hughen et al.*, 1996; *Black et al.*, 1999; *Peterson et al.*, 2000; *Haug et al.*, 2001]. The results of *Peterson et al.* [2000] illustrate this point nicely. Figure 1 compares color reflectance (550 nm) of a Cariaco sediment core (Ocean Drilling Program (ODP) Hole 1002C) (a measure of surface productivity in the tropical ocean) to the ice core $\delta^{18}O$ record from the Greenland Ice Sheet Project (GISP) representative of temperature changes over the North Atlantic. The immediate impression is that the north and tropical Atlantic records vary on a range of timescales, and almost in lockstep. These conclusions are corroborated by other studies that show linkages during the Younger Dryas [*Hughen et al.*, 1996; *Haug et al.*, 2001] and also during the last eight centuries [*Black et al.*, 1999]. These records prompt the following two questions: (1) what pattern of climate variations are they representing, and (2) why is there such a strong linkage between the tropical and North Atlantic?

[3] All the studies cited above suggest that the Cariaco Basin is recording changes in the north-south position of the Atlantic Intertropical Convergence Zone (ITCZ) and the band of maximum rainfall embedded in it. A shift in the ITCZ can impact sediment records either through change in the runoff that feeds into the basin [e.g., *Peterson et al.*, 2000], or through change to the northeasterly trades, upwelling, and therefore nutrient supply and biological productivity [e.g., *Hughen et al.*, 1996]. This scenario resembles, and is likely modeled after, a leading mode of present-day interannual-to-decadal climate variability in the tropical Atlantic. This mode can be derived using maximum covariance analysis (MCA; also known as singular value decomposition (SVD) [*Bretherton et al.*, 1992]) of anomalies in tropical Atlantic sea surface temperature (SST) and surface wind or heat flux fields [e.g., *Chang et al.*, 1997; J. C. H. Chiang, and D. J. Vimont, Analogous meridional modes of tropical atmosphere-ocean variability in the tropical Pacific and tropical Atlantic, submitted to *Journal of Climate*, 2003]. Figure 2 show the leading mode of an MCA analysis of SST (left field) and both components of surface wind (right field) from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalyses data set 25°S–35°N across the tropical Atlantic basin (see Figure 2 caption for details). Figure 2a shows the spatial pattern for the leading

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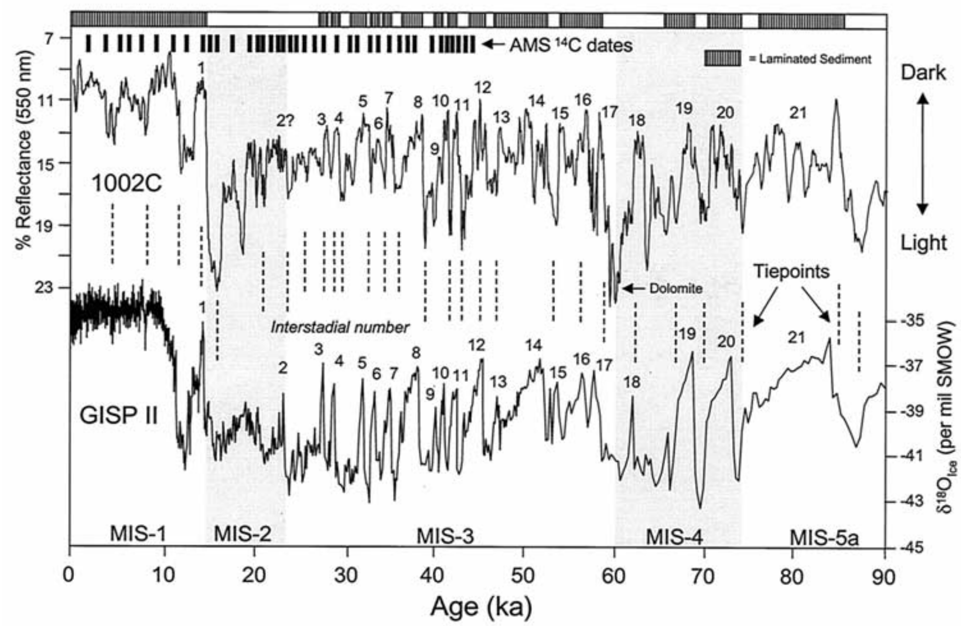

Figure 1. Demonstation of the coherence in paleoclimate variability between the tropical Atlantic and the North Atlantic. (Reprinted with permission from *Peterson et al.* [2000]. Copyright 2000 American Association for the Advancement of Sciences, http://www.sciencemag.org.) Measured color reflectance (550 nm) of Cariaco Basin sediments from ODP Hole 1002C is compared to δ^{18}O from the GISP II ice core. The reader is referred to Peterson et al. for details on dating and tie points. *Peterson et al.* [2000, p. 1948] makes this particular comment for this figure: "Deposition of dark, generally laminated sediments preferentially occurs during warm interglacial or interstadial times (numbered events), whereas deposition of light colored bioturbated sediments was restricted to colder stadial intervals of the last glacial. Sediment color variations in the Cariaco Basin are driven by changing surface productivity, with increased organic rain leading to darker sediments and, through remineralization reactions, periods of anoxic or near-anoxic conditions in the deep basin."

mode SST and winds, and Figure 2b shows the regression on the normalized SST expansion coefficients from the MCA of a merged satellite precipitation data set (CMAP [*Xie and Arkin*, 1997]). The SST pattern shows an anomalous meridional SST gradient across the position of the mean ITCZ and a displacement of the ITCZ and cross-equatorial flow toward the warmer hemisphere. This is the trademark spatial pattern of the dominant mode of tropical Atlantic covariability on interannual-decadal timescales [*Nobre and Shukla*, 1996], what we will refer to as the "meridional mode," following *Servain et al.* [1999]. The meridional SST gradient drives the cross-equatorial flow by creating a surface pressure gradient through hydrostatic adjustment of the atmospheric boundary layer [*Lindzen and Nigam*, 1987; *Hastenrath and Greischar*, 1993], which then drives a cross-equatorial boundary layer flow, changing the meridional position of maximum surface wind convergence and therefore the ITCZ. The ITCZ response is very sensitive to small SST gradients [*Chiang et al.*, 2002], an anomalous SST gradient of only ~1 K across the tropical Atlantic 15°N to 5°S, is able to displace the ITCZ hundreds of kilometers.

[4] To answer the second question why is there a strong linkage between the tropical the North Atlantic, we resort again to the present-day variability analogue. The meridional mode is not known to be self-sustaining and is thought to be driven from external forcing [*Kushnir et al.*, 2002]. The two known forcings are the North Atlantic Oscillation (NAO) and El Niño-Southern Oscillation (ENSO). The NAO describes interannual fluctuations in the northern Atlantic wintertime atmospheric circulation that impact SST throughout the entire North Atlantic basin, thus linking the variability of the north and tropical Atlantic climate [*Xie and Tanimoto*, 1998; *Marshall et al.*, 2001]. In fact, the North Atlantic anomaly sea level pressure (SLP) pattern associated with the MCA mode 1 (Figure 2c) resembles the NAO. Both ENSO and the NAO appear to drive the tropical Atlantic meridional mode through modulating the north tropical Atlantic (NTA) trade wind strength and therefore the latent heat fluxes and SST [e.g., *Seager et al.*, 2000; *Czaja et al.*, 2002]. The NTA SST anomaly thus formed projects on the meridional mode. Furthermore, there is evidence that the tropical Atlantic can feed back on the NAO through atmospheric teleconnections driven by the tropical convection changes [*Okumura et al.*, 2001], implying that the tropical and North Atlantic climate may, in fact, act as a cohesive unit [*Xie and Tanimoto*, 1998]. The present-day variability therefore allows us to make two points: that the tropical Atlantic is responsive to external forcing (and may in fact feed back on the forcing) and that

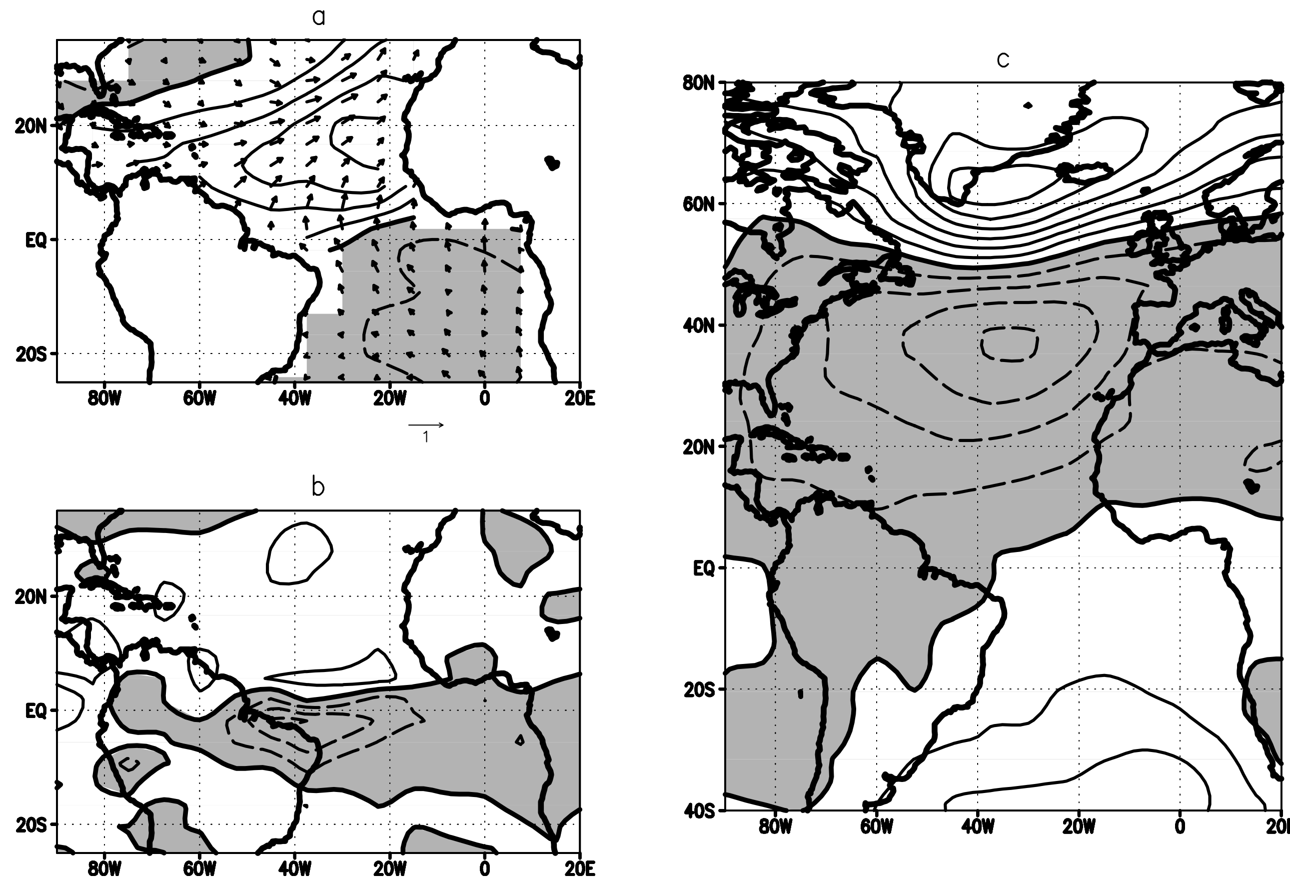

Figure 2. Meridional mode associated with meridional displacements of the Atlantic ITCZ, which here is extracted as the leading mode of a MCA analysis of SST (left field) and the zonal and meridional 1000 mbar wind field (right field) anomalies from the reanalyses. Prior to analyses, a light 3 month running mean was applied, and an index of the ENSO (the cold tongue index: SST anomalies averaged over 6°S–6°N and 180°W–90°W) was removed linearly from all fields. The analysis follows one similar to Chiang and Vimont (submitted manuscript, 2003). The removal of ENSO facilitates comparison with a similar analysis using the AGCM results (Figure 3), given that the AGCM does not simulate ENSO. (a) SST and wind fields shown as regression maps on the normalized SST expansion coefficients from the leading MCA mode. The contour interval is 0.1 kyr, and the reference wind vector is 1 m/s. (b) Regression on the CMAP precipitation data set 1979–2001, contour interval 0.1 mm/d. (c) Regression on surface pressure anomalies, contour interval 20Pa. In Figures 2a–2c, negative regions are shaded, and the thick line is zero contour.

there is a modern-day climate phenomenon that can link the climate of the tropical and North Atlantic.

[5] This study explores the tropical Atlantic response to paleoclimate-timescale forcing in a specific atmospheric general circulation model (AGCM). We hypothesize (1) that the sensitivity of the present-day Atlantic meridional mode to external forcing and small SST gradients makes it a preferred way for the tropical Atlantic to respond to past climate forcings and (2) that the model of atmospheric circulation linking the North Atlantic to the tropical Atlantic today may also be applicable in past climate and in understanding the differences between the modern-day climatology and the climatology under different boundary conditions (e.g., the LGM). Therefore the dynamical framework already developed for the meridional mode may be a useful model for past change.

[6] The model that we choose reflects these hypotheses; we use a full AGCM coupled to a simple thermodynamic slab ocean model (SOM). This dynamical model configuration is the simplest that we know of that incorporates a meridional-mode-like behavior as part of its internal variability (as shown by *Biasutti* [2000]; *Xie and Saito* [2001] also showed this in a slightly more elaborate atmospheric GCM-slab ocean model), in other words, the position of the model's Atlantic ITCZ varies from year to year in a similar fashion to observations. Furthermore, the model's leading mode of atmosphere-ocean variability in the tropical and North Atlantic resembles the meridional mode seen in observations.

[7] We will show that the tropical Atlantic responds to LGM conditions, specifically, land ice, sea ice coverage, and ocean heat transport (OHT) changes due to a weakened

Atlantic thermohaline circulation, preferentially through the meridional mode pattern. Furthermore, we will show that the Laurentide ice sheet has a large influence on the tropical Atlantic via trade wind changes, a mechanism analogous to that responsible for the tropical Atlantic response to the NAO. However, the meridional mode response can be elicited by other means, as in the case of the response to sea ice and Atlantic thermohaline OHT changes. We will also show that two other potential influences, a "permanent El Niño" condition and orbital changes due to precession, do not have as large an influence as might be suggested by our current understanding of meridional mode dynamics.

[8] Section 2 introduces the model and shows its applicability to our problem. In section 3 we describe the response to LGM forcings (greenhouse gas, orbital, and land ice sheet boundary conditions) and break it down into its components. In section 4 we explore how the land ice sheet influence reaches the tropics and displaces the ITCZ. In sections 5 and 6 we test the sensitivity of the LGM Atlantic ITCZ to changes to OHT and sea ice coverage by imposing them as external forcings. Section 7 reports on an experiment that uses 6 ka insolation to test the sensitivity of the ITCZ to orbital precession. Conclusions, and a discussion, are given in section 8.

2. Model Description and Simulation of Tropical Atlantic Variability

[9] We use the Community Climate Model version 3 (CCM3) general circulation model [*Kiehl et al.*, 1998] used extensively in simulations of tropical Atlantic climate variability [e.g., *Chang et al.*, 2000; *Saravanan and Chang*, 2000; *Giannini et al.*, 2001]. It is coupled to a fixed 50 m depth ocean mixed layer that allows thermodynamic atmosphere-ocean interactions, but no ocean dynamics. While inclusion of ocean dynamics is clearly preferable, our use of a thermodynamic-only slab ocean is a useful first step as suggested by two considerations. First, an ocean mixed layer model forced by observed winds has been shown to reproduce the dominant decadal variability of Atlantic SST [*Seager et al.*, 2000], with ocean dynamics playing largely a damping role in the tropics [*Seager et al.*, 2001]. Second, the zonal ENSO-like mode (also known as the Atlantic Niño) in the tropical Atlantic is substantially weaker than its equivalent in the tropical Pacific [*Zebiak*, 1993]. Consistent with these findings, the CCM3 coupled to a slab ocean has been shown to simulate the spatial structure of the variance in tropical Atlantic March-May SST, the season when the ITCZ variability peaks, realistically [*Saravanan and Chang*, 1999]. However, we note that several other studies have postulated a significant role for ocean dynamics in the coupled behavior in the tropical Atlantic: *Xie* [1999] showed the potential role of Ekman advection in changing the spatial pattern and frequency of the meridional mode, and *Chang et al.* [1997] suggested that advection of temperature by the mean surface currents plays a primary role in switching its phase. Furthermore, despite its relative weakness, the Atlantic Niño explains a substantial fraction of the interannual variability in the tropical Atlantic and is likely to respond to paleoclimate forcing in its own right. We will discuss the potential role of the Atlantic Niño in section 8.

[10] We use a fairly coarse model resolution (a triangular truncation keeping 31 basis functions in the meridional dimension and 15 for the zonal, equivalent to 48 longitude and latitude grid points) that has been used successfully to simulate the extratropical climate [*Yin and Battisti*, 2001] and tropical Atlantic variability [*Biasutti*, 2000]. The coarser resolution implies an order of magnitude increase in model throughput relative to the CCM3 standard resolution (T42), which allows us to perform longer simulations and to test our hypotheses with various experimental configurations, without unduly sacrificing realism. Our control run has imposed modern day orbital, greenhouse gases, sea ice, and land surface conditions. A flux correction Q (hereafter referred to as the Q flux) is applied to the surface ocean fluxes to force the slab ocean to follow the present-day SST monthly mean climatology:

$$\rho_0 C_0 h_0 \frac{\partial T}{\partial t} = F + Q \tag{1}$$

where ρ_0 is the density of seawater (taken to be 1.026×10^3 kg/m^3), C_0 is the heat capacity (3.93×10^3 J/K kg), h_0 is the mixed layer depth, and F is the net surface flux into the ocean. Q is derived from an uncoupled simulation using fixed climatological monthly mean SST [*Kiehl et al.*, 1996]:

$$Q^m = \rho_0 C_0 h_0^m \frac{(T^{m+1} - T^{m-1})}{(d^{m+1} - d^{m-1})} - F^m \tag{2}$$

where T^m is the climatological monthly mean SST and F^m is the monthly mean net surface flux derived from that same run. In our case, the F^m is averaged over 30 years. The monthly mean Q^m values are linearly interpolated to obtain the instantaneous Q required in equation (1). We interpret the application of the Q flux as the imposition of the modern-day OHT, though in reality it also corrects for model bias.

[11] Sea ice coverage is also prescribed and varies in time to the present-day monthly climatology. Because sea ice is not interactive, we impose the additional condition that the open ocean temperature is always above the model sea ice freezing temperature threshold of $-1.7999°$C. This constraint, while unphysical, in practice has negligible impact on our simulations of the mean climate. Using imposed sea ice coverage provides two benefits: the imposed Q flux, which is computed from a fixed climatological SST run with the same sea ice coverage, is consistent with the given model boundary conditions, and the climate of the control slab ocean run is very close to that of the fixed-SST simulation.

[12] The model exhibits a realistic seasonal cycle of tropical Atlantic SST and winds (not shown) and a qualitatively correct meridional mode variability. Figure 3a shows the results of a MCA analysis of tropical Atlantic SST and surface winds similar to the one in Figure 2a but obtained using monthly mean anomaly fields from the last 50 years of a 65-year control model run. The characteristic pattern of

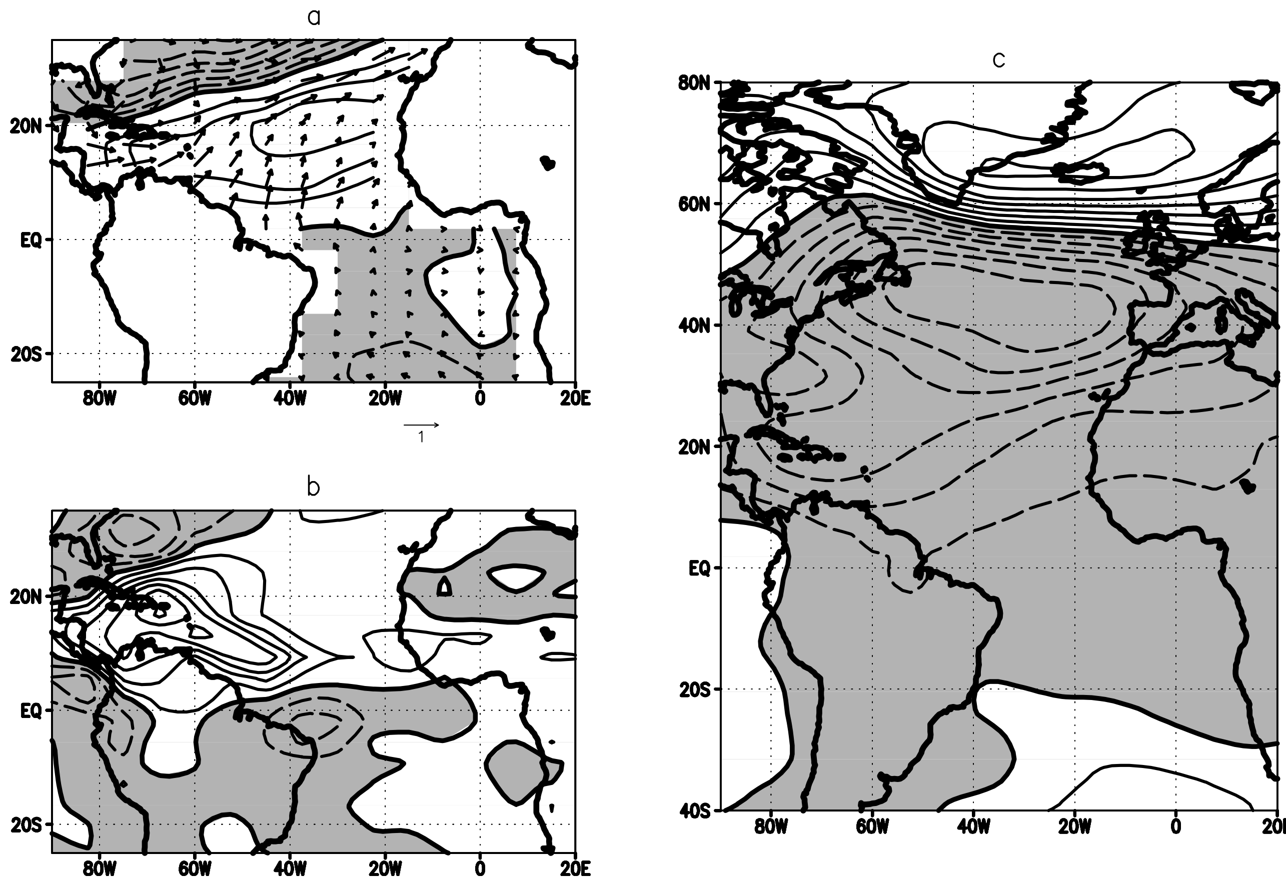

Figure 3. Same as Figure 2, but using monthly mean fields from a 50 year control run of the CCM3 AGCM with the 50 m slab ocean. The model MCA mode 1 resembles the observed MCA 1 mode, demonstrating that the model is capable of simulating meridional mode variability as the dominant tropical Atlantic atmosphere-ocean variability in the model.

meridional SST gradient and cross-equatorial flow is apparent in the model MCA leading mode. Figure 3b shows the model precipitation response. While the ITCZ "displacement" is apparent, the structure is unrealistic, in particular in the concentrated northern lobe of precipitation over northern South American and the Caribbean. This is a known deficit in CCM3 [e.g., *Saravanan and Chang*, 2000] which has to do with the unrealistically strong mean precipitation over that region. Hence, in the paper we interpret precipitation patterns like this as the model expression of meridional mode variability, rather than changes to Caribbean and Northern South American precipitation. Regression on the model SLP (Figure 3c) reveals a NAO-like pattern, indicating a midlatitude atmospheric forcing on the meridional mode similar to what is seen in the real world. A more detailed lead lag analysis (not shown) supports this interpretation.

[13] There are clear deficits in the model simulation of tropical Atlantic and North Atlantic variability, of which we mention the two most glaring. First, the amplitude of the model SST tripole pattern is too large compared to the observed, particularly at higher latitudes. Second, the persistence in the model Atlantic SST anomalies is longer than observed, for instance, the decorrelation time (defined here as the time lag taken for the autocorrelation to reach e^{-1}) of the SST expansion coefficients derived from the MCA analysis above is around 7.2 months for the model, compared to 5.7 months for the observed. The excessive variability is caused, at least partly, by the underestimation of the mixed layer depth at higher latitude, where the mixed layer depth is more likely to be around 150–200 m during winter. The increased persistence in SST is likely caused by the lack of ocean dynamics, which has been shown to damp Atlantic SST anomalies associated with the tripole SST pattern [e.g., *Seager et al.*, 2001]. Also, it has been shown [*Saravanan and Chang*, 1999] that using a realistic spatially varying (but time invariant) mixed layer depth rather than a uniform slab (as we apply here) significantly improves the simulation of variance in tropical Atlantic monthly mean SST anomalies. However, these shortcomings are primarily related to the transient response, given that the mixed layer ocean has a characteristic timescale comparable or longer than the persistence timescale of the atmospheric forcing. However, since we are simulating the equilibrated changes to annual mean climate forced by steady external forcings, these drawbacks should not unduly affect our model results. We tested this assumption by repeating one of our perturbation experiments, the 21 ka land ice sheet simulation (see

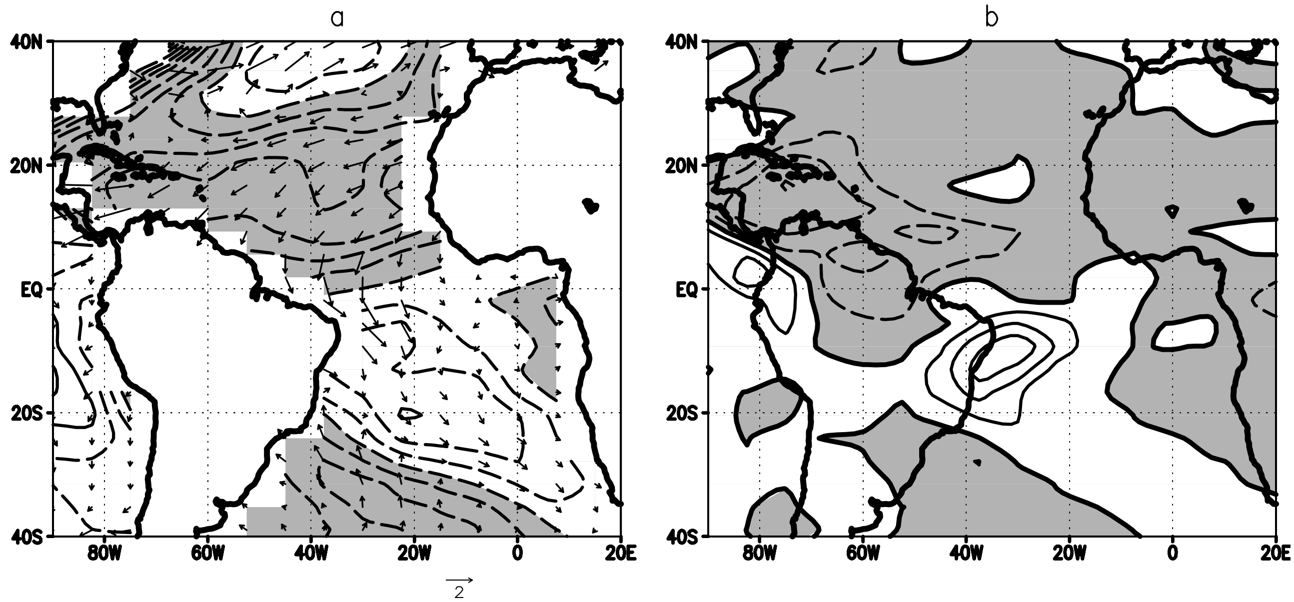

Figure 4. Changes to the CCM3 slab ocean climate given LGM boundary conditions. This shows the annual mean difference between the 21K ALL (CO_2, orbital, and land ice forcing) and the control run. (a) SST (CI 0.3K) and surface winds (reference vector 2 m/s); (b) precipitation (CI 1 mm/d). For SST, regions below -0.9 K are shaded. For precipitation, negative values are shaded, and the thick line is the zero contour.

section 4), using realistic spatially varying mixed layer depth. Indeed, the response is essentially the same as that for the uniform 50 m slab case. Overall, we believe our model does a reasonable job of simulating both the seasonal cycle and the meridional mode in the tropical Atlantic and can be used to examine the tropical Atlantic response in this system to LGM forcing.

3. Model Tropical Atlantic Response to LGM Forcing

[14] We apply the following changes to the modern-day CCM3 configuration to simulate the 21 ka LGM condition (hereafter referred to as the 21K ALL simulation): (1) CO_2 levels to 200 ppm from its standard "present-day" level of 355 ppm, (20) 21 ka land ice coverage, using the ice sheet reconstruction by *Peltier* [1994], and (3) 21 ka orbital configuration [*Berger*, 1978]. These forcings occur on timescales much longer than the adjustment timescale of the atmosphere and surface ocean and so can be imposed as external forcing in our experiments. The model configuration is essentially the same as used by *Yin and Battisti* [2001] (their LGM-SLAB run), except that in our case sea ice coverage is prescribed to present-day values. Our model lacks the ability to simulate the response of OHT and sea ice to LGM forcing. The sensitivity of the tropical Atlantic climate to changes in OHT and sea ice will be addressed in section 5. We do not address changes to LGM vegetation, or their possible feedback on the climate.

[15] All model results reported here are means over the last 20 years of 35-year simulations, allowing 15 years for spin up (which we found to be more than sufficient time for the model to equilibrate given LGM forcing). Significances in the differences of the means presented here are assessed using a two-sided t test, but for clarity of presentation we do not show them explicitly. Rather, in the text we only describe features that are significant at the 95% level.

[16] Figure 4 shows the difference between the annual mean 21K ALL run and the control run. Two features immediately stand out: the generally cooler tropical SST, and the spatial signature of the meridional mode, i.e., stronger trades in the NTA region, and a southward displaced ITCZ associated with a change in the meridional SST gradient across the equator. The spatial patterns resemble the model meridional mode pattern in Figure 3.

[17] What is the relative contribution of each 21 ka forcing? We performed additional simulations for each individual 21 ka forcing: 21 ka orbital configuration (hereafter 21K INSOLATION), 21 ka CO_2 concentration (hereafter 21K CO2), and 21 ka land ice sheet (hereafter 21K LAND ICE). Adding the anomalies due to individual forcing results in fields that resemble the anomaly field of the 21K ALL run in both pattern and amplitude (not shown), suggesting that to first order the response to the forcing is linear. We interpret the runs with individual forcing under this assumption. The orbital effect on the tropical Atlantic (not shown) is negligible, not surprisingly, as the LGM orbital configuration is similar to present-day. Figure 5 shows the tropical Atlantic SST and lowest model level wind for the 21K CO2 and 21K LAND ICE. The 21K CO2 (Figure 5a) contributes to a roughly uniform cooling of around 1 K throughout the entire basin. Interestingly, the precipitation response (not shown) is very small, despite the cooler SSTs, and the surface wind response is negligible. The lion's share of the response in wind, precipitation and SST gradient comes from the land ice sheet forcing

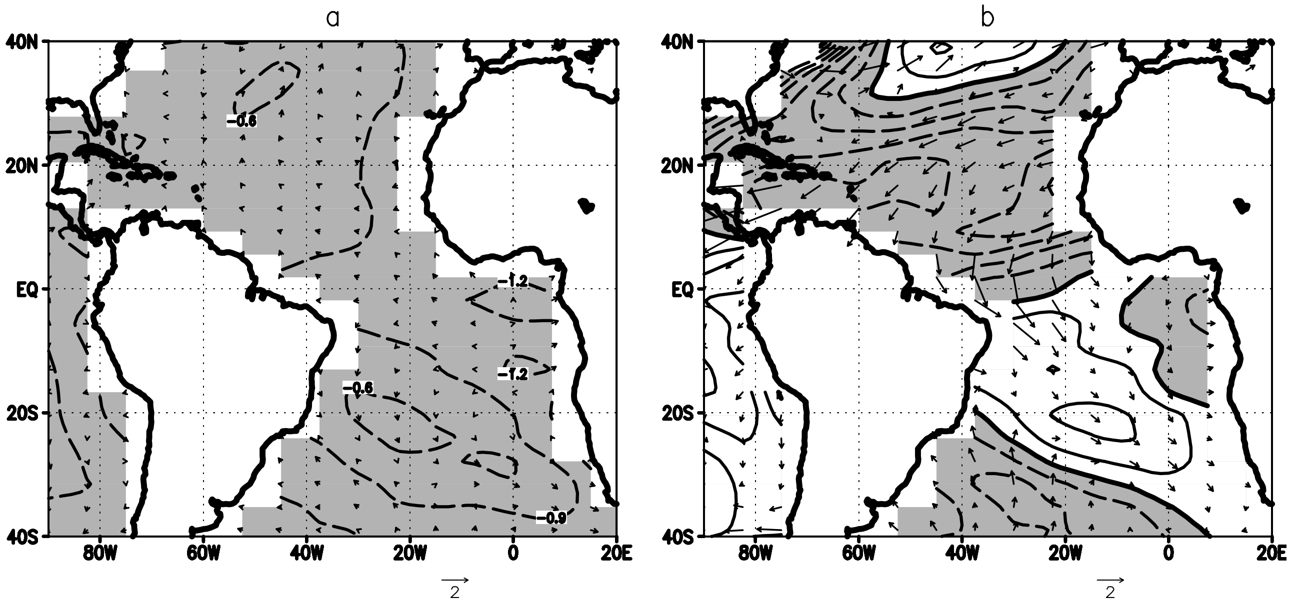

Figure 5. Contribution of each LGM boundary condition to the overall LGM response in Figure 4. This shows annual mean difference SST and surface winds of the 21 ka individual forcing runs and the control. (a) 21K CO2; (b) 21K LAND ICE. The contour intervals and wind vector scaling are the same as in Figure 4; the only difference is that for SST, regions below 0 K are shaded.

(Figure 5b) which acts to cool the north tropical Atlantic while warming the south tropical Atlantic SST. A seasonal breakdown of the SST and wind anomalies shows that this is true for all seasons.

[18] These results suggest that the simulated LGM climate response can be usefully broken down in terms of the uniform basinwide and meridional mode responses. Accordingly, we define these summary statistics: (1) the tropical mean SST, taken to be the SST averaged across the basin from 20°N to 20°S, (2) the trade wind strength in the NTA region, taken to be the zonal wind speed averaged over 5°N–25°N over the entire tropical North Atlantic, and (3) the meridional SST gradient, taken as the average of the SST 5°N–20°N across the Atlantic basin, subtracted by the average of the SST 10°S–5°N, also across the Atlantic basin. A positive value of this measure implies that the north is warmer that the south.

[19] Figure 6 shows the difference in summary statistics between the perturbation runs and the control, averaged seasonally over December–February (DJF), March–May (MAM), June–August (JJA), and September–November (SON). The statistics corroborate the roles of the individual forcings summarized above. The basinwide SST cooling for the 21K ALL run is around 1 K, with relatively little seasonality in the response. The 21K CO2 is responsible for the bulk of this response, but 21K LAND ICE also makes a sizable contribution. The 21K INSOLATION warms the basin slightly. The northeasterly trades are stronger during all seasons in the 21K ALL run, but with strong seasonality: the strongest anomalies are in DJF followed by SON; the changes during MAM and JJA are about half that in DJF. The 21K LAND ICE accounts for almost all of the NTA trade wind increase. Likewise, 21K LAND ICE is also responsible for the lion's share of the negative (cool north-warm south) meridional SST gradient and equatorial meridional wind response seen in the 21K ALL run. The SST gradient anomalies are of the order of ~1 kyr; they are strongest in the DJF and SON months, but the seasonality of this response is small, presumably because of the thermal inertia of the slab ocean.

4. Origins of the Meridional Mode Response

[20] How does the land ice sheet, exclusively in the midlatitudes and higher latitudes, communicate its influence to the tropical Atlantic? Figure 7 compares Atlantic surface wind and precipitation anomalies in the 21K LAND ICE run and a similar 21K LAND ICE run that uses fixed present-day SST instead of the slab ocean (note that the fields for the fixed SST run are subtracted from a run with the same SST and present-day land conditions, for consistency). While the midlatitude wind anomalies are qualitatively similar in the two simulations, the tropical winds are very different. Two points can be made from this comparison: (1) the North Atlantic anomalous surface circulation in the slab ocean 21K LAND ICE run is directly forced by the land ice sheet, and the thermodynamic SST feedback plays a secondary role in modifying the midlatitude surface circulation; whereas (2) the anomalous tropical circulation is almost entirely a consequence of the thermodynamic ocean-atmosphere feedback to the 21K LAND ICE forcing. Particularly noticable differences are the strengthened northeasterly trades and the emergence of a southward cross-equatorial flow and a southward shifted ITCZ. The interactive SSTs do have a quantitative effect on the extratropical surface circulation, however. Figure 7c shows the difference between the 21K LAND ICE interactive and fixed SSTs, a measure of the thermodynamical ocean

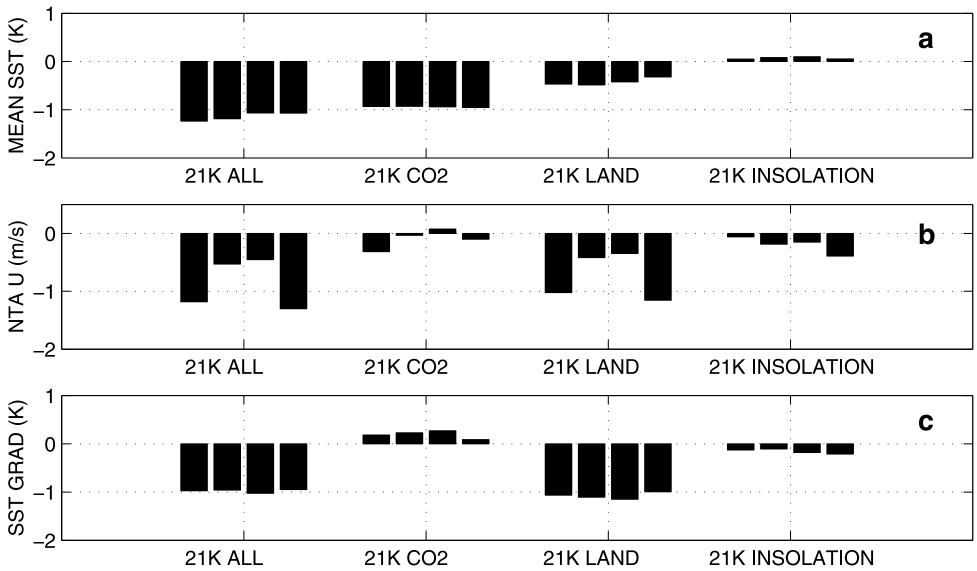

Figure 6. Summary statistics of how the 21K ALL and each individual LGM boundary condition impacts the tropical Atlantic climate. The summary statistics are for (a) tropical mean SST; (b) North tropical Atlantic trade wind strength; and (c) meridional SST gradient. The four columns for each entry represent, from left to right, December–February, March–May, June–August, and September–November averages. See the text for definitions of each statistic.

feedback: interactive SSTs generate an anomalous anticyclonic flow centered over the mid-Atlantic around 40°N, increasing the mean westerlies between 50°N and 60°N and generating anomalous easterlies between 25°N and 35°N. It is unclear, however, whether the North Atlantic atmosphere circulation is forced locally by the North Atlantic Ocean or remotely from the tropics, though we note that our North Atlantic anomalous circulation is consistent with the atmospheric response to a forced southward shift in the Atlantic ITCZ as shown by *Okumura et al.* [2001].

[21] The mechanism generating the ITCZ response in the 21K LAND ICE run resembles the dominant mechanism driving the Atlantic meridional mode on interannual-decadal timescales. In the meridional mode, increased

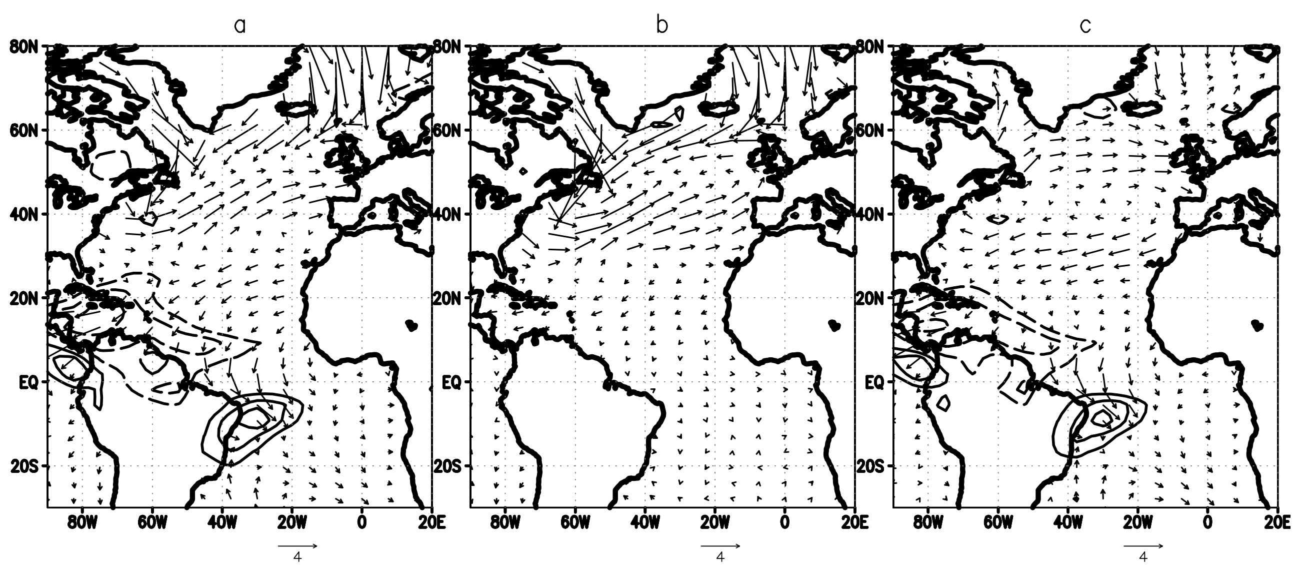

Figure 7. The impact of the land ice sheet on the North Atlantic circulation. This shows annual mean surface wind and precipitation difference of the 21K LAND ICE and control simulation, for (a) the interactive ocean and (b) fixed SST simulation. The reference wind vector is 4m/s, and precipitation CI is 1 mm/d. (c) Difference between the fields shown in Figures 7a and 7b, showing the contribution from the thermodynamic SST feedback. Note that the zero contour in precipitation is not shown.

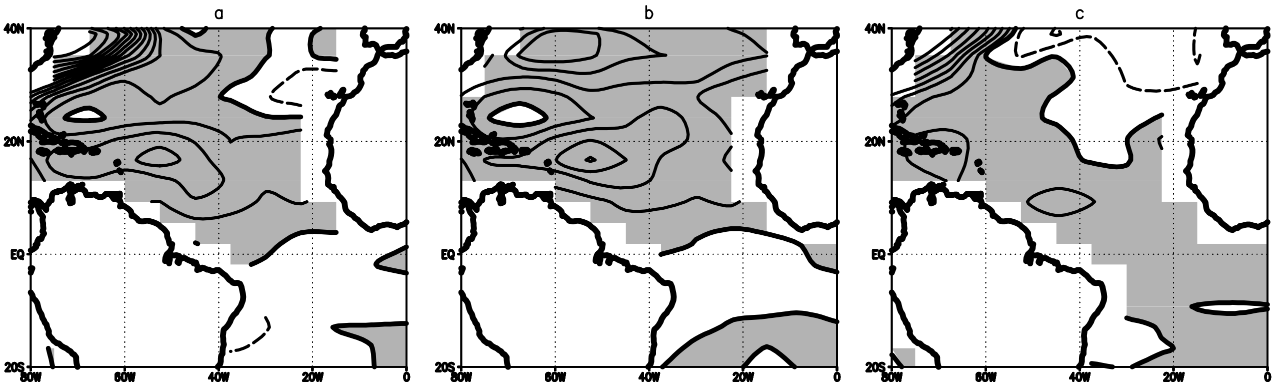

Figure 8. Surface fluxes associated with forcing the meridional mode response in the land ice simulation. (a) Annual mean latent heat flux difference between the 21K LAND ICE run with fixed SST and control. The contour interval is 5 W/m^2 and shaded regions are positive (heat flux out of the ocean). Estimates of the relative contributions of (b) wind speed and (c) air-sea specific humidity difference contributions to the latent heat flux show that the wind speed influence dominates the tropical Atlantic region.

northeasterly trades during boreal winter drive increased latent heat flux, which cools the SST down by boreal spring, causing a cooler north-warmer south SST gradient and a southward displacement of the ITCZ [*Xie and Tanimoto*, 1998]. In the 21K LAND ICE situation, latent heat flux anomalies in the fixed SST run (Figure 8a) show that circulation changes caused by the land ice forcing initiate cooling of the NTA SST; the other surface flux components (shortwave and longwave flux and sensible heat flux) are small by comparison and are not shown. Following a linearization argument made by *Saravanan and Chang* [2000], we decompose the latent heat flux anomalies into wind-driven anomalies and anomalies driven by air-sea specific humidity difference. The wind-driven component (Figure 8b) is approximated by, $F'_w = (w'/\overline{w})\overline{F}$, where w is lowest level model wind speed and F is the latent heat flux, with primes indicating anomalies and bars indicating the basic state. The humidity-driven component (Figure 8c) is approximated by $F'_q = (\Delta q'/\overline{\Delta q})\overline{F}$, where Δq is the air-sea specific humidity difference. The analysis shows that wind speed is responsible for the bulk of the latent flux anomalies that set up the SST gradient. However, air-sea specific humidity difference contributes to the generation of SST anomalies in the Caribbean region, consistent with *Saravanan and Chang*'s [2000] result. We note as an aside that the air-sea humidity difference dominates the evaporation response in the region of the Gulf Stream.

[22] The ice sheet forces the atmospheric circulation anomalies responsible for the LGM precipitation response in the tropical Atlantic. The land ice sheet can impact the atmosphere via its albedo, and topography. We identify the effect of each aspect of the ice sheet by imposing the albedo and the topography of the ice sheet in isolation. The ''ice albedo'' experiment retains the albedo of the ice over the entire horizontal extent of the land ice sheet, while setting the ice sheet thickness to zero. The ''ice topography'' experiment retains the ice sheet thickness, but sets the albedo of the ice sheet to that for present-day albedo at each grid point. Each forcing component impacts the tropical Atlantic in distinctly different ways. The ice albedo forcing (Figure 9a) increases the subtropical trades, cools the NTA SST, and displaces the ITCZ southward, as with the 21 ka land ice situation. The ice topography forcing (Figure 9b), however, reduces the northern subtropical trades, producing warmer NTA SST. The ice sheet topography is also responsible for the structure of the anomalous surface circulation over the midlatitude and polar North Atlantic, a cyclonic circulation centered off the basin around 45°N with anomalous westerlies to the south and easterlies to the north (compare Figures 9b and 7a). The impact of the Laurentide ice sheet on the North Atlantic stationary wave circulation is well known [e.g., *Manabe and Broccoli*, 1985], and this particular result is in agreement with a linear model analysis of the ice age stationary waves by *Cook and Held* [1988]. The ice albedo produces anomalous westerlies across the basin around 50°N and northeasterlies running parallel to the North American coastline, a direct response to the anomalous thermal highs over North American and polar regions.

[23] The land ice forcing is highly nonlinear, in that the ice albedo and ice topography forcings do not add linearly to give the response seen in the 21K LAND ICE run in the north and tropical Atlantic. This is especially noticable for the circulation in the NTA region, where the winds effectively cancel each other. An additional experiment, in which the ice sheet thickness was lowered by half its LGM value while the ice sheet extent and albedo were kept the same, highlights the control of the albedo over the tropical response and that of topography over the midlatitude response (Figure 9c): the northeasterly trade and cross-equatorial flow anomalies remain about the same (albeit slightly weaker) as topography is lowered, but the midlatitude circulation anomalies completely change to resemble more the ice albedo simulation (Figure 9a) and less the ice topography simulation (Figure 9b).

5. Sensitivity to Ocean Heat Transport

[24] The lack of interactive ocean heat transport (OHT) and sea ice coverage in our model precludes a meaningful

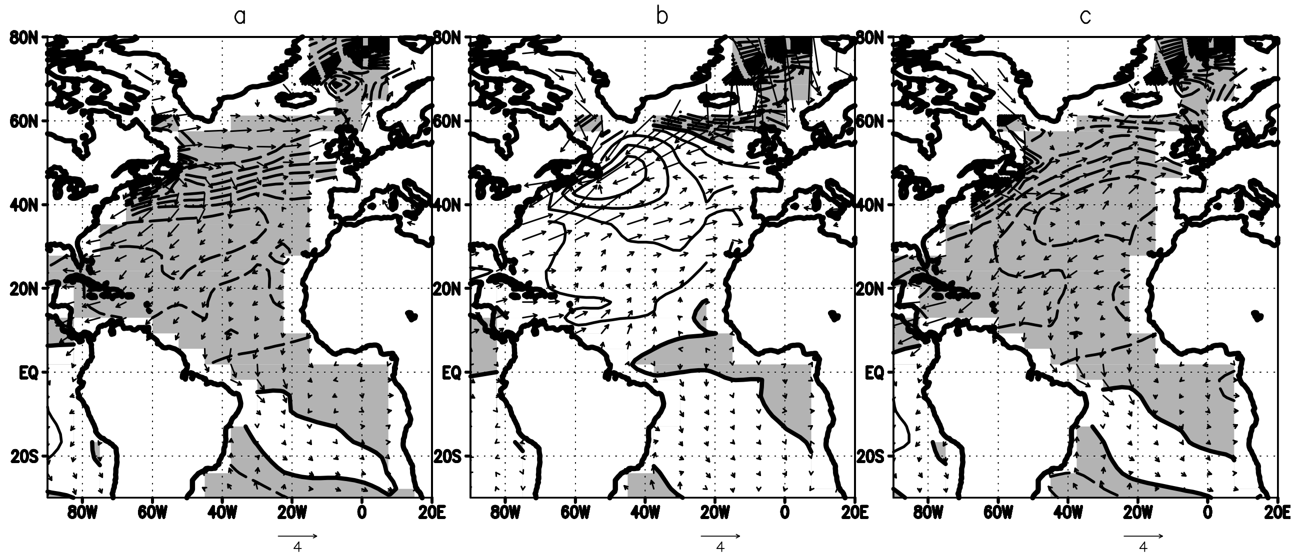

Figure 9. Relative roles of the ice albedo and ice topography components of the land ice sheet in forcing the response in Figure 7. Annual mean SST (CI 0.5 K) and surface wind (4 m/s reference vector) difference between the (a) ice albedo, (b) ice topography, and (c) 21 ka land ice with ice thickness reduced by 50% and the control run. Thick contour is the zero contour, and negative regions are shaded.

comparison between our model results and paleorecords of the tropical Atlantic climate. However, we can estimate the sensitivity of the Atlantic ITCZ to changes in OHT (this section) and sea ice (section 6) by imposing them in a meaningful way. Current understanding of modern-day variability suggests that changes in both the tropical Pacific and in the thermohaline circulation in the Atlantic can potentially affect the Atlantic ITCZ:

[25] 1. Changes in the Atlantic thermohaline circulation influence surface OHT convergence changes in the tropical oceans through ocean planetary wave propagation [e.g., *Yang*, 1999; *Johnson and Marshall*, 2002]. In particular, recent coupled GCM studies of simulated North Atlantic thermohaline circulation shutdown by *Vellinga and Wood* [2002] and *Dong and Sutton* [2002] demonstrates the potentially potent effect of a thermohaline circulation shutdown on Atlantic ITCZ position.

[26] 2. The impact of ENSO on the Atlantic ITCZ variability is well established [e.g., *Curtis and Hastenrath*, 1995]. In particular, ENSO is known to have significant control over the variability of the northeasterly trades and hence the NTA SST anomalies [*Saravanan and Chang*, 2000]. Changes in the mean equatorial zonal SST gradient in the Pacific from coupled ocean-atmosphere processes, similar to changes that occur with the El Niño-Southern Oscillation (ENSO), may also affect the Atlantic paleo-ITCZ.

[27] There is observational and model evidence for change in the Atlantic thermohaline circulation, but quantitative estimates thereof remain poorly constrained. We use the ocean model estimate by *Fichefet et al.* [1994], who forced a two-dimensional ocean model with output from an AGCM run with LGM boundary conditions and obtained an estimate of the zonal mean Atlantic OHT induced by a weakened LGM thermohaline circulation. We took the difference of this annual mean OHT estimate from present-day model estimates, computed the OHT convergence and applied it uniformly in time and longitude across each latitude band across the Atlantic from 40°S to 80°N as additional Q flux. This amounts to the assumption that all OHT divergence is in the upper 50 m of the ocean. The resulting additional Q flux is shown in Figure 10a.

[28] Observational estimates of the Pacific equatorial zonal gradient in SST during LGM are so poorly constrained that even the sign remains in dispute. A "no-analog" reconstruction of LGM SST [*Mix et al.*, 1999] produced increased Pacific zonal gradients, whereas another recent estimate [*Koutavas et al.*, 2002] reduced the LGM gradient by about 1 K over the width of the basin. In the face of such uncertainty, we appealed to our simulations for clues on how to proceed. The 21K ALL run (land ice, CO_2, and orbital forcing) produces SST anomalies in the tropical Pacific that weaken the zonal gradient by 1–2 K across the width of the basin, and shift the ITCZ in the central Pacific southward, an El Niño-like response. We estimated what the ENSO dynamics might do with this initial SST change by applying these SST anomalies on an intermediate coupled ocean-atmosphere model for the tropical Pacific [*Battisti*, 1988]. The LGM influence was applied to the intermediate coupled model by linearly damping the coupled model anomalous mixed layer temperature to the 21K ALL run SST anomaly with a timescale of 125 days (125 days is also the Newtonian damping timescale used by the intermediate coupled model ocean mixed layer and serves as a proxy for surface flux damping). The intermediate coupled model amplified the initial SST anomalies to produce a mean warming of the eastern equatorial Pacific with the same spatial structure as the ENSO SST anomalies. We transferred the tropical Pacific feedback to the AGCM-SOM by modifying the original monthly mean Q flux with the OHT

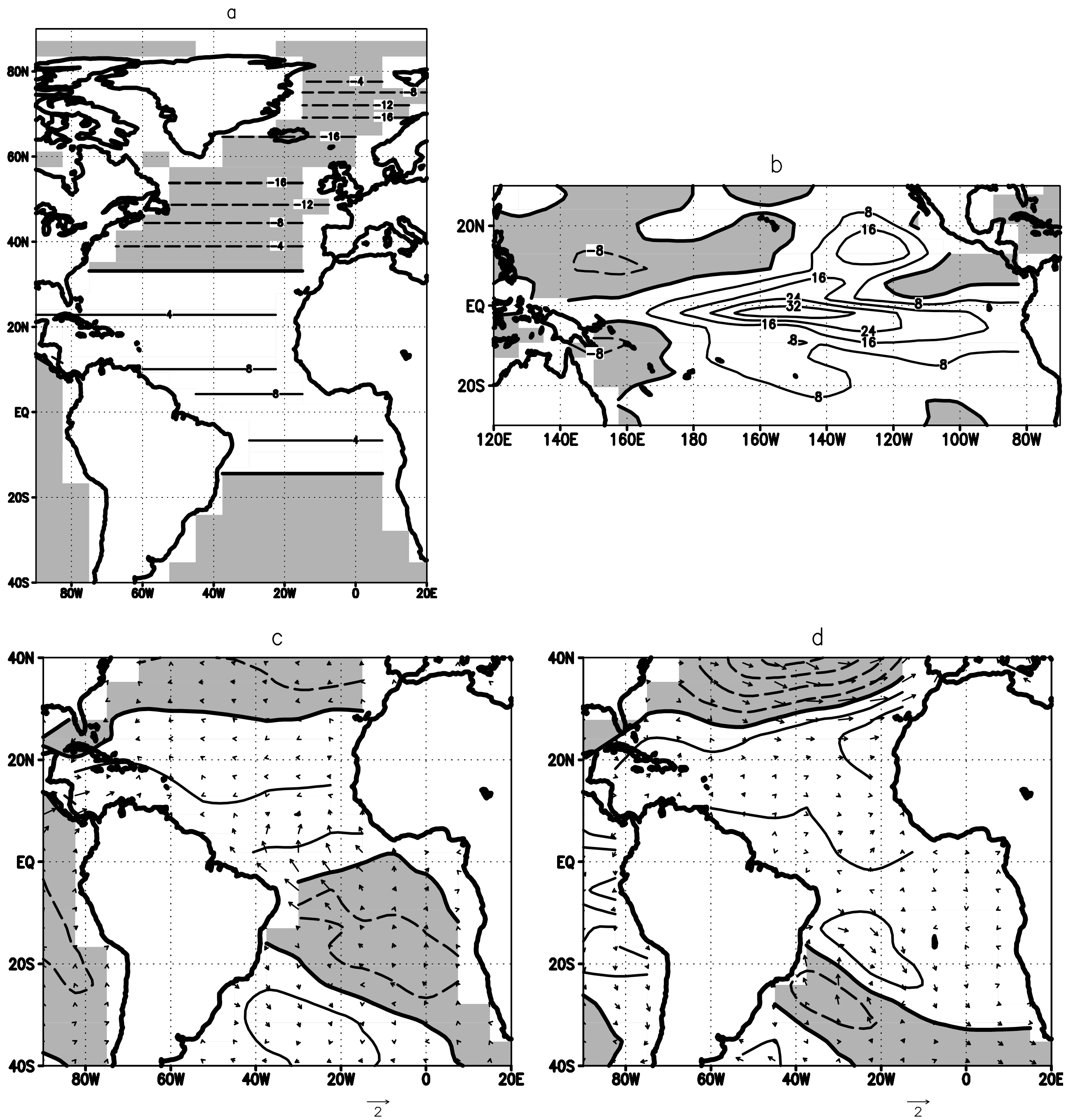

Figure 10. Testing the sensitivity of the tropical Atlantic climate to OHT changes associated with a permanent El Niño and weakening Atlantic thermohaline circulation. Additional Q flux applied to the slab ocean model for the (a) thermohaline OHT change and (b) reduced equatorial zonal SST gradient. CI is 4 W/m^2, and negative regions (heat extracted from the ocean) are shaded. The values shown are annual means. Response in the tropical Atlantic to these forcings: annual mean difference of SST (CI 0.3 K) and surface wind (reference wind vector 2 m/s) between the (c) thermohaline OHT and control and (d) reduced Pacific zonal SST gradient and control. The thick contour is zero contour, and negative regions are shaded.

anomalies derived from the coupled model run. The additional Q flux (annual mean) is shown in Figure 10b. The AGCM-SOM run with this modified tropical Pacific Q flux reduces the across-basin equatorial Pacific SST gradient by an additional 2 K across the entire basin (not shown), a permanent El Niño-like response.

[29] To assess the impact of each OHT perturbation, we added each one separately, in addition to the forcing applied

for the 21K ALL run, and compared them to the 21K ALL case. The Atlantic thermohaline OHT change produces a meridional mode response that shifts the Atlantic ITCZ northward (Figure 10c), reducing the anomalous SST gradient set up by the 21K ALL run by about 0.5 K. While the applied OHT also causes a cooling of North Atlantic SST by about 2 K, there is minimal surface circulation response throughout the entire North Atlantic outside the deep tropics; in other words, the tropical meridional mode response does not come about through changes to the trade wind perturbations (as for the land ice response) nor through forcing from the North Atlantic. Rather, the meridional mode response appears to come directly through the applied OHT convergence; the slightly larger OHT convergence in the NTA region relative to the south tropical Atlantic is apparently sufficient to cause a meridional SST gradient across the equator.

[30] The permanent El Niño-like forcing, on the other hand, does not influence the meridional ITCZ position in the Atlantic (Figure 10d), in contrast to what seen in modern-day interannual variability. Rather, it produces a weak but significant warming of the entire tropical Atlantic basin, with only slightly higher amplitude in the northern subtropical Atlantic. There is a strong anomalous westerly response at the northern subtropical latitudes (~30°N) that might suggest a surface flux forcing on the NTA SST. However, the anomalous winds are located at the boundary between the trades and the midlatitude westerlies and not in the trades themselves, apparently too far north to impact the surface fluxes in the NTA region. The basinwide tropical Atlantic SST warming appears consistent with a mechanism proposed by *Chiang and Sobel* [2002] for ENSO impact over the remote tropics through warming of the tropical troposphere. The zonally averaged tropospheric temperature warms over the model tropics in response to the permanent El Niño condition (albeit small ~0.4°C in the mid and upper troposphere), similar to what is seen during present-day El Niño events. This warming, according to Chiang and Sobel, is communicated through convective processes to warming over the tropical oceans outside the Pacific source region.

6. Sensitivity to Sea Ice Changes

[31] Sea ice turns out to be a major influence on Atlantic ITCZ position. We applied a crude representation of LGM sea ice to the model in the following way: from the 21K ALL run we identified those regions where the monthly mean SST dipped below 0°C and marked them as additional sea ice for the subsequent run. Note that this prescription likely underestimates the extent of sea ice increase because it omits the positive feedback of the additional sea ice on itself; however, we chose a higher (0°C) temperature threshold than the ocean water freezing temperature of −1.8°C partly to compensate for this. In January (Figure 11a), additional sea ice forms in the North Atlantic region and throughout the Southern Ocean, and in July (Figure 11b) it forms exclusively over the Southern Ocean. The influence of this additional sea ice on the Atlantic is shown in Figure 11c, which shows SST and surface wind anomalies relative to the 21K ALL run. The net effect of the sea ice feedback, like that for the thermohaline OHT change, is to reduce the anomalous SST gradient set up by the 21K ALL run by ~0.5 K.

[32] To understand the role played by the sea ice coverage in each hemisphere, we ran two additional experiments applying the sea ice coverage changes in one hemisphere only. Figures 11d and 11e show the change in annual mean surface temperature and winds over the tropical Atlantic due to Northern Hemisphere-only and Southern Hemisphere-only sea ice changes, respectively. In both cases, significant cooling of surface temperature occurs in the hemisphere with sea ice, and a somewhat weaker cooling occurs in the opposing hemisphere. However, it is primarily the Southern Hemisphere sea ice that elicits the cross equatorial flow response in the original run, possibly due to the larger sea ice coverage changes in the Southern Hemisphere relative to the north.

[33] How does the atmosphere communicate the sea ice influence to the tropical Atlantic? This is still under investigation and will be elaborated in a future paper, so we will not go into details here. However, it is clear from Figure 11 that like the thermohaline OHT influence and unlike the land ice influence, the sea ice influence is not mediated through variations in the subtropical trades. Rather, we think it arises through sea ice influence on the mean hemispheric air temperature and humidity; specifically, the more sea ice there is in one hemisphere, the cooler and drier that hemisphere. The greater cooling and drying in the Southern Hemisphere relative to the Northern Hemisphere in the simulation with additional sea ice cover provides the hemispheric asymmetry necessary for the Atlantic ITCZ response. In support of this interpretation, we note that precipitation in the Pacific and Indian Oceans is also displaced northward in the sea ice run relative to the 21K ALL run (not shown).

[34] The effect of interhemispheric difference in radiative forcing on the global circulation has been noted before, notably by *Manabe and Broccoli* [1985], and more recently by *Broccoli* [2000] in the context of the LGM land ice sheet.

7. Orbital Forcing on the Tropical Atlantic

[35] For completeness, we investigate the sensitivity of the Atlantic ITCZ to precessional forcing. *Clement et al.* [2003] have argued, also from a AGCM-slab ocean perspective, that the precessional forcing is equally important to the tropical hydrological cycle as the influence of land ice. Since the orbital configuration at 21 ka is similar to the present one, the LGM case does not reveal the full potential for orbital changes to influence the Atlantic ITCZ. We investigate the effect of 6 ka orbital parameters; the change in precession affects the seasonal hemispheric gradients in insolation (Figure 12a) but not the annual mean gradient (Figure 12b). Annual mean precipitation changes from a run with 6 ka orbital forcing, and otherwise, modern-day conditions are presented in Figure 12c. The change does not resemble a meridional mode response (at least, in the annual mean): rather, the annual mean response shows a slight (0.5–1 mm/d) precipitation strengthening at 10°N and 5°S,

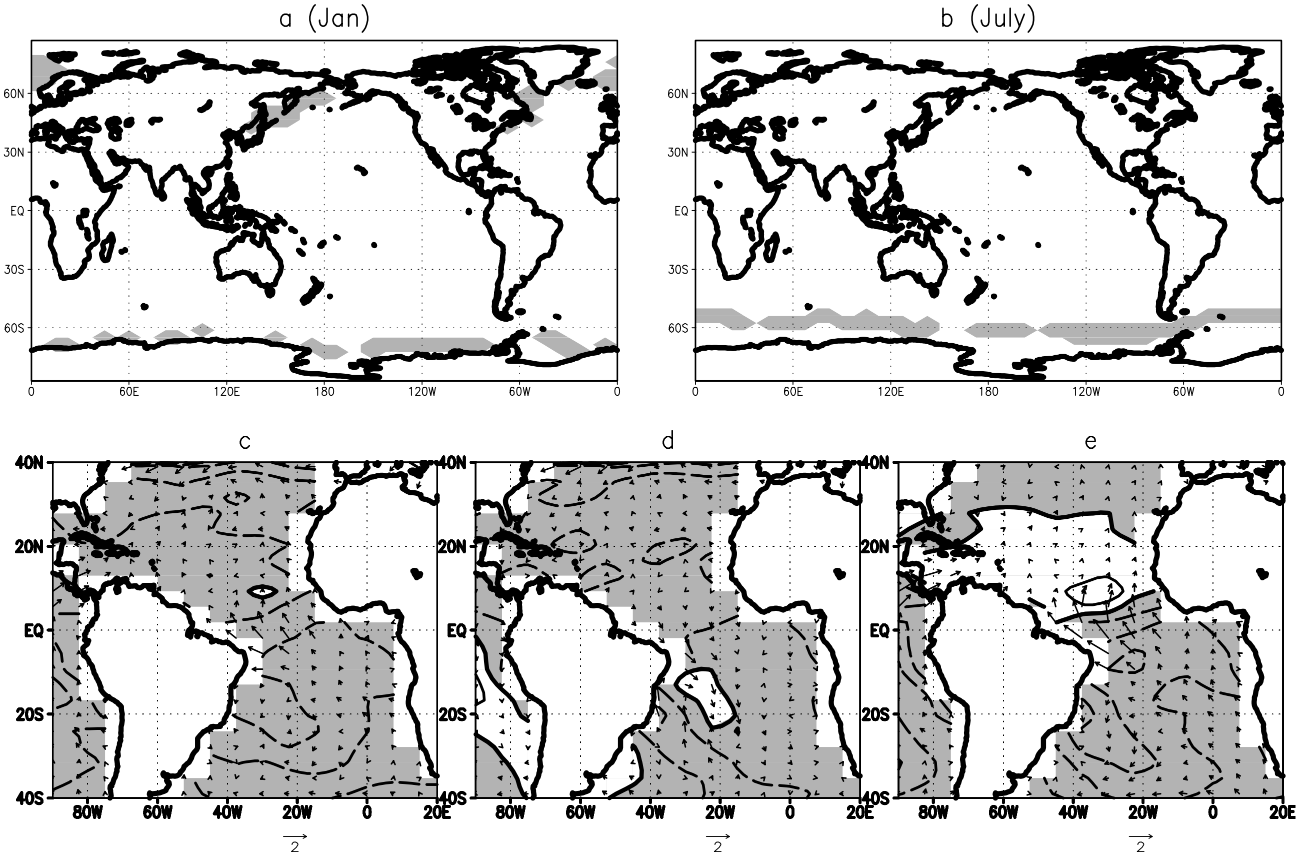

Figure 11. Testing the sensitivity of the tropical Atlantic to change in sea ice cover. Change in the sea ice coverage for the 21K ALL with sea ice feedback simulation relative to the 21K ALL simulation for (a) January and (b) July. (c) Annual mean SST (CI 0.3 K) and surface wind (reference vector 2 m/s) difference between the 21K ALL with sea ice and 21K ALL simulation. (d) Same as Figure 11c, but with just the Northern Hemisphere sea ice imposed; and (e) same as Figure 11c but with just the Southern Hemisphere sea ice imposed. In panels Figures 11c–11e, the thick contour is zero contour, and negative regions are shaded.

and weakening over the mean ITCZ latitude of around 5°N (Figure 12c). The SST and wind responses (not shown) are weak, relative to the other perturbation experiments. The interesting ITCZ feature is over the West African region, where there is a clear northward displacement of the land ITCZ.

[36] We think that the difference between the land and ocean responses has to do with the different surface thermal inertia in land and ocean and also with the seasonality of the mean ITCZ behavior. The oceanic ITCZ does have a strong seasonal response as might be expected from seasonally strong hemispheric gradients in insolation (Figure 12a), but such response is cancelled out to a significant degree in the annual mean (Figure 12c), in sympathy with the zero annual mean equatorial meridional insolation gradient (Figure 12b). The seasonal displacement of the ITCZ appears to phase lag the changes in insolation gradient. The strongest anomalous oceanic ITCZ displacement to the south occurs during March through June (Figure 12d), 6 months after the maximum cooling of the north tropical Atlantic and warming of the south tropical Atlantic by the anomalous insolation. This phase difference is caused in part by the relatively large ocean thermal inertia, which ensures that ocean SST does not equilibrate relative to the seasonal insolation changes. This interpretation is supported by an additional run where we reran both the 6 ka experiment and the control run using a 1 m mixed layer depth (not shown). There, the SST was roughly equilibrated to the insolation, and the seasonal ITCZ displacement was pronounced and in phase with the applied hemispheric insolation gradient anomaly.

[37] The West African anomalous land ITCZ response in boreal summer is in phase with the anomalous insolation (Figure 12e), given the small surface thermal inertia of land. It remains to be explained why the land ITCZ displacement survives in the annual mean. We think it is simply due to the fact that there is no convection there in boreal winter, so that insolation perturbations have little impact on the amount of boreal winter rainfall. In contrast, the latitude range of oceanic ITCZ migration is smaller, and the seasonal reversal in the anomalous insolation gradient can achieve a significant degree of cancellation in the annual mean response.

[38] Finally, it is interesting to contrast this 6 ka insolation case, which presents a strong West African ITCZ response but a weak tropical Atlantic response, to the 21K LAND

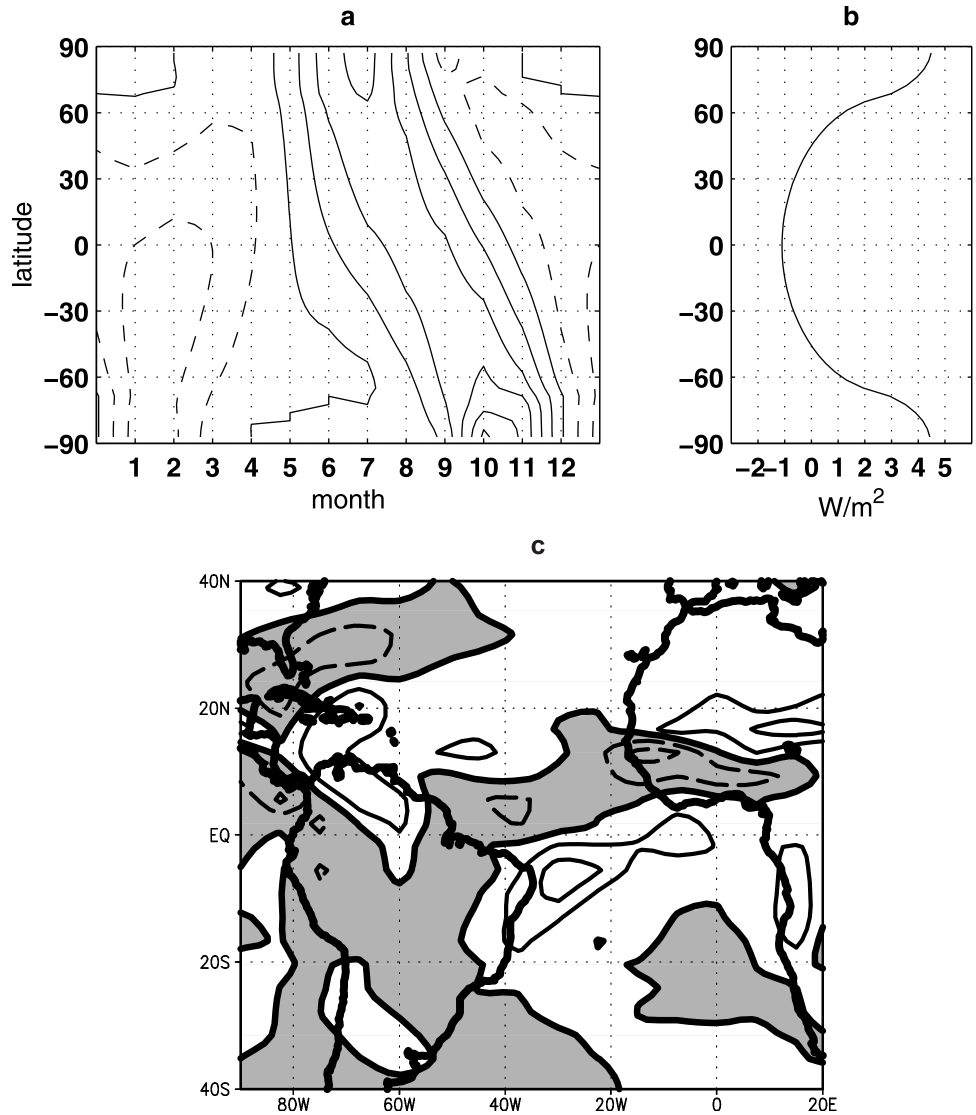

Figure 12. Sensitivity of the tropical Atlantic to precessional changes to the insolation. (a) Top of atmosphere difference in the model incoming shortwave insolation between the 6 ka run and control. CI is 10 W/m^2, and the thick line is zero contour. (b) Annual mean top of atmosphere difference in incoming shortwave insolation 6 ka minus present-day. (c) Annual mean difference between the 6 ka and control precipitation (CI 0.5 mm/d). Note that the CI for precipitation is half that for the other precipitation plots in this paper, indicating the relatively weak nature of the annual mean precipitation response to 6 ka orbital forcing. The thick line is zero contour, and negative regions are shaded. (d) Latitude-time contours for precipitation, 6 ka minus present-day, for the mid-Atlantic (30°W). The contour interval is 1 mm/d, negative contours are dashed, and the zero contour is in bold. (e) Same as Figure 12d, but for West Africa (0°E). Figures 12d and 12e illustrate the seasonal nature of the precipitation response to precession.

ICE case, which presents the opposite scenario. We think that the difference in the land and ocean thermal inertia plays a crucial role in determining such opposite responses. The oceanic ITCZ, because of its slower response time and sensitivity to SST gradients, responds more effectively to small perturbations in hemispheric asymmetry over the annual mean, like the impact of the land ice sheet on the tropical SST. The land ITCZ, because of its faster response time, can respond effectively to seasonal forcing in north-south contrasts (like precessional perturbations to insolation).

8. Summary and Discussion

[39] Paleoevidence from the last glacial period and deglaciation show significant changes in the position of the Atlantic ITCZ that occur concomitantly with climate changes in the North Atlantic. We explore the hypothesis

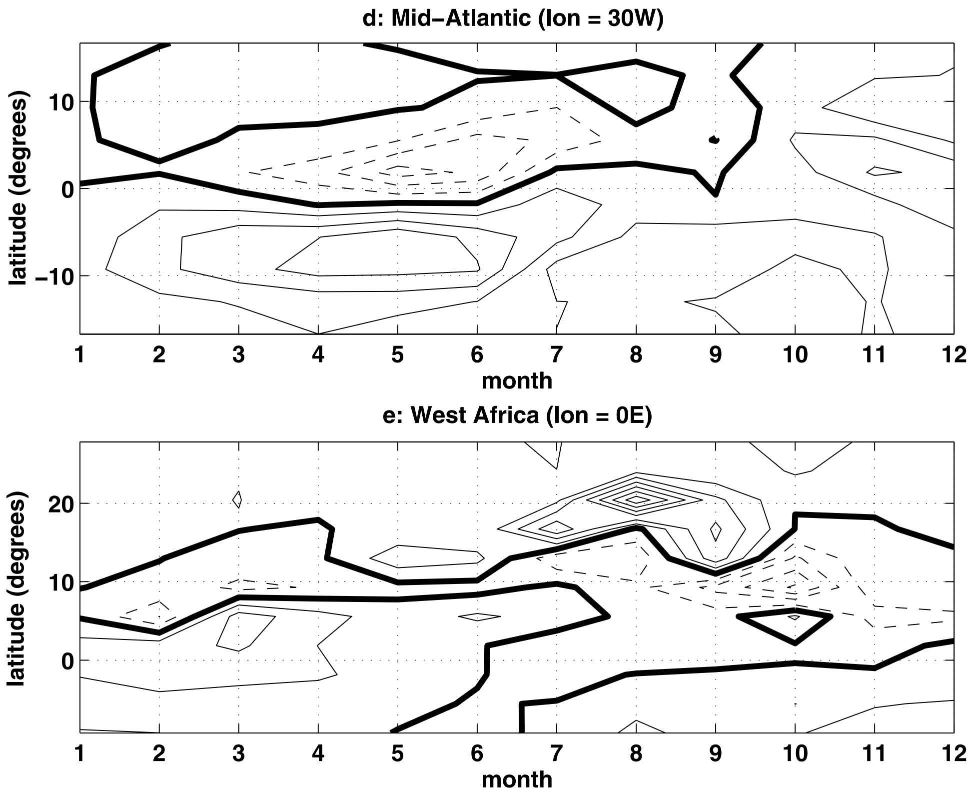

Figure 12. (continued)

that past changes in the tropical Atlantic mean state can be produced by the same processes that produce the modern-day leading mode of interannual-to-decadal variability in the tropical Atlantic, namely, the meridional mode. This modern-day analog suggests that an AGCM coupled to a slab ocean model should be capable of reproducing the Atlantic response to paleoclimate forcings. Mechanisms that can affect the ITCZ but are not captured by the model physics, such as changes in sea ice cover and ocean heat transport, are addressed through sensitivity experiments. We use CCM3 coupled to a slab ocean model with fixed 50 m depth, and with a prescribed OHT tuned to reproduce modern-day mean SSTs in our control run. The model is capable of qualitatively reproducing the year-to-year variations in the Atlantic ITCZ and near-equatorial meridional SST gradient as part of its internal variability.

[40] We applied the following three forcings to simulate the 21 ka climate: carbon dioxide concentration to 200 ppm, land ice sheet, and orbital changes. Apart from a $\sim$1 K uniform cooling of tropical basinwide SST due to the reduction of CO_2, the dominant response of the tropical Atlantic to these LGM forcing resemble the meridional mode response. The land ice sheet forcing is responsible for virtually all of the meridional SST gradient change. The land ice sheet communicates with the tropics through the perturbation of the North Atlantic stationary wave circulation by the Laurentide ice sheet, which produces a strengthening of the NTA trades. The tropical ocean-atmosphere interactions associated with the meridional mode amplify this initial perturbation, cooling the mean NTA SST and displacing southward the Atlantic ITCZ. Further experiments suggested that it is the albedo component of the land ice that shifts the ITCZ southward, although the decomposition into albedo and topography effects is nontrivial and the responses to individual forcing components do not add linearly to give the land ice sheet response.

[41] We performed sensitivity experiments to assess the effect of LGM ocean heat transport (OHT) changes associated with permanent ENSO conditions and with a weaker Atlantic thermohaline circulation and the effect of sea ice coverage changes. We find that the permanent ENSO does not affect the Atlantic ITCZ, while both the weaker thermohaline circulation and the hemispheric asymmetric expansion of sea ice force a significant meridional mode response that acts to shift the Atlantic ITCZ to the north, in opposition to the southward shift induced by the LGM land ice sheet. Furthermore, the mechanisms forcing the meridional mode response to the thermohaline OHT and sea ice coverage changes are not through northern tropical trade wind changes. Preliminary results suggest that differing sea ice coverage changes in the northern and southern polar regions cause differential changes to hemispheric temperatures and thus are able to displace the Atlantic ITCZ, but this hypothesis is still under investigation.

[42] Finally, we tested the Atlantic response to changes in insolation by imposing a 6 ka precession. The Atlantic ITCZ responds with large seasonal shifts but small annual mean changes. By contrast, the annual mean West African ITCZ is displaced to the north relative to present-day. We attributed the difference in behavior between the land and

the ocean ITCZs to the different response time and memory in the land surface and the oceanic mixed layer.

[43] How do the model results contribute to the interpretation of paleorecords of Atlantic ITCZ variability? Our model results in general support the interpretation of the Cariaco Basin paleoproxy climate studies that infer changes in the position to the ITCZ. Similar to the modern day variations in tropical Atlantic climate, a preferred mode of tropical Atlantic response to various LGM boundary conditions (land ice sheet, thermohaline OHT, and sea ice in particular) appears to be a displacement of the ITCZ and associated changes to SST and atmospheric circulation. A weak uniform tropical basinwide SST warming or cooling (as exhibited by the model response to CO_2 and to the reduced zonal SST gradient in the tropical Pacific) may be another such preferred mode of behavior in the tropical Atlantic. The lack of any sizable surface circulation or precipitation changes associated with the Atlantic tropical basinwide SST change suggests that this response may be difficult to detect from proxy records.

[44] The results from the experiments with prescribed changes in land and sea ice coverage and in ocean heat transport suggest that the synchronicity in the tropical and high-latitude records may be due to the high latitudes driving the variability in the tropical Atlantic, both at the slow glacial-interglacial timescale and in the context of rapid climate change. In order to substantiate this suggestion, let us compare the model response to land ice, sea ice, and thermohaline ocean heat transport with the paleo record.

[45] In agreement with *Dong and Sutton* [2002] and *Vellinga and Wood* [2002], our results suggest that thermohaline circulation can exert a strong influence on the Atlantic ITCZ position, either directly through tropical Atlantic OHT changes or indirectly through its influence on sea ice. These results lend support to the hypothesis that thermohaline variations are the cause for the observed covariability in the Greenland and Cariaco basin record in rapid climate change. Our reasoning is as follows. Consider the onset of a Dansgaard-Oeschger (D/O) event. If the onset of a D/O event is indeed due to a surge in the OHT to the North Atlantic and consequent reduction in sea ice, these two processes will have opposite effects in the tropical Atlantic: the direct OHT influence shifts the ITCZ southward, while the reduced sea ice shifts the ITCZ northward. We know from paleoproxy records that warm stadial events over Greenland (stronger overturning circulation [e.g., *Rahmstorf*, 2002]) correspond to a northward shifted ITCZ [*Peterson et al.*, 2000]. Our sensitivity studies imply that the impact of sea ice wins out over the direct OHT influence on the tropical Atlantic. We are, however, not confident that the OHT changes we apply is necessarily correct for a weakened Atlantic thermohaline state. In our case, the ITCZ shifts north; however, *Yang* [1999] suggests the opposite for a weakened thermohaline state, based on transient results of an idealized ocean general circulation model (OGCM) study. It is beyond the scope of this study to resolve this apparent discrepancy, though we note that there are many differences between the Fichefet et al. and the Yang studies that preclude direct comparison; the former studied the LGM climate using a zonally averaged OGCM, whereas the latter examined the impact of present-day Laborador deep water formation changes using a full ocean GCM (albeit with idealized boundaries). However, the point that Atlantic ITCZ is sensitive to thermohaline OHT changes remains valid.

[46] The chain of events outlined above to explain the covariability of North Atlantic sea ice and tropical Atlantic precipitation in terms of high-latitude control of the tropics hinges on the hypothesis, yet unproven, that the ocean heat transport determines sea ice coverage. If instead we assume that the sea ice extent is determined by the surface winds stress (either directly or through its influence on the northward extent of the gyre circulation and ocean heat transport), we can interpret the observational record in an alternative way. Let us assume that for some unknown reason, the mean position of the ITCZ shifts to the south. The remote response to such a change in the tropical heating will change the surface wind in the Atlantic high latitudes: indeed, *Okumura et al.* [2001] show GCM evidence for tropical Atlantic ITCZ influence on the NAO. Figure 7c suggests that the changes would be in the direction of increased surface wind speed, which causes stronger heat fluxes from the ocean, and of a southward Ekman transport. The net effect of a southward shift in the ITCZ would thus likely be a southward extension of North Atlantic sea ice. What effect, if any, a shift of the ITCZ can have of the Atlantic OHT cannot be established with our modeling framework. The shift in the sea ice would, in turn, influence the location of the ITCZ. Our modeling results may therefore be consistent with contrasting interpretations of the observational data: one that underscores how the high latitudes can force changes in tropical precipitation and another that underscores the opposite. Further modeling work is needed to sort out which interpretation is more relevant for variability at different paleo timescales and what the role of the feedbacks is.

[47] Does any paleoproxy evidence suggest a connection between ice volume and Atlantic ITCZ displacements? *Harris and Mix* [1999] conclude from a sediment record from the Ceara rise that the dominant variability in the position of Amazon convection over the last million years follows a $\sim$100 kyr cycle characterized by a more southward position during glacial periods, and is remarkably synchronous with the ice volume record. To the extent that Atlantic ITCZ variations can be translated into Amazon convection variations, we suggest that this link is causal: the land ice directly influences the atmospheric circulation in the midlatitudes and subtropics, and the coupled atmosphere-ocean meridional mode physics in the tropical Atlantic brings this influence into the tropics and displaces the ITCZ. We note, however, that Harris and Mix do not themselves interpret their record in this way because they find an apparent lead in the precipitation data relative to the ice volume data. They suggest instead that Amazonian rainfall follows the 100 kyr cycle because of a nonlinear amplification of insolation forcing at precessional frequencies, and Amazon rainfall changes, in turn, drive variations at higher latitudes that ultimately bring about the glacial-interglacial variations.

[48] The surprising nonlinearity of the land ice sheet influence on the atmospheric circulation over the tropical Atlantic, in particular, the opposing effects of albedo and topography, may have interesting implications for Atlantic climate during periods of significant ice sheet evolution. Could rapid transitions in the climate regime over the tropical and North Atlantic occur with gradual growth of the Laurentide ice sheet? This may well be an interesting avenue of research to explore.

[49] We note that while our model results show that precessional variations do not induce a shift of the annual mean position of the Atlantic ITCZ, paleorecords suggest otherwise. *Haug et al.* [2001] infer from titanium and iron concentration data from the Cariaco Basin that the ITCZ was shifted northward during the Holocene ''thermal maximum'' between 5 and 10 ka, after which it drifted gradually southward to its present-day position. Our model 6 ka simulation does not show marked change to the annual mean ITCZ position, though it shows a marked seasonal response. Since the surface ocean response to seasonal insolation changes is generally not equilibrated, a rectifying effect, as *Haug et al.* [2001] argue, is possible. One possibility is that our model, which uses a uniform 50 m mixed layer depth and does not model ocean heat transport changes, cannot capture the different characteristic response time of the surface ocean at different locations. Consequently, the response of the ITCZ to anomalous insolation may be stronger at a certain point of its seasonal march than at another, and a rectified signal may emerge.

[50] The ability for sea ice coverage change to significantly impact ITCZ position in the Atlantic (and of the marine tropics as a whole) is a surprising result, as there is no modern analog for this connection, neither from a observational viewpoint (sea ice coverage is known to vary with ENSO, but the causal influence is thought to be from the tropical Pacific to the polar regions) nor from a mechanistic viewpoint. The relevance of sea ice variations to climate variability in the tropics, and how the sea ice manages to communicate its influence there, are very much open questions.

[51] We conclude with a cautionary note on the omission of equatorial ocean dynamics in our study. The tropical Atlantic is not dominated by any single mode of climate variability as ENSO does over the tropical Pacific (eloquently discussed by *Sutton et al.* [2000]). The Atlantic Niño has sizable control of the climate variability over the equatorial Atlantic and neighboring land regions, especially along the Gulf of Guinea and also along the equatorial Amazon regions [*Zebiak*, 1993]. It is thought to arise from a coupled ocean-atmosphere processes similar to ENSO and characterized by a warm (cold) SST anomaly over the Atlantic cold tongue region, anomalous surface wind convergence (divergence) into this SST anomaly, and intensification (reduction) of convection over the warm (cold) anomaly. It is likely that this mode will also respond to paleoclimate forcing, potentially complicating the tropical Atlantic response. This mode could arise in the following ways: (1i) from interaction with the meridional mode to give rise to a combined behavior [e.g., *Servain et al.*, 1999]; (2) forced by ENSO [e.g., *Latif and Barnett*, 1995], in which case it might provide a larger response in the tropical Atlantic to the permanent El Niño condition than what a slab ocean could provide; and (3) from amplifying the tropical Atlantic response to orbital forcing, given that the West African monsoon responds to orbital forcing and plays a substantial role in setting up the Atlantic cold tongue [e.g., *Li and Philander*, 1997; *Xie and Saito*, 2001]. This being said, two recent fully coupled model simulations of the LGM [*Bush and Philander*, 1999; *Kitoh and Murakami*, 2002] appear to show meridional mode-like change in the tropical Atlantic relative to the present-day simulations, with increased north tropical Atlantic trades and colocated cooler SST anomalies relative to the south tropical Atlantic (neither paper discusses their tropical Atlantic response at any length, though the responses are discernable from the LGM minus present-day fields plotted for surface wind and SST). These simulations suggest that the meridional mode-like response still dominates tropical Atlantic changes to LGM conditions, despite the presence of ocean dynamical feedback.

[52] **Acknowledgments.** Special thanks to Jeff Yin for his help with setting up the model boundary conditions and advice on designing the model experiments. Our paper benefited from exchange with Cecile Bitz, Christa Farmer, Tom Koutavas, Camille Li, Gerard Roe, Dan Vimont, and two anonymous reviewers. J.C.H.C. was supported through a NOAA Climate and Global Change postdoctoral fellowship, administered by the University Corporation for Atmospheric Research, and NSF grant ATM-99-86072; and M.B. and D.S.B. are funded by the Joint Institute for the Study of the Atmosphere (JISAO) under NOAA Cooperative agreement NA17RJ1232, contribution 1003.

References

Battisti, D. S., Dynamics and thermodynamics of a warming event in a coupled tropical atmosphere ocean model, *J. Atmos. Sci.*, *45*, 2889–2919, 1988.

Berger, A. L., Long-term variations of caloric insulation resulting from Earth's orbital elements, *Quat. Res.*, *9*, 139–167, 1978.

Biasutti, M., Decadal variability in the tropical Atlantic as simulated by the Climate System Model and the CCM3 coupled to a slab ocean model, M.S. thesis, 43 pp., Dep. of Atmos. Sci., Univ. of Wash., Seattle, 2000.

Black, D. E., L. C. Peterson, J. T. Overpeck, A. Kaplan, M. N. Evans, and M. Kashgarian, Eight centuries of North Atlantic Ocean atmosphere variability, *Science*, *286*, 1709–1713, 1999.

Bretherton, C. S., C. Smith, and J. M. Wallace, An intercomparison of methods for finding coupled patterns in climate data, *J. Clim.*, *5*, 541–560, 1992.

Broccoli, A. J., Tropical cooling at the Last Glacial Maximum: An atmosphere-mixed layer ocean model simulation, *J. Clim.*, *13*, 951–976, 2000.

Bush, A. B. G., and S. G. H. Philander, The climate of the Last Glacial Maximum: Results from a coupled atmosphere-ocean general circulation model, *J. Geophys. Res.*, *104*, 24,509–24,525, 1999.

Chang, P., L. Ji, and H. Li, A decadal climate variation in the tropical Atlantic Ocean from

thermodynamic air-sea interactions, *Nature*, *385*, 516–518, 1997.
Chang, P., R. Saravanan, L. Ji, and G. C. Hegerl, The effect of local sea surface temperatures on the atmospheric circulation over the tropical Atlantic sector, *J. Clim.*, *13*, 2195–2216, 2000.
Chiang, J. C. H., and A. H. Sobel, Tropical tropospheric temperature variations caused by ENSO and their influence on the remote tropical climate, *J. Clim.*, *15*, 2616–2631, 2002.
Chiang, J. C. H., Y. Kushnir, and A. Giannini, Deconstructing Atlantic Intertropical Convergence Zone variability: Influence of the local cross-equatorial sea surface temperature gradient, and remote forcing from the eastern equatorial Pacific, *J. Geophys. Res.*, *107*(D1), 4004, doi:10.1029/2000JD000307, 2002.
Clement, A. C., A. Hall, and A. J. Broccoli, The importance of precessional signals in the tropical climate, *Clim. Dyn.*, in press, 2003.
Cook, K. H., and I. M. Held, Stationary waves of the ice age climate, *J. Clim.*, *1*, 807–819, 1988.
Curtis, S., and S. Hastenrath, Forcing of anomalous sea-surface temperature evolution in the tropical Atlantic during Pacific warm events, *J. Geophys. Res.*, *100*, 15,835–15,847, 1995.
Czaja, A., P. van der Vaart, and J. Marshall, A diagnostic study of the role of remote forcing in tropical Atlantic variability, *J. Clim.*, *15*, 3280–3290, 2002.
Dong, B. W., and R. T. Sutton, Adjustment of the coupled ocean-atmosphere system to a sudden change in the thermohaline circulation, *Geophys. Res. Lett.*, *29*(15), 1728, doi:10.1029/2002GL015229, 2002.
Fichefet, T., S. Hovine, and J. C. Duplessy, A model study of the Atlantic thermohaline circulation during the Last Glacial Maximum, *Nature*, *372*, 252–255, 1994.
Giannini, A., J. C. H. Chiang, M. A. Cane, Y. Kushnir, and R. Seager, The ENSO teleconnection to the tropical Atlantic Ocean: Contributions of the remote and local SSTs to rainfall variability in the tropical Americas, *J. Clim.*, *14*, 4530–4544, 2001.
Harris, S. E., and A. C. Mix, Pleistocene precipitation balance in the Amazon Basin recorded in deep sea sediments, *Quat. Res.*, *51*, 14–26, 1999.
Hastenrath, S., and L. Greischar, Circulation mechanisms related to Northeast Brazil rainfall anomalies, *J. Geophys. Res.*, *98*, 5093–5102, 1993.
Haug, G. H., K. A. Hughen, D. M. Sigman, L. C. Peterson, and U. Rohl, Southward migration of the intertropical convergence zone through the Holocene, *Science*, *293*, 1304–1308, 2001.
Hughen, K. A., J. T. Overpeck, L. C. Peterson, and S. Trumbore, Rapid climate changes in the tropical Atlantic region during the last deglaciation, *Nature*, *380*, 51–54, 1996.
Johnson, H. L., and D. P. Marshall, Localization of abrupt change in the north Atlantic thermohaline circulation, *Geophys. Res. Lett.*, *29*(6), 1083, doi:10.1029/2001GL014140, 2002.
Kiehl, J. T., J. J. Hack, G. B. Bonan, B. A. Boville, B. P. Briegleb, D. L. Williamson, and P. J. Rasch, Description of the NCAR community climate model, *NCAR Tech. Note 152*, Natl. Cent. for Atmos. Res., Boulder, Colo., 1996.
Kiehl, J. T., J. J. Hack, G. B. Bonan, B. A. Boville, D. L. Williamson, and P. J. Rasch, The National Center for Atmospheric Research Community Climate Model: CCM3, *J. Clim.*, *11*, 1131–1149, 1998.
Kitoh, A., and S. Murakami, Tropical Pacific climate at the mid-Holocene and the Last Glacial Maximum simulated by a coupled ocean-atmosphere general circulation model, *Paleoceanography*, *17*(3), 1047, doi:10.1029/2001PA000724, 2002.
Koutavas, A., J. Lynch-Stieglitz, T. M. Marchitto, and J. P. Sachs, El Nino-like pattern in ice age tropical Pacific sea surface temperature, *Science*, *297*, 226–230, 2002.
Kushnir, Y., R. Seager, J. Miller, and J. C. H. Chiang, A simple coupled model of tropical Atlantic decadal climate variability, *Geophys. Res. Lett.*, *29*(23), 2133, doi:10.1029/2002GL015874, 2002.
Latif, M., and T. P. Barnett, Interactions of the tropical oceans, *J. Clim.*, *8*, 952–964, 1995.
Li, T. M., and S. G. H. Philander, On the seasonal cycle of the equatorial Atlantic Ocean, *J. Clim.*, *10*, 813–817, 1997.
Lindzen, R. S., and S. Nigam, On the role of sea-surface temperature gradients in forcing low-level winds and convergence in the tropics, *J. Atmos. Sci.*, *44*, 2440–2458, 1987.
Manabe, S., and A. J. Broccoli, The influence of continental ice sheets on the climate of an ice age, *J. Geophys. Res.*, *90*, 2167–2190, 1985.
Marshall, J., Y. Kushnir, D. Battisti, P. Chang, A. Czaja, R. Dickson, J. Hurrell, M. McCartney, R. Saravanan, and M. Visbeck, North Atlantic climate variability: Phenomena, impacts and mechanisms, *Int. J. Climatol.*, *21*, 1863–1898, 2001.
Mix, A. C., A. E. Morey, N. G. Pisias, and S. W. Hostetler, Foraminiferal faunal estimates of paleotemperature: Circumventing the no-analog problem yields cool ice age tropics, *Paleoceanography*, *14*, 350–359, 1999.
Nobre, P., and J. Shukla, Variations of sea surface temperature, wind stress, and rainfall over the tropical Atlantic and South America, *J. Clim.*, *9*, 2464–2479, 1996.
Okumura, Y., S.-P. Xie, A. Numaguti, and Y. Tanimoto, Tropical Atlantic air-sea interaction and its influence on the NAO, *Geophys. Res. Lett.*, *28*, 1507–1510, 2001.
Peltier, W. R., Ice-age paleotopography, *Science*, *265*, 195–201, 1994.
Peterson, L. C., G. H. Haug, K. A. Hughen, and U. Rohl, Rapid changes in the hydrologic cycle of the tropical Atlantic during the last glacial, *Science*, *290*, 1947–1951, 2000.
Rahmstorf, S., Ocean circulation and climate during the past 120,000 years, *Nature*, *419*, 207–214, 2002.
Saravanan, R., and P. Chang, Oceanic mixed layer feedback and tropical Atlantic variability, *Geophys. Res. Lett.*, *26*, 3629–3632, 1999.
Saravanan, R., and P. Chang, Interactions between tropical Atlantic variability and El Nino-Southern Oscillation, *J. Clim.*, *13*, 2177–2194, 2000.
Seager, R., Y. Kushnir, M. Visbeck, N. Naik, J. Miller, G. Krahmann, and H. Cullen, Causes of Atlantic Ocean climate variability between 1958 and 1998, *J. Clim.*, *13*, 2845–2862, 2000.
Seager, R., Y. Kushnir, P. Chang, N. H. Naik, J. Miller, and W. Hazeleger, Looking for the role of the ocean in tropical Atlantic decadal climate variability, *J. Clim.*, *14*, 638–655, 2001.
Servain, J., I. Wainer, J. P. McCreary Jr., and A. Dessier, Relationship between the equatorial and meridional modes of climatic variability in the tropical Atlantic, *Geophys. Res. Lett.*, *26*, 485–488, 1999.
Sutton, R. T., S. P. Jewson, and D. P. Rowell, The elements of climate variability in the tropical Atlantic region, *J. Clim.*, *13*, 3261–3284, 2000.
Vellinga, M., and R. A. Wood, Global climatic impacts of a collapse of the Atlantic thermohaline circulation, *Clim. Change*, *54*, 251–267, 2002.
Xie, P. P., and P. A. Arkin, Global precipitation: A 17-year monthly analysis based on gauge observations, satellite estimates, and numerical model outputs, *Bull. Am. Meteorol. Soc.*, *78*, 2539–2558, 1997.
Xie, S.-P., A dynamic ocean-atmosphere model of the tropical Atlantic decadal variability, *J. Clim.*, *12*, 64–70, 1999.
Xie, S.-P., and K. Saito, Formation and variability of a northerly ITCZ in a hybrid coupled AGCM: Continental forcing and oceanic-atmospheric feedback, *J. Clim.*, *14*, 1262–1276, 2001.
Xie, S.-P., and Y. Tanimoto, A pan-Atlantic decadal climate oscillation, *Geophys. Res. Lett.*, *25*, 2185–2188, 1998.
Yang, J., A linkage between decadal climate variations in the Labrador Sea and the tropical Atlantic Ocean, *Geophys. Res. Lett.*, *26*, 1023–1026, 1999.
Yin, J. H., and D. S. Battisti, The importance of tropical sea surface temperature patterns in simulations of Last Glacial Maximum climate, *J. Clim.*, *14*, 565–581, 2001.
Zebiak, S. E., Air-sea interaction in the equatorial Atlantic region, *J. Clim.*, *6*, 1567–1568, 1993.

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