Understanding the Mid-Holocene Climate

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ABSTRACT

Paleoclimatic evidence suggests that during the mid-Holocene epoch (about 6000 yr ago) North America and North Africa were significantly drier and wetter, respectively, than at present. Modeling efforts to attribute these differences to changes in orbital parameters and greenhouse gas (GHG) levels have had limited success, especially over North America. In this study, the importance of a possibly cooler tropical Pacific Ocean during the epoch (akin to a permanent La Niña–like perturbation to the present climate) in causing these differences is emphasized. Systematic sets of atmospheric general circulation model experiments, with prescribed sea surface temperatures (SSTs) in the tropical Pacific basin and an interactive mixed layer ocean elsewhere, are performed. Given the inadequacies of current fully coupled climate models in simulating the tropical Pacific climate, this intermediate coupling model configuration is argued to be more suitable for quantifying the contributions of the altered orbital forcing, GHG levels, and tropical Pacific SST conditions to the different mid-Holocene climates. The simulated responses in this configuration are in fact generally more consistent with the available evidence from paleovegetation and sedimentary records.

Coupling to the mixed layer ocean enhances the wind–evaporation–SST feedback over the tropical Atlantic Ocean. The net response to the orbital changes is to shift the North Atlantic intertropical convergence zone (ITCZ) northward, and make North Africa wetter. The response to the reduced GHG levels opposes, but does not eliminate, these changes. The northward-shifted ITCZ also blocks the moisture supply from the Gulf of Mexico into North America. This drying tendency is greatly amplified by the local response to La Niña–like conditions in the tropical Pacific. Consistent with the paleoclimatic evidence, the simulated North American drying is also most pronounced in the growing (spring) season.

1. Introduction

The climate research community is increasingly interested in understanding how aspects of regional climates, especially the hydrological cycle, might change under future climate conditions. One way to build confidence in future climate projections is to assess our ability to simulate past climate changes, for which, unlike the future, at least some paleoclimatic proxy data are available for verification. We focus here on the mid-Holocene epoch (~ 6000 yr ago), when atmospheric CO₂ levels were lower than those at present (Raynaud et al. 1993) and the seasonal cycle was stronger in the Northern Hemisphere because of different orbital parameters (Berger 1978).

The mid-Holocene paleoclimatic records reveal a distinct pattern of drier conditions across central North America and wetter conditions across North Africa (e.g., Jolly et al. 1997; Forman et al. 2001; Viau and Gajewski 2001). These differences from the modern climate were consistently underestimated in the atmospheric general circulation model (AGCM) simulations performed under the Paleoclimate Modeling Intercomparison Project (PMIP), with mid-Holocene orbital parameters, CO₂ levels, and prescribed modern climato-
logical sea surface temperatures (SSTs; Joussaume et al. 1999). The PMIP simulations were reasonably successful in the core monsoon regions, but their failure over North America and North Africa suggests that other factors might also have been in play.

Several possible contributing mechanisms that may not have been accurately represented in the PMIP simulations have been proposed. Forman et al. (2001) speculated from a qualitative analysis of paleoclimatic reconstructions that reduced remote tropical Pacific SST variability associated with a permanent La Niña–like state may have been important in causing the North American drought. Strong regional vegetation–climate feedbacks might also have been important, especially in making North Africa wetter (McAvaney et al. 2001; Levis et al. 2004). The two effects were likely not independent; the regional feedbacks might themselves have been triggered by the remote SST changes.

The reconstructed mid-Holocene pattern of North American dryness/North African wetness is strikingly similar to the leading EOF pattern of the observed modern (1900–95) Palmer Drought Severity Index (PDSI), which is known to be well correlated with the El Niño–Southern Oscillation (ENSO) (e.g., Dai et al. 1998). Moreover, modern-day dryness over central North America has been linked to the cold phase of ENSO in several studies (Palmer and Brankovic 1988; Ting and Wang 1997; Hoerling and Kumar 2003). Indeed, much of the recent success in the seasonal forecasting of subcontinental-scale temperature and precipitation is attributable to improved predictions of the regional response to tropical ENSO-related SST variations. In a coral and tree-ring study of climate variability over the last 300 yr, Cole et al. (2002) found a link between multiyear La Niña events and persistent droughts across central North America. Of course, we do not know that the mean difference between the mid-Holocene and modern tropical Pacific SSTs was precisely La Niña–like. Nonetheless, these studies, and also the recent study of Barsugli and Sardeshmukh (2002) establishing the general sensitivity of North American precipitation to tropical Pacific SSTs, suggest a potentially important role for remote tropical Pacific conditions in causing the relatively dry mid-Holocene climate over North America.

What was the tropical Pacific doing at the mid-Holocene? The reconstructions, though sparse, are suggestive of persistent La Niña–like conditions and/or weaker ENSO variability (Fig. 1a; see also Fig. 10-3 of Gagan and Thompson 2005). By “persistent La Niña–like conditions” we mean tropical Pacific SSTs that were generally colder and warmer than those at present in the eastern and western parts of the basin, respectively, consistent with the occurrence of weaker and less frequent El Niño events. Gagan et al. (1998) produced an annually resolved mid-Holocene record from a geochemical analysis of coral material from the western tropical Pacific (specifically, from the Great Barrier Reef off Australia) that indicated about 1.5°C warmer and more saline conditions there than those at present. Overpeck and Webb (2000) interpreted the Gagan et al. data, along with paleoclimate reconstructions from lake sediments, fossil pollen, fossil marine fauna, and terrestrial sediments at the eastern and western edges of the tropical Pacific, as evidence for weaker ENSO variability and a warmer western tropical Pacific, consistent with mean La Niña–like conditions. The reconstructions of Tudhope et al. (2001) at the western edge of the Pacific warm pool near Papua New Guinea, and of Sandweiss et al. (2001) in coastal Peru, also suggest weaker ENSO variability during the mid-Holocene. The analysis by T. Quinn (2004, personal communication) of West Pacific warm-pool corals suggests little change in the amplitude of ENSO variability, but an increase of at least 1°C SST, which is qualitatively consistent with Tudhope et al.’s warming estimate of 2.3°C. The overall evidence of a warmer western tropical Pacific, combined with that of a cooler eastern tropical Pacific near the Galapagos Islands (Koutavas et al. 2002), strongly suggests that persistent La Niña–like conditions prevailed during the mid-Holocene. These considerations motivate us here to assess the regional impacts of altered tropical Pacific SST conditions around the globe, especially on the North American dryness/North African wetness difference from the modern climate, in a dynamically consistent framework.

Ideally, one should address this issue using climate models in which a dynamically active atmosphere is fully coupled to a dynamically active ocean, and that accurately simulate the altered tropical Pacific SSTs themselves in response to prescribed orbital and CO₂ changes. Such fully coupled model simulations of the mid-Holocene are, however, inconsistent with paleoclimatic records, especially in regions of modern ENSO influence. For example, the current generation of models show either a slight decrease or an increase of precipitation across central North America, and relatively weak increases of precipitation across North Africa that also are too far south (Harrison et al. 2003; Liu et al. 2003, hereafter L03; Braconnot et al. 2004). This model response is inconsistent with the paleoclimatic evidence of widespread drought across central North America and wetness across North Africa (Jolly et al. 1997; Forman et al. 2001; Viau and Gajewski 2001). It is noteworthy that these fully coupled models also have diffi-
difficulty in simulating modern tropical Pacific conditions and their global impacts. Indeed, this is true of current fully coupled models in general; most of the participating models in the Coupled Model Intercomparison Project (CMIP) only generate ENSO-like variability, in a higher and narrower frequency band than is observed, and do not reproduce many of the observed remote ENSO teleconnections (AchutaRao and Sperber 2002). The CMIP models without surface flux adjustments also have mean SST biases of as large as ±3°C in the tropical Pacific (see Fig. 1 of AchutaRao and Sperber 2002). Such deficiencies compromise the usefulness of the fully coupled models for investigating tropical Pacific SST conditions and their global impacts in altered climates, such as the mid-Holocene.

One is thus faced with the dilemma that while accounting for at least some air–sea coupling is apparently essential to resolve the issues at hand, full coupling as in the current models introduces unacceptable errors. Our partial solution is to prescribe hypothesized mid-Holocene SST climatologies in the tropical Pacific and include air–sea coupling to a mixed layer ocean elsewhere. We find that an AGCM with this intermediate ocean coupling, and with prescribed mid-Holocene atmospheric trace gas concentrations and orbital parameters, produces improved simulations of regional mid-Holocene climate differences from the present, especially over North America and North Africa. We confirm that some coupling is indeed essential, but suggest that the coupling employed here is adequate. We also demonstrate that accounting for tropical Pacific conditions is crucial for a realistic simulation of the mid-Holocene North American drying.

2. Models and experimental setup

The AGCM used here is version 3.10 of the Community Climate Model (CCM3.10; Kiehl et al. 1998), developed at the National Center for Atmospheric Research (NCAR). It employs a spatial discretization of T42 in the horizontal (about 2.8° in latitude and longi-
Table 1. Boundary conditions specified for the modern and mid-Holocene simulations. The orbital parameters are from Berger (1978), and the greenhouse gas concentrations are from Raynaud et al. (1993).

<table>
<thead>
<tr>
<th>Boundary conditions</th>
<th>Modern (0-kyr BP)</th>
<th>Mid-Holocene (6-kyr BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Solar constant (W m(^{-2}))</td>
<td>1367</td>
<td>1367</td>
</tr>
<tr>
<td>Orbital parameters</td>
<td>Eccentricity</td>
<td>0.016 724</td>
</tr>
<tr>
<td></td>
<td>Axial tilt</td>
<td>23.446°</td>
</tr>
<tr>
<td></td>
<td>(\omega - 180^\circ)</td>
<td>102.04°</td>
</tr>
<tr>
<td>GHG concentrations (ppm)</td>
<td>CO(_2)</td>
<td>355</td>
</tr>
<tr>
<td></td>
<td>CH(_4)</td>
<td>1.714</td>
</tr>
<tr>
<td></td>
<td>N(_2)O</td>
<td>0.311</td>
</tr>
<tr>
<td></td>
<td>CFC-11</td>
<td>0.280 \times 10^{-3}</td>
</tr>
<tr>
<td></td>
<td>CFC-12</td>
<td>0.503 \times 10^{-3}</td>
</tr>
</tbody>
</table>

tude) and 18 levels in the vertical. Unless stated otherwise, all of our experiments with this AGCM were performed by prescribing SSTs in the tropical Pacific basin (15°S–15°N, and from 120°E to the west coast of the Americas) and coupling elsewhere to an oceanic mixed layer model (MLM; e.g., Yulaeva et al. 2001). To produce a more realistic modern control global climate in this model with prescribed modern climatological-mean tropical Pacific SSTs, a surface heat flux correction was applied to the MLM. This correction was estimated as the additional surface flux necessary in a separate 40-yr run with prescribed modern climatological tropical Pacific SSTs to bring the drifting MLM temperatures elsewhere back to the observed daily climatological SST values every day. It thus crudely accounts for both the CCM surface heat flux parameterization errors over the oceans and the absence of oceanic heat transports in the MLM. With this correction, the MLM temperature at all oceanic grid points was calculated as

\[
\frac{\partial T_m}{\partial t} = \frac{Q + Q_c}{\rho C_p h_m},
\]

where \(T_m\) is the oceanic mixed layer temperature, \(\rho\) is the seawater density (1000 kg m\(^{-3}\)), \(C_p\) is the specific heat capacity of seawater (4000 J kg\(^{-1}\) K\(^{-1}\)), \(Q\) and \(Q_c\) are the net surface heat flux from the CCM and applied heat flux correction, respectively, and \(h_m\) is the prescribed oceanic mixed layer depth. Daily values of the mixed layer depth were interpolated from observed modern monthly climatologies (Levitus 1982).

In the tropical Pacific, the calculated mixed layer temperatures (equivalent to SSTs) were replaced by various hypothesized SST fields (e.g., modern climatology, El Niño, La Niña, La Niña \(\times -1\), and La Niña + 1°C; see Fig. 1b) to investigate the sensitivity of the mid-Holocene climate to them. The climatological-mean, El Niño, and La Niña SST fields were derived from the 50-yr (1951–2000) Global Sea Ice and SST (GISST) dataset (Parker et al. 1995). The experiment with the climatology plus sign-reversed La Niña anomaly field (La Niña \(\times -1\)) was performed to assess the linearity of the tropically forced responses. The experiment with the spatially uniform 1°C warming imposed on the La Niña field (La Niña + 1°C) was performed to assess the impacts of La Niña–like conditions, but with a warmer western tropical Pacific inferred from some paleoclimate records (Gagan et al. 1998; Overpeck and Webb 2000).

For the mid-Holocene simulations, the orbital parameters (eccentricity, obliquity, and longitude of perihelion) were set to the mid-Holocene values as in Berger (1978; see Table 1). Given the uncertainties of the vegetation distribution, we simply prescribed the modern vegetation distribution in all of our runs. The concentrations of the atmospheric trace gases (CO\(_2\), CH\(_4\), and N\(_2\)O) were set to the mid-Holocene values deduced from the Greenland and Antarctic ice cores (Raynaud et al. 1993; Table 1).

The experiments performed with the mid-Holocene forcings are listed in Table 2. In addition to the modern control (Exp0), we performed eight sensitivity experiments with the mid-Holocene forcings to examine the contributions of the different orbital forcings (Exp1 – Exp0), the La Niña–like tropical Pacific SSTs (Exp2 – Exp1), and greenhouse gas (GHG) levels (Exp7 – Exp1 or Exp8 – Exp2) to the total simulated mid-Holocene difference from the present climate (Exp8 – Exp0). Table 2 also lists additional experiments performed to estimate the importance of the remote air–sea coupling (Exp6 – Exp0), to test the linearity and sensitivity of the responses to various hypothesized tropical Pacific SST fields in the mid-Holocene orbital setting (Exp3, –4, –5 – Exp1), and to assess the extent to which the sensitivity of the mid-Holocene climate to a La Niña–like SST perturbation (Exp8 – Exp7) might have differed from that in the modern climate (Exp9 – Exp0). All integrations with the MLM coupling were 50 yr long, of which the last 40 yr were analyzed. These much longer runs than those in previous paleoclimate modeling studies make our conclusions much less sus-
ceptible to sampling errors. The prescribed (i.e., uncoupled) SST run (Exp6) was 20 yr long, of which the last 15 yr were analyzed.

3. Results

The simulated modern area-averaged seasonal cycles of precipitation in the coupled control run (Exp0) over North America (20°–50°N, 120°–70°W) and North Africa (10°–30°N, 20°W–30°E) are shown in Fig. 2. They may be compared with the corresponding observational cycles derived from the 50-yr National Centers for Environmental Prediction (NCEP)–NCAR reanalysis (Kistler et al. 2001) and the Xie–Arkin (Xie and Arkin 1997) precipitation datasets (Fig. 2). [These data are available from the National Oceanic and Atmospheric Administration–Cooperative Institute for Research in Environmental Sciences (NOAA–CIRES) Climate Diagnostics Center (CDC), in Boulder, Colorado, and can be found online at http://www.cdc.noaa.gov.] In view of the known observational uncertainties, the control simulation is reasonably accurate in our target areas of interest.

This section is organized as follows. Section 3a provides further evidence of the adequacy of our intermediate coupling strategy through comparisons of our simulated changes of the monsoons with the available paleoclimatic information and with previous fully coupled simulations. Section 3b demonstrates the importance of coupled air–sea interactions over the tropical Atlantic and La Niña–like tropical Pacific SSTs in causing a wetter North Africa and a drier North America during the mid-Holocene in response to the orbital and GHG changes. Our success in capturing the correct seasonal dependence of these differences from the modern climate is emphasized. We provide compelling evidence in section 3c that the sensitivity of the North American drought to cool eastern tropical Pacific SSTs may have been the same during the mid-Holocene as in the modern climate. The latter is discussed in detail in section 3d, and is shown to be rather robust. The main argument developed in sections 3c

![Fig. 2. The simulated present-day seasonal cycles of the precipitation (mm day⁻¹; black solid lines) over (a) North America (20°–50°N, 120°–70°W) and (b) North Africa (10°–30°N, 20°W–30°E). The climatological precipitation data from 50-yr NCEP–NCAR reanalysis (gray solid lines) and Xie–Arkin (gray dashed lines) datasets are also shown for comparison.](image-url)

<p>| Table 2. List of experiments and boundary conditions. The numbers in parentheses indicate the number of simulated years averaged for the analysis. |</p>
<table>
<thead>
<tr>
<th>Expt</th>
<th>Orbital parameters</th>
<th>GHGs</th>
<th>Tropical Pacific SSTs*</th>
<th>Run length (averaging period)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>Modern control</td>
<td>0-kyr BP</td>
<td>0-kyr BP</td>
<td>Climatology</td>
</tr>
<tr>
<td>1</td>
<td>Mid-Holocene (orbital)</td>
<td>6-kyr BP</td>
<td>0-kyr BP</td>
<td>Climatology</td>
</tr>
<tr>
<td>2</td>
<td>6-kyr BP</td>
<td>0-kyr BP</td>
<td>La Niña</td>
<td>50 (40)</td>
</tr>
<tr>
<td>3</td>
<td>6-kyr BP</td>
<td>0-kyr BP</td>
<td>La Niña × 1°C</td>
<td>50 (40)</td>
</tr>
<tr>
<td>4</td>
<td>6-kyr BP</td>
<td>0-kyr BP</td>
<td>El Niño</td>
<td>50 (40)</td>
</tr>
<tr>
<td>5</td>
<td>6-kyr BP</td>
<td>0-kyr BP</td>
<td>La Niña + 1°C</td>
<td>50 (40)</td>
</tr>
<tr>
<td>6</td>
<td>6-kyr BP</td>
<td>0-kyr BP</td>
<td>Climatology</td>
<td>20 (15)**</td>
</tr>
<tr>
<td>7</td>
<td>Mid-Holocene (orbital plus GHGs)</td>
<td>6-kyr BP</td>
<td>6-kyr BP</td>
<td>Climatology</td>
</tr>
<tr>
<td>8</td>
<td>6-kyr BP</td>
<td>6-kyr BP</td>
<td>La Niña</td>
<td>50 (40)</td>
</tr>
<tr>
<td>9</td>
<td>0-kyr BP</td>
<td>0-kyr BP</td>
<td>La Niña</td>
<td>50 (40)</td>
</tr>
</tbody>
</table>

* Tropical Pacific SSTs were calculated from the composite of a 50-yr SST dataset (Parker et al. 1995; Fig. 1).
** In Exp6, the uncoupled run, the prescribed seasonal cycle of global SST was identical to the 40-yr-mean seasonal cycle simulated in Exp0.
and 3d is that a severe mid-Holocene North American drought could easily have been induced by persistent La Niña-like SST conditions in the tropical Pacific, but, even more importantly, that they did not have to be precisely La Niña-like for the drought to occur.

a. Overview of observed and simulated monsoon precipitation changes

The differences of the simulated mid-Holocene precipitation seasonal cycles (Exp1–5) from the control (Exp0) over the core monsoon regions are shown in Fig. 3, along with a summary of the available paleoclimatic evidence. To facilitate comparisons with the fully coupled study of L03, these regions were chosen to correspond closely to theirs. The main difference is that our North American (NA) box is shifted 10° northward to overlap better with the area of the greatest available paleoclimatic information.

Our simulations evidently capture the major elements of the summer monsoon precipitation changes. In the Northern Hemisphere, the monsoon precipitation is generally increased over North America (Fig. 3b), North Africa (Fig. 3c), and Asia (Fig. 3d) regardless of the imposed tropical Pacific SST conditions. This response is consistent with previous fully coupled simulations (e.g., Figs. 4 and A1 of L03). In L03, the coupling further amplified the land–sea contrast already amplified by an orbitally enhanced seasonal cycle, and hence the monsoon circulation and rainfall. The fact that we are able to reproduce their results without explicitly considering ocean dynamics suggests that thermodynamic air–sea coupling through simple mixed layer physics is sufficient for this purpose.

In the Southern Hemisphere, our simulations show reduced summer monsoon rainfall over South America (Fig. 3e), South Africa (Fig. 3f), and Australia (Fig. 3g). The reductions over South America and South Africa are consistent with those of L03, but the changes over Australia are inconsistent with both L03 and the paleoclimatic record (Johnson et al. 1999; Bowler et al. 2001). L03 suggested the importance of the local ocean dynamic feedback in intensifying the Australian rainfall, which is not included in our mixed layer model simulations.

The paleoclimatic evidence in Fig. 3a suggests widespread drought over North America during the mid-Holocene. Our simulations also show substantially decreased precipitation in fall and winter. For the La Niña SST forcing, the effect lingers through and intensifies in the spring and summer seasons. For this forcing there is a decrease of about 0.13 mm day$^{-1}$, even in the annual-mean precipitation rate.

b. Simulated changes over the Western Hemisphere

1) Annual mean

The experiments listed in Table 2 have been analyzed to isolate the contributions of various forcings and processes on the annual-mean fields of surface temperature, precipitation rate, surface winds, and sea level pressure during the mid-Holocene (Fig. 4). The isolated contributions are

(i) orbital changes without air–sea coupling (Exp6 – Exp0);
(ii) orbital changes with air–sea coupling (Exp1 – Exp0);
(iii) La Niña-like tropical Pacific SST changes (Exp2 – Exp1);
(iv) GHG changes (Exp8 – Exp2); and
(v) total (combined orbital, La Niña-like tropical Pacific SST, and GHG) changes (Exp8 – Exp0).

The coupled response to the orbital changes (Fig. 4b) accounts for a large part of the total response (Fig. 4e). It is not surprising that the coupling alters the near-surface air temperature response over the ocean. Perhaps more surprising is that it also radically alters and amplifies the precipitation response over the tropical Atlantic. The net effect is to shift the Atlantic intertropical convergence zone (ITCZ) northward, make North Africa wetter, and large portions of the southern and eastern United States drier. The main impacts of the La Niña-like tropical Pacific SSTs are in the Pacific–American sector, and are associated with a pronounced warming and drying over North America and a cooling and moistening over tropical South America. These changes in turn result in reduced and increased soil moisture over North America and tropical South America, respectively (figure not shown). The response to the GHG changes generally opposes but does not nullify these responses to the orbital and La Niña SST changes. The GHG radiative forcing$^2$ of $-2.12$ W m$^{-2}$ is somewhat exaggerated in our simulations. A modern CO$_2$ concentration value of 318 ppm (average of twen-

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$^2$ This response initiates the warm and dry conditions over North America (Fig. 4b), although the precipitation response is weak and limited to the south-central United States. Our coupled orbital response qualitatively agrees with previous studies using AGCMs coupled to a slab mixed layer ocean (Vettoretti et al. 1998; Sawada et al. 2004).

$^3$ The radiative forcing resulting from the reduced atmospheric trace gas concentrations was calculated using the equations in Ramaswamy et al. (2001). The contribution of each gas to the total forcing of $-2.12$ W m$^{-2}$ was as follows: CO$_2$ = $-1.30$ W m$^{-2}$, CH$_4$ = $-0.47$ W m$^{-2}$, N$_2$O = $-0.12$ W m$^{-2}$, CFC-11 = $-0.07$ W m$^{-2}$, and CFC-12 = $-0.16$ W m$^{-2}$.
tieth century), instead of the 355 ppm (A.D. 1990) we used, would arguably have been more representative of the twentieth century, and more consistent with its SST climatology. For instance, our modern control run using 355 ppm is approximately 0.5°C warmer over North America than a transient twentieth century CO₂ run using the same model (result not shown).

The total precipitation response (Exp8 – Exp0) in
Fig. 4. Simulated annual-mean changes of (left) surface temperature (°C), (middle) precipitation (mm day⁻¹), and (right) surface winds (m s⁻¹) and sea level pressure (hPa) at the mid-Holocene. (a) The uncoupled response to orbital changes, (b) the coupled response to orbital changes, (c) the coupled response to a hypothesized La Niña-like tropical Pacific SSTs, (d) the coupled response to GHG changes, and (e) the total coupled response to the orbital, La Niña-like SSTs, and GHG changes are shown. Negative values in sea level pressure changes are gray shaded.

This part of the globe is dominated by the northward-shifted Atlantic ITCZ. This is consistent with the previous fully coupled simulations of Kutzbach and Liu (1997), Otto-Bliesner (1999), Braconnot et al. (2004), and L03. Our distinctive contribution is to highlight the sensitivity of this total precipitation response to the tropical Pacific SST forcing. Without this forcing, our simulations would show an increase of precipitation over the United States and a net surface cooling. As in the modern climate, one may reasonably expect that both the drying and warming were important contributors to the mid-Holocene U.S. drought inferred from the paleovegetation and sedimentary records.

We also generated mid-Holocene simulations for other prescribed hypothetical tropical Pacific SST fields to assess the sensitivity of the simulated North American drought to them (Table 2). The substantial annual-mean surface temperature and precipitation responses to the La Niña + 1°C forcing (Exp5 – Exp1), the negative La Niña forcing (Exp3 – Exp1), and the El Niño forcing (Exp4 – Exp1), shown in Fig. 5, again attest to the sensitivity of the regional mid-Holocene climates to tropical Pacific SSTs. Note that all of these responses are generally opposite to the response to the La Niña forcing shown in Fig. 4. The opposite responses to the negative La Niña and El Niño forcings are not surprising, and provide reassuring evidence of the approximate linearity of these forcing–response relationships.

The response to the La Niña + 1°C forcing is more troubling. Like the response to the La Niña forcing in Fig. 4, it also shows a warming over North America, consistent with the drought, but an opposite precipita-
tion response, inconsistent with the drought. One way to reconcile this inconsistency is to recall that our hypothesized basinwide La Niña + 1°C SST pattern was inferred from Gagan et al’s (1998) West Pacific proxy warming data, and to allow for the possibility that those local data may not be indicative of basin-wide conditions at the mid-Holocene. Koutavas et al. (2002) have since published results indicative of about a −2°C cooling in the eastern tropical Pacific near the Galapagos Islands, further supporting the notion that the basin-wide SST field may have been La Niña–like. One might also expect a basin-wide La Niña + 1°C warming to be inconsistent with a “global cooling” scenario associated with reduced CO₂ levels during the mid-Holocene.

2) Seasonal Cycle

The orbital, tropical Pacific SST, and GHG changes have markedly different impacts on the simulated seasonal cycles of surface temperature and precipitation over North America and North Africa (Fig. 6). Some of these differences were masked in the annual averages discussed in the previous section, but are important for fully understanding the proxy paleovegetation and sedimentary records. For instance, reduced North American paleovegetation evidence may more directly reflect a drought in the growing spring season than in midsummer.

The primary orbitally induced seasonal change arises from increased summertime (June, July, and August) insolation, which enhances the summertime land–sea contrast, and thence the monsoonal circulation and rainfall. The effect is clear in Fig. 6 in the summertime surface temperature and rainfall over both North America and North Africa. The response to the GHG changes is more uniform across the seasons, and is generally opposite in summer. Indeed, the summertime magnitude of the North African rainfall is just as dramatically reduced by the GHGs as it is increased by the orbital forcing. As noted previously, these rainfall changes are closely tied to the shifts of the North Atlantic ITCZ, which are strongly amplified by the local coupled wind–evaporation–SST (WES) feedback (which was absent in the uncoupled PMIP simulations; see Fig. 4a). The orbital forcing and GHG changes efficiently excite this “coupled mode” of
Atlantic ITCZ variability, but with opposite signs (see Figs. 4b and 4d).

The La Niña forcing has a minor effect on North Africa. Over North America, however, it has a warming and drying effect throughout the year, which is strongest in spring [March–May (MAM)]. This is an unambiguous drought-inducing response and is largest in the “correct” growing spring season. The orbital forcing also induces drying, though arguably in the “wrong” (winter) season. It is possible that the orbital and La Niña effects acted in concert to produce a year-round drought during the mid-Holocene.

c. Linearity of responses

One potential difficulty in quantifying the contributions of the orbital, tropical Pacific SST, and GHG changes to the mid-Holocene climate concerns the linearity of the responses to these forcings. If the responses were substantially nonlinear and thus large enough to interact with one another, then our estimated contributions of these component forcings (orbital, La Niña, and GHG) would depend, for instance, on the order in which we added these forcings. Substantial nonlinearity would also introduce another ambiguity in assessing the true impact of a component forcing—is it the response to that component in isolation, or is it the difference between the responses to the total forcing with and without that component? One way to settle these ambiguities would be to conduct additional experiments with all possible permutations and combinations of our component forcings. A less comprehensive but simpler way to demonstrate linearity would be to show that equal and opposite responses are obtained for equal and opposite forcings. The generally good correspondence between the opposite responses to the La Niña forcing in Fig. 4 and the negative La Niña forcing in Fig. 5 is reassuring in this regard.

Important additional evidence that the responses to our different forcings interact negligibly with each other is provided in Fig. 7. The figure shows the coupled annual-mean response to the La Niña forcing in isolation, that is, for modern orbital parameters and GHG levels (Fig. 7a; Exp9 – Exp0), for mid-Holocene orbital parameters and modern GHG levels (Fig. 7b; Exp2 – Exp1, repeated from Fig. 4), and for mid-Holocene orbital parameters and GHG levels (Fig. 7c; Exp8 – Exp7). It is remarkable how the precipitation and surface temperature responses to La Niña in these
three experiments with one, two, and all three perturbation forcings included are identical within the sampling error. This is basically because the responses to the orbital and GHG changes, although substantial, are apparently not large enough to significantly alter the background state on which the planetary Rossby wave response to tropical SST forcing evolves.

d. General sensitivity of North American drought to tropical SSTs

Our simulations of the mid-Holocene North American drought are clearly sensitive to the SSTs prescribed in the tropical Pacific. Given the sparseness and ambiguities of the paleoclimatic record in the tropical Pacific (see Fig. 1a), it is important to assess the robustness of this sensitivity. Some indication of this was provided in Fig. 5. The fact that this sensitivity was also shown in Fig. 7 to be practically identical to that in the present climate provides further reassurance in this regard. Figure 7 strongly suggests that the mid-Holocene climate change induced by the orbital and GHG changes was not large enough to alter the sensitivity of the remote atmospheric responses to tropical SST forcing. This fact leads us to make a more general assessment of the sensitivity of the mid-Holocene North American drought to tropical SSTs using results from a just-completed study of global atmospheric sensitivity in the present climate to prescribed tropical SST anomalies in the same NCAR CCM3.10 (P. D. Sardeshmukh et al. 2006, unpublished manuscript). Briefly, the AGCM responses were determined for an array of 43 localized SST anomaly patches over the Indian, Pacific, and Atlantic Oceans (see appendix A and Fig. A1). Ensembles of 16 or more 2-yr integrations were made for each patch, with both positive and negative SST anomalies. The ensemble-mean responses were then combined using a thin-plate smoothing-spline procedure (described in Barsugli and Sardeshmukh 2002) to produce sensitivity maps for target quantities of interest. The target quantity of interest in this study is the precipitation response in spring (MAM) averaged over the NA box (see Fig. 3a). The value plotted in Fig. 8 at each tropical location is the NA precipitation response
to a $1^\circ$C SST cooling over a $10^5$ km$^2$ area centered at that location. A sensitivity of $-25 \, [10^{-3} \, \text{mm day}^{-1} \, (10^6 \, \text{km}^2 \, \text{K})^{-1}]$ in these units means that a $1^\circ$C SST cooling averaged over a $10^6 \times 10^6$ area would force a precipitation deficit of 0.025 mm day$^{-1}$ over the NA box. For the same cooling over a larger $100^\circ \times 100^\circ$ area (about the size of the Niño-3.4 region), the deficit would be 0.25 mm day$^{-1}$, which is comparable to that observed in modern ENSO events. Figure 8 demonstrates a robust sensitivity of the NA drought to cooling in almost all portions of the tropical Pacific and Indian Oceans. Any SST anomaly (or climate change) pattern in the Indo-Pacific with a positive projection on the pattern in Fig. 8 will induce a North American drought. A La Niña–like SST change is an efficient, but certainly not the only, way through which this can be accomplished.

The results shown in Fig. 8 are reassuring, but one might still question our choice of the precipitation deficit averaged over a rather large NA box as an appropriate measure of regional drought, because it ignores the surface warmth associated with the drought and also the possibility of large climate signals of opposite sign within the box. Furthermore, North American droughts do not occur in isolation, but in conjunction with significant anomalies in other parts of the Northern Hemisphere. To the extent that they are regional manifestations of hemispheric anomaly (or climate change) patterns, it is more appropriate to establish the sensitivity of the hemispheric anomaly patterns to tropical SST forcing, instead of just their NA portion as in Fig. 8.

One way to address these issues is to examine the dominant combined EOFs of the 43 ensemble-mean springtime (MAM) response fields of land surface precipitation, land surface temperature, and 500-hPa height over the Northern Hemisphere obtained for the 43 SST anomaly patches (Fig. 9). The first two EOFs together account for about 47% of the response structure in these 43 forcing experiments. Maps of the associated principal component (PC) series$^3$ for these EOFs (Figs. 9c and 9f) depict the amplitude with which the EOF is excited by a $+1^\circ$C SST anomaly patch at each tropical location. For example, a value of $-2.5$ near the equatorial date line in Fig. 9c means that the response to a $1^\circ$C cold patch there would project onto EOF-1 with a magnitude of 2.5. These sensitivity maps may also be interpreted as optimal SST anomaly patterns for forcing the corresponding EOF patterns with the largest magnitude.

For our purposes here, the most important message of Fig. 9 is that springtime North American drought (precipitation deficit accompanied by surface warmth) is a robust feature of the dominant Northern Hemispheric climate sensitivity to tropical SST change. The drought is efficiently excited by La Niña–like SST changes, but it can also be excited by any other SST change pattern with positive projections on the sensitivity maps in Figs. 9c and 9f. One can imagine how this condition could be readily met for a wide range of hypothesized tropical Pacific cooling patterns that are consistent with the sparse paleoclimatic tropical Pacific data.

4. Discussion and conclusions

Our mid-Holocene simulations generated using the NCAR CCM3.10 coupled to an MLM outside the tropical Pacific, with prescribed modern La Niña–like climatological SST boundary conditions in the tropical Pacific, are generally more consistent with the available paleoclimatic information than previous simulations. The AGCMs participating in PMIP ignored some important air–sea coupling effects, such as WES feedbacks in the tropical Atlantic sector that were included.

$^3$ Note that we are showing here an EOF analysis of an “experiment series” of response maps, with each experiment associated with an SST anomaly patch at a specific tropical location, as opposed to the more traditional EOF analysis of a “time series” of maps.
in our simulations. The fully coupled climate model simulations, on the other hand, were adversely affected by their climate biases in the tropical Pacific sector. These biases were absent in our simulations by design.

Our simulations suggest that the following mechanisms led to a wetter North Africa and a drier North America at the mid-Holocene (see Fig. 10). The orbital change was associated with a stronger seasonal cycle in the Northern Hemisphere and a weaker cycle in the Southern Hemisphere. This increased the cross-equatorial SST gradient in the tropical Atlantic and the cross-equatorial wind, which further amplified the SST gradient through a positive WES feedback. The net effect was to shift the tropical Atlantic ITCZ northward. This shift made neighboring North Africa wetter by increasing the flux of moist maritime air into inland

![Diagram of EOF analysis](image-url)

**Fig. 9.** (a), (b) The first combined EOF of the 500-hPa height (m), surface temperature (°C), and precipitation (mm day⁻¹) responses in spring (MAM), and (c) its sensitivity to SST forcing in different tropical locations. Negative 500-hPa height responses are dashed. (d), (e) The second combined EOF and (f) its sensitivity to SST forcing in different tropical locations.
areas (see Fig. 4). The northward-shifted ITCZ was also associated with a greater moisture influx from higher subtropical latitudes and a weaker subtropical high. The net effect was to block the moisture supply from the Gulf of Mexico into North America and contribute to the North American drought. This blockage of Gulf of Mexico moisture, however, was not the only, or even the major, reason for the North American drought. Our simulations suggest that a severe North American drought would not have occurred without the presence also of persistent La Niña–like cold SSTs in the tropical Pacific, which forced an upper-tropospheric planetary Rossby wave train over the Pacific–North American (PNA) region with a dry, warm surface footprint over the United States. The combination of La Niña and reduced moisture supply from the Gulf of Mexico likely led to the severe North American drought. It is plausible that the reduced soil moisture and surface evaporation associated with the warm, dry surface conditions further amplified the drought through a positive soil moisture feedback (e.g., Oglesby and Erickson 1989), although we did not explicitly demonstrate this.

According to our simulations, the reduced mid-Holocene GHG concentrations (e.g., 280 ppm of CO₂) would have damped the changes to a warmer and drier environment over North America. We recognize that our prescribed radiative forcing change of about −2.12 W m⁻² from the modern control was somewhat exaggerated, and therefore the simulated damping effect was also likely overestimated. For example, our modern control run using CO₂ forcing for A.D. 1990 (Exp0) is approximately 0.5°C warmer over North America than a twentieth-century run of the same model with transient CO₂ forcing.

The sensitivity of our simulated North American drought to tropical Pacific cooling is striking. We provided considerable evidence that this sensitivity is robust, that is, it is not dependent upon the SSTs there being precisely La Niña–like. As discussed earlier, some limited paleoclimate records do suggest mean SST conditions during the mid-Holocene that are similar to those during modern ENSO episodes. Most of the coral data suggest a mean shift to warmer and more saline conditions in the western tropical Pacific, with some records suggesting near-modern ENSO variability, and others suggesting greatly diminished ENSO variability at the mid-Holocene. Presumably, these differences among paleorecords partly reflect local effects on temperature and salinity monitored at different coral sites. However, at least some portion of these differences is likely also indicative of the impact of a mean shift in tropical Pacific SSTs on local SST variability. At a minimum, our mid-Holocene simulations with persistent SST states, resembling those encountered during modern ENSO phases, represent interesting sensitivity tests. More importantly, we believe that these experiments provide a reasonable representation of the mid-Holocene climate with near-modern ENSO variability, but with mean shifts of the eastern tropical Pacific SSTs to generally cooler values than those at present.

Our emphasis on the role of a cooler tropical Pacific in inducing the North American drought in no way denies the possible importance of regional vegetation–climate feedbacks in further amplifying the drought. To assess these feedbacks, we conducted two additional experiments with two different prescriptions of mid-Holocene vegetation cover (see appendix B). The results (summarized in Fig. B1) indicate that over North America, the vegetation changes may indeed have amplified the La Niña–induced changes. Over North Africa, the results suggest that the vegetation feedbacks amplified the response to orbital changes (and to a lesser extent, the La Niña–induced changes). Thus, veg-
etation–climate feedbacks were important in amplifying externally induced regional climate changes at the mid-Holocene, where the “external” induction was primarily by the cool tropical Pacific SSTs over North America and by the orbital changes over North Africa.

Finally, we should stress that our specified tropical Pacific SST conditions ignored the SST variability and its likely difference during the mid-Holocene from that at the present. A better accounting of the tropical Pacific SST variability, especially on decadal scales, may throw additional light on the altered likelihood of extreme climate anomalies during the mid-Holocene, such as extended North American droughts. This is a topic of ongoing research.

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APPENDIX A

Anomalous SST Patch Experiments

The experiments were performed with an AGCM, the NCAR CCM3.10, with prescribed idealized anomalous tropical SST patches in the present-day climate setting (Table 1). For each localized patch, the anomalous SST $T_k$ $(k = 1, \ldots, 43)$ was defined as

$$T_k(\lambda, \phi) = B \cos^2\left(\frac{\pi}{2} \left(\frac{\phi - \phi_k}{\phi_k}\right)\right) \cos^2\left(\frac{\pi}{2} \left(\frac{\lambda - \lambda_k}{\lambda_k}\right)\right)$$

(A1)

where $\lambda$ is longitude, $\phi$ is latitude, $(\lambda_k, \phi_k)$ is the center point of the $k$th patch $T_k$, $(\lambda_w, \phi_w)$ is the half-width of the patch, and $B$ is the maximum amplitude of the SST anomaly ($\pm 2^\circC$ in this study). We specified patch half-widths of $10^\circ$ ($10^\circ$) latitude and $44^\circ$ ($28^\circ$) longitude in the tropical Indo-Pacific (Atlantic) Ocean. The locations and shape of the SST patches are shown in Fig. A1. These SST anomaly patches are then added to the observed present-day climatological-mean seasonally varying SSTs to set the lower boundary conditions in CCM3.10.

Each patch experiment consisted of a 16 (20)-member ensemble of 2-yr runs with both warm and cold SST anomalies in the tropical Indo-Pacific (Atlantic) Ocean. The linear response to each SST anomaly patch was estimated as half the difference between the ensemble-mean responses in the warm and cold runs. The atmospheric noise was estimated from a separate 100-yr control climate run with climatological-mean seasonally varying SSTs.

APPENDIX B

Vegetation Sensitivity Experiments

To assess the vegetation effect on the regional climate changes, we performed auxiliary experiments

| Table B1. List of experiments and boundary conditions for the vegetation sensitivity experiments. The numbers in parentheses indicate the number of simulated years averaged for the analysis. |
|----------------|---------------|----------------|----------------|
| **Expt** | **Orbital parameters** | **GHGs** | **Tropical Pacific SSTs** | **Vegetation dataset** | **Run length (averaging period)** |
| 1v | 6-kyr BP | 6-kyr BP | La Niña | Diffenbaugh and Sloan (2002) | 50 (40) |
| 2v | 6-kyr BP | 6-kyr BP | La Niña | Lynch et al. (2003) | 50 (40) |

* Tropical Pacific SSTs were calculated from the composite of a 50-yr SST dataset (Parker et al. 1995; Fig. 1).
with two existing mid-Holocene vegetation reconstructions (Diffenbaugh and Sloan 2002; Lynch et al. 2003). The details of the experimental design are summarized in Table B1. We isolated the effect of the vegetation changes by subtracting the simulated response in Exp8 [with a modern vegetation distribution (Table 1)] from the responses obtained in Exp1v and 2v (Table B1). For ease of quantitative comparisons with other mid-Holocene forcings, we have plotted the vegetation effect on the already shown Fig. 6 (Fig. B1). The region between the temperature and precipitation responses obtained for the two vegetation reconstructions is shaded gray to indicate the uncertainty in these effects.

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