Sensitivity of Last Glacial Maximum climate to uncertainties in tropical and subtropical ocean temperatures

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Abstract

The faunal and floral gradients that underlie the CLIMAP (1981) sea-surface temperature (SST) reconstructions for the Last Glacial Maximum (LGM) reflect ocean temperature gradients and frontal positions. The transfer functions used to reconstruct SSTs from biologic gradients are biased, however, because at the warmest sites they display inherently low sensitivity in translating fauna to SST and they underestimate SST within the euphotic zones where the pycnocline is strong. Here we assemble available data and apply a statistical approach to adjust for hypothetical biases in the faunal–based SST estimates of LGM temperature. The largest bias adjustments are distributed in the tropics (to address low sensitivity) and subtropics (to address underestimation in the euphotic zones). The resulting SSTs are generally in better agreement than CLIMAP with recent geochemical estimates of glacial–interglacial temperature changes. We conducted a series of model experiments using the GENESIS general atmospheric circulation model to assess the sensitivity of the climate system to our bias-adjusted SSTs. Globally, the new SST field results in a modeled LGM surface-air cooling relative to present of 6.4 cooling 1°C cooler than that of CLIMAP). Relative to the simulation with CLIMAP SSTs, modeled precipitation over the oceans is reduced by 0.4 mm d\textsuperscript{-1} (an anomaly 0.4 versus 0.0 mm d\textsuperscript{-1} for CLIMAP) and increased over land (an anomaly 0.2 versus 0.5 mm d\textsuperscript{-1} for CLIMAP). Regionally strong responses are induced by changes in SST gradients. Data-model comparisons indicate improvement in agreement relative to CLIMAP, but differences among terrestrial data inferences and simulated moisture and temperature remain. Our SSTs result in positive mass balance over the northern hemisphere ice sheets (primarily through reduced summer ablation), supporting the hypothesis that tropical and subtropical ocean temperatures may have played a role in triggering glacial changes at higher latitudes.

1. Introduction

One of the primary goals of the Climate Long Range Mapping and Prediction (CLIMAP) program was to reconstruct a snapshot of the ice–age winter (February) and summer (August) sea-surface temperatures (SSTs) and sea ice conditions. Initial SST estimates (CLIMAP, 1976) were used as boundary conditions in an atmospheric general circulation model (AGCM) that was run to provide insights into the nature of the climate during the Last Glacial Maximum (LGM) and to identify possible mechanisms associated with LGM changes in the state of the Earth System (Gates, 1976). An updated and more complete data set (CLIMAP, 1981) has been used extensively for nearly 25 years both as a boundary condition and a target for atmospheric and ocean model experiments.

The largest changes in glacial SSTs estimated by CLIMAP (1976, 1981) are found in the mid-to-high latitudes where, for example, the subpolar North Atlantic is up to 12°C colder than present. Moderate cooling on the order 1–2°C is estimated by CLIMAP in the western Pacific and the Pacific and Atlantic warm pools, whereas cooling in the eastern equatorial oceans is on the order of 2–3°C. One of the more interesting features of the CLIMAP reconstruction is found in the subtropical gyres in the Pacific Basin where LGM SSTs exceed those of present.
CLIMAP SST reconstructions were derived from the “transfer function” methodology first developed by Imbrie and Kipp (1971) from analysis of the population of species of various fossil groups including primarily foraminifera, radiolarians, coccoliths and diatoms. Two critical assumptions of the transfer function method are (1) that the spatial distribution of a species is, to a first order, systematically related to (but not necessarily directly determined by) temperature, and (2) that a statistical relationship based on analyzing the abundances of a multivariate set of species can be inverted to yield temperature estimates that are valid for the LGM. Thus, the underlying statistical basis for the transfer function methodology implicitly requires the assumption that the ecological response of species has not changed through time.

In addition to transfer function techniques, new methodologies for estimating past SSTs have emerged, including statistical approaches such as modern analog matching methods (e.g., Pflaumann et al., 1996; Ortiz and Mix, 1997; Waelbroeck et al., 1998, Trend-Staïd and Prell, 2002) and geochemical methods based on alkenones (e.g., Müller et al., 1998; Prahl et al., 1988) and Mg/Ca ratios recorded in indicator species (e.g., Lea, 1999; Klinkhammer et al., 2004). Alkenone methods have been applied to estimate SST changes in high-productivity regions, whereas Mg/Ca ratios have been used to reconstruct temperatures mainly in low-latitude regions. Comparison of LGM SST anomalies (relative to modern core-top estimates) based on alkenone and Mg/Ca analyses with the changes estimated by faunal transfer functions (Mix et al., 1999) show that, within calibration errors of the methods, there is little or no systematic difference among the methods in some areas; although each of the proxies contain sources of bias (e.g., Brown and Elderfield, 1996; Dekens et al., 2002; Lea et al., 2005; Prahl et al., 2006), and some regional differences continue to stir important debate about the processes of climate change (e.g., Mix, 2003).

One such region of debate is in the tropics and subtropics where an apparent LGM warm-temperature bias based on faunal reconstruction methods of CLIMAP calls into question the applicability and interpretation of the transfer function methodology (Lea et al., 2000). Mix et al. (1999) addressed the apparent lack of cooling in the upwelling regions of the eastern equatorial Pacific and tropical Atlantic gyres by applying modified foraminifera-based transfer functions (e.g., Imbrie and Kipp, 1971) to address

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Fig. 1. Modern distribution of major faunal and floral fossil groups from surface sediments (Polar—dark blue; subpolar—light blue; transitional—green; subtropical—yellow; and warm pool—tan; from Moore et al. (1981). Superimposed are annual SST estimates from the World Ocean Atlas (1998).
the “no-analog” problem, i.e., faunal assemblage in LGM samples unlike any assemblages found in the modern ocean. Mix et al. (1999) developed a strategy to address the potential no-analog problem by defining faunal assemblages using modern spatial variations and faunal data from past-ocean conditions. Their solution proved to be statistically stable, and yielded estimates of greater LGM cooling in the tropics, in agreement with estimates based on CLIMAP-style radiolarian transfer functions (Pisias and Mix, 1997), but maintain the apparent thermal stability of the subtropical gyres. A number of more recent microfossil studies have added further support for greater ice age cooling in the tropical Pacific based on foraminifera (Feldberg and Mix, 2002; Martinez et al., 2003; Morey et al., 2005) and in the tropical Atlantic based on faunal analyses (Niebler et al., 2003; Pfau mann et al., 2003; Kucera et al., 2005; Morey et al., 2005).

Substantial modeling research has also been undertaken to reconcile the apparent discrepancies between oceanic and terrestrial LGM cooling in the tropics, particularly in the Pacific Ocean where SST reconstructions are relatively sparse (e.g., Rind and Peteet, 1985; Pollard and Thompson, 1997; Pinot et al., 1999, Crowley, 2000). A key component of most of the model-based research has been assessing the sensitivity and response of the climate system to SST change prescribed either throughout the tropics and low latitudes or regionally (e.g., Hostetler and Mix, 1999; Zhao et al., 2004) and application of mixed-layer climate models in which SSTs are computed (e.g., Dong and Valdez, 1998).

In the case of prescribed SSTs, most studies derived alternative LGM SSTs from CLIMAP by applying some arbitrarily specified degree of cooling (e.g., Yin and Battisti, 2001; Rodgers et al., 2003, 2004). While those experiments are useful for quantifying the range and nature of climate sensitivity to changes in tropical SSTs, the oceanic temperature gradients in such experiments are not realistic.

Here we consider the possible source of error in faunal-based SST reconstructions that could lead to a warm bias in the tropics and subtropics. We present a modified LGM SST field based on CLIMAP and more recent faunal
temperature estimates (Mix et al., 1999) that we derived by applying plausible bias adjustments. The temperature field displays greater LGM cooling in the low-latitude regions while preserving realistic oceanic temperature gradients. We assess the potential sensitivity of global climate to the bias-adjusted SST field by analyzing AGCM results from experiments in which both the CLIMAP and bias-adjusted SSTs were specified as boundary conditions.

2. Strategy

2.1. Tropical SST biases

The spatial pattern of modern faunal and floral assemblages corresponds well with the latitudinal distribution of SST (Fig. 1). Marked changes in these faunal and floral distributions during the LGM likely reflect changes in both the spatial distribution and gradients of glacial SSTs. The location of the implied, large-scale thermal gradients defined by the LGM faunal and floral data are probably correct, even though the quantitative transfer functions used to reconstruct SSTs may have inherent biases. We suggest that there are two fundamental biases in the faunal-based reconstructions: (1) underestimation of very warm temperatures, particularly in the tropics and (2) failure to account for vertical temperature gradients in the upper water column (i.e., the strength of the pycnocline). These biases were discussed by Mix et al. (1999); our Fig. 2, who demonstrate a relationship between the residuals of surface temperature estimates and density gradients within the euphotic zone.

2.2. Bias adjustments

We start from the global SST reconstruction of Mix et al. (1999), which includes improved reconstructions of SSTs in the eastern tropical Pacific and tropical Atlantic Oceans, and is elsewhere identical to CLIMAP (1981). Our goal was to adjust for the apparent warm temperatures in this baseline LGM reconstruction without substantially modifying the location of the thermal gradients as defined by the underlying faunal/floral data sets; the location of these regional gradients is a key feature of the climate system (Yin and Battisti, 2001). We adjusted the gridded 2° latitude by 2° longitude LGM SST field based on both the warm-temperature and pycnocline biases identified by Mix et al. (1999). At each grid point, the CLIMAP temperature was adjusted as

\[
SST_{\text{adj}} = \left[ 1 + Ae^{-k(SST_{\text{CLIMAP}} - 30)} \right] SST_{\text{CLIMAP}} + B \sigma_r(0-100\text{ m}) + C,
\]

where \( SST_{\text{adj}} \) is the adjusted temperature, \( A, B, C, \) and \( k \) are empirical coefficients, \( SST_{\text{CLIMAP}} \) is the SST of Mix et al. (1999), and \( \sigma_r \) is the pycnocline gradient represented by the water temperature difference between the surface and \(-100\text{ m}\). The first term on the right-hand side of Eq. (1) determines the magnitude of the adjustment for the warm-temperature bias, the second term determines the magnitude of the pycnocline adjustment, and the constant \( C \) shifts the mean of the entire SST field. (These coefficients are the slope \( B \) and intercept \( C \) of the fitted lines plotted in Fig. 2d.) The exponential form of the first term is selected to weight warm bias adjustments to be greatest for higher temperatures in Equatorial region. The value of 30.0 in the exponent of the

Fig. 4. Gradient in the modern ocean pycnocline (\( \sigma_r \) at 0–100 m) from the World Ocean Atlas (1998). Units in \( \sigma_r = (\text{density water} - 1) \ast 1000. \)
Fig. 5. (a) SST anomaly map (Warm Pool–CLIMAP) for August. The Warm Pool SST field is 1.0 °C cooler than CLIMAP in August. (b) SST anomaly map (Warm Pool–CLIMAP) for February. The Warm Pool SST field is 1.1 °C cooler than CLIMAP in February.
Table 1
Comparison between CLIMAP (as modified by Mix et al., 1999) and independent estimates of LGM to Modern changes in SST

<table>
<thead>
<tr>
<th>Core</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Method</th>
<th>Estimated LGM Δt</th>
<th>CLIMAP Feb. Δt</th>
<th>Aug. Δt</th>
<th>Mean Δt</th>
<th>Ref.</th>
</tr>
</thead>
<tbody>
<tr>
<td>TN057-21-PC2</td>
<td>41.13S</td>
<td>007.82E</td>
<td>U$_{57}^S$</td>
<td>−5.0</td>
<td>−2.2</td>
<td>−2.8</td>
<td>−2.7</td>
<td>1</td>
</tr>
<tr>
<td>W8709A-8PC</td>
<td>42.54N</td>
<td>127.68W</td>
<td>U$_{57}^S$</td>
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<td>−3.3</td>
<td>−3.5</td>
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<td></td>
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<tr>
<td>ODP8906B</td>
<td>0.32N</td>
<td>159.37E</td>
<td>Mg/Ca</td>
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<td>−0.7</td>
<td>−3.0</td>
<td>−2.0</td>
<td>3</td>
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<tr>
<td>MD9821-81</td>
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<td>125.83E</td>
<td>Mg/Ca</td>
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<td>−0.2</td>
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<td>−2.0</td>
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<td>TR163-20B</td>
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<td>−2.4</td>
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<td>TR163-22</td>
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<td>TR163-18</td>
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<td>089.85W</td>
<td>Mg/Ca</td>
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<td>−2.4</td>
<td>−1.9</td>
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<tr>
<td>TR163-19</td>
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<td>090.95W</td>
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<tr>
<td>TR163-30</td>
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<td>089.68W</td>
<td>Mg/Ca</td>
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<td>−3.1</td>
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</tr>
<tr>
<td>ODP1242</td>
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<td>Mg/Ca</td>
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<td>ODP1233</td>
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<tr>
<td>MD9821-62</td>
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<td>−0.8</td>
<td>−1.1</td>
<td>−1.5</td>
<td>8</td>
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</table>

Mean CLIMAP is calculated based on the annual cycle used in GCM experiments. Mean difference between CLIMAP and geochemical methods is −1.1 °C. References: 1—Sachs et al. (2001); 2—Prahl et al. (1995); 3—Lea et al. (2000); 4—Stott et al. (2002); 5—Koutavas et al. (2002); 6—Benway (2005); 7—Lamy et al. (2004); 8—Visser et al. (2003).

Table 2
Comparison between independent SST estimates and LGM boundaries from this study

<table>
<thead>
<tr>
<th>Core</th>
<th>Method</th>
<th>Estimated LGM Δt</th>
<th>Warm Pool (EXP2), Mean Δt</th>
<th>Difference</th>
<th>Combined (E2PYC)</th>
<th>Diff</th>
<th>Ref.</th>
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</thead>
<tbody>
<tr>
<td>TN057-12-PC2</td>
<td>U$_{57}^S$</td>
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<tr>
<td>W8709A-8PC</td>
<td>U$_{57}^S$</td>
<td>−4.0</td>
<td>−3.7</td>
<td>0.3</td>
<td>−4.1</td>
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<tr>
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<td>−4.5</td>
<td>−1.5</td>
<td>−4.5</td>
<td>1.5</td>
<td>3</td>
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<td>MD9821-81</td>
<td>Mg/Ca</td>
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<td>−3.0</td>
<td>0.0</td>
<td>−3.1</td>
<td>−0.1</td>
<td>4</td>
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<tr>
<td>TR163-20B</td>
<td>Mg/Ca</td>
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<td>−3.2</td>
<td>−0.3</td>
<td>−2.7</td>
<td>0.2</td>
<td>3</td>
</tr>
<tr>
<td>TR163-22</td>
<td>Mg/Ca</td>
<td>−2.4</td>
<td>−3.2</td>
<td>−0.8</td>
<td>−2.7</td>
<td>−0.3</td>
<td>3</td>
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<td>TR163-18</td>
<td>Mg/Ca</td>
<td>−3.1</td>
<td>−3.2</td>
<td>−0.1</td>
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<td>3</td>
</tr>
<tr>
<td>TR163-19</td>
<td>Mg/Ca</td>
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<td>−3.2</td>
<td>−0.4</td>
<td>−2.4</td>
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<td>3</td>
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<tr>
<td>V21-30</td>
<td>Mg/Ca</td>
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<td>−4.0</td>
<td>−2.8</td>
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<tr>
<td>ODP1242</td>
<td>Mg/Ca</td>
<td>−4.0</td>
<td>−3.6</td>
<td>0.4</td>
<td>−2.4</td>
<td>1.6</td>
<td>6</td>
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<tr>
<td>ODP1233</td>
<td>U$_{57}^S$</td>
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<td>2.1</td>
<td>−4.0</td>
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<td>MD9821-62</td>
<td>Mg/Ca</td>
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<td>−0.2</td>
<td>−3.4</td>
<td>0.1</td>
<td>8</td>
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</table>

Average of absolute value of differences 0.91 0.88

References are listed in Table 1.
Average of absolute values of differences without core V21-30 are 0.75 and 0.71 for the EXP2 and E2PYC experiments, respectively. Combined (E2PYC): $a = −0.1, b = +0.4$ $C = −1$.

Table 3
Annual average 2-m air temperatures (°C) as simulated in the model experiments

<table>
<thead>
<tr>
<th></th>
<th>Global</th>
<th>Global land</th>
<th>Global ocean</th>
<th>Tropical land</th>
<th>Tropical ocean</th>
<th>NH land</th>
<th>NH ocean</th>
<th>NH ice</th>
<th>SH land</th>
<th>SH ocean</th>
<th>SH ice</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>13.7</td>
<td>13.0</td>
<td>16.1</td>
<td>23.2</td>
<td>24.6</td>
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<td>18.2</td>
<td>−18.5</td>
<td>21.2</td>
<td>3.9</td>
<td>−2.6</td>
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<tr>
<td>CLIMAP</td>
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<td>10.7</td>
<td>14.1</td>
<td>20.0</td>
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<td>7.4</td>
<td>16.4</td>
<td>−24.8</td>
<td>18.3</td>
<td>3.7</td>
<td>−37.7</td>
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<tr>
<td>EXP2</td>
<td>7.9</td>
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<td>17.7</td>
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<td>−27.3</td>
<td>16.4</td>
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<tr>
<td>E2PYC</td>
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<td>12.5</td>
<td>17.5</td>
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<td>4.9</td>
<td>14.6</td>
<td>−28.1</td>
<td>16.0</td>
<td>3.3</td>
<td>−42.8</td>
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</table>

Table 4
Annual average 2-m air temperature anomalies (experiment minus control) (°C) as simulated in the model experiments

<table>
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<tr>
<th></th>
<th>Global</th>
<th>Global land</th>
<th>Global ocean</th>
<th>Tropical land</th>
<th>Tropical ocean</th>
<th>NH land</th>
<th>NH ocean</th>
<th>NH ice</th>
<th>SH land</th>
<th>SH ocean</th>
<th>SH ice</th>
</tr>
</thead>
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<tr>
<td>CLIMAP</td>
<td>−4.5</td>
<td>−2.4</td>
<td>−2.0</td>
<td>−3.2</td>
<td>−1.3</td>
<td>−2.8</td>
<td>−1.8</td>
<td>−6.3</td>
<td>−2.9</td>
<td>−0.2</td>
<td>−35.2</td>
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<tr>
<td>EXP2</td>
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<td>−3.1</td>
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<td>−4.9</td>
<td>−3.2</td>
<td>−8.8</td>
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<td>−39.0</td>
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<tr>
<td>E2PYC</td>
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<td>−3.6</td>
<td>−9.6</td>
<td>−5.2</td>
<td>−0.6</td>
<td>−40.3</td>
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</table>
Fig. 6. (a) SST anomaly map (Combined–CLIMAP) for August. The combined SST field is 1.5°C cooler than CLIMAP. (b) SST anomaly map (Combined–CLIMAP) for February. The combined SST field is 1.6°C cooler than CLIMAP.
Fig. 7. (a) SST anomaly map (Combined–control) for August. The combined SST field is 1.5°C cooler than the Control. (b) SST anomaly map (Combined–control) for February. The combined SST field is 2.6°C cooler than the Control.
term is specified because the warm bias adjustment is particularly applicable to the warmest faunal-based temperatures (the maximum temperature in the faunal reconstruction is 29°C).

For $A < 0$, the SSTs calculated by Eq. (1) are colder than the baseline temperatures; for $A = -0.1$ a CLIMAP temperature of 30°C is reduced by 3–27°C. The constant $k$ changes the magnitude of the adjustments for temperatures <30°C. For $k = 0$, the term $(1 + A)$ yields a linear temperature decrease, i.e., at 20°C the cooling is 2°C and at 10°C the cooling is 1°C. As the value of $k$ is increased the overall cooling of the LGM reconstruction is reduced (Fig. 3), the change in thermal gradients at higher latitude (areas of lower temperatures) is reduced, and the largest adjustments are applied to the warmest ocean temperatures.

The second and third terms in Eq. (1) adjust SSTs for biases associated with the pycnocline gradient. We use the modern gradients obtained from the World Ocean Atlas (1998); (Fig. 4) in all adjustments to the baseline LGM SSTs. Ideally, LGM estimates of pycnocline strength would be used to determine the magnitude of the adjustment, but lacking sufficient LGM data, we make the bias adjustments using modern pycnocline strengths on the premise that the first-order regional patterns of pycnocline strength did not change much during the LGM (Ravelo and Andreasen, 1999). Systematic errors associated with using modern pycnocline gradients for adjusting LGM temperatures are likely small relative to the improvements in the regional SST estimates associated with approximate bias adjustments. The modern pycnocline gradients yield a bias adjustment that is greatest in the eastern equatorial Pacific, north of the equatorial divergence zone (Fig. 4); very small adjustments are applied at higher latitudes.

3. LGM surface ocean temperature adjustments

We applied a temperature-based bias correction calculated with $k = 0.1$ and $A = -0.1$ ($B = 0$ and $C = 0$) in Eq. (1). As discussed above, increasing $k$ values applies the greatest cooling to the warmest SSTs while reducing the net cooling at lower temperatures. Much of the cooling relative to the baseline case is thus located in the western warm pool of the Pacific and, to a lesser extent, in the eastern equatorial oceans (Fig. 5). The resulting change in area-weighted global average SST is $-0.97°$C (in addition to the $1.8°$C glacial–interglacial cooling of the unadjusted SSTs). With the exception of core V21-30 from the eastern equatorial Pacific, agreement among the geochemical proxy SST estimates and the adjusted SSTs is better than that of the baseline case (Tables 1 and 2), especially in the warm pool of the western Pacific.

To apply the pycnocline adjustment, we set coefficients $B$ and $C$ to values of 0.4 and $-1.0$, respectively, in Eq. (1) (with $k = 0.1$ and $A = -0.1$ as above) Adding the pycnocline bias adjustment (Fig. 6) results in an additional cooling (relative to CLIMAP) of $1.5°$C in area-weighted global average SST. The combined adjustments to the baseline LGM SSTs reduce discrepancies between the LGM temperature field and various geochemical estimates of SST (Tables 1 and 2). In the tropical Pacific, agreement is improved over the Panama Basin and the western Pacific warm pool. In the western equatorial Pacific, the bias-corrected LGM SST change is essentially identical to Mg/Ca estimates (Table 2). Our glacial–interglacial SST anomalies in the South China Sea ($-4–5°$C) are similar to recent faunal–based estimates ($3–4.5°$C; Chen et al., 2005). Mid-latitude cooling in the eastern Pacific and southwest Pacific is also in better agreement with alkenone–derived LGM temperature estimates. The biggest change from the baseline reconstruction is the marked reduction in the region of warmer-than-modern SSTs in the central gyres of the North and South Pacific (Fig. 7).

Additional insight into the nature of our SST field is illustrated by plotting and comparing the zonal average temperature in the western Pacific along a transect from 35°S to 60°N averaged over the longitudinal band 150°E to 180°E (Fig. 8a) and in the equatorial Pacific along a transect averaged from 10°S to 10°N (Fig. 8b). The mid- to high-latitude SSTs in our bias-corrected SSTs are similar to those reconstructed by Kucera et al. (2005). In agreement with other geochemical-based estimates of SST, our SST field displays substantially more cooling than Kucera et al. (2005) in the low-latitude equatorial transect (Fig. 8b). We note that Kucera et al. (2005) reconstructed LGM SSTs with zonal averages warmer than modern averages.

4. Sensitivity tests with a climate model

The goal of our modeling experiment is to assess the sensitivity of the climate system to changes in tropical and subtropical temperatures. We present four, 30-year climate simulations conducted with the GENESIS global climate model to quantify the effects of modifying the CLIMAP SST reconstruction. Three ice-age simulations were run in addition to a present-day (labeled CTL) simulation: (1) ice-age
in which the baseline SST field was specified (labeled CLIMAP), (2) ice-age in which our warm pool adjustment is specified (labeled EXP2), and (3) ice-age in which our combined SST field was specified (labeled E2PYC). Other than the SST fields, standard LGM boundary conditions [21 ka coastlines, continental ice sheets, atmospheric composition (200 ppmV CO$_2$), and orbital parameters (21 ka, which are nominally the same as present)] were prescribed in the three paleosimulations. The first 10 years of each model run are excluded as model spin-up, and data analyses are over the last 20 years of each model run.

4.1. Atmospheric general circulation model responses

The colder sea surface in the warm pool and the subtropics induces greater atmospheric cooling in the EXP2 and E2PYC experiments than is produced by the CLIMAP experiment (Tables 2 and 3). Regional patterns of atmospheric cooling are primarily the result of near-field changes in SSTs, and secondarily the result of transmission and propagation through circulation at the surface (changes in sea level pressure patterns and winds) and aloft (changes in mid-troposphere wind patterns and planetary wave patterns). These changes are somewhat amplified in the Northern Hemisphere by the presence of the continental ice sheets and the sensitivity of the model to the ice. Ice age global surface-air cooling is 4.5, 4.8, and 6.4 °C, respectively, for the CLIMAP, EXP2, and E2PYC experiments (Tables 3 and 4). Global cooling over land (as determined separately by the present and LGM land masks) ranges from 2.4 to 4.8 °C, which falls within the range of 1.8–5.4 °C of the mixed-layer models in the PMIP experiments (Pinot et al., 1999). The ratio of terrestrial to oceanic cooling in our simulations (1.2–1.4 °C) is also in agreement with the mean value of 1.3 °C cited for the mixed-layer models.

Comparison of our model results in the tropics (30° N to 30° S) with results from the mixed-layer models used in the PMIP experiments (Pinot et al., 1999) and several coupled atmosphere–ocean models of varying complexity (Fig. 9) indicate that: (1) the magnitude of cooling in both the EXP2 and E2PYC experiments falls on the high end of the cooling simulated by the mixed-layer models and some coupled models, and (2) roughly twice as
much cooling is simulated over land than is prescribed over the oceans. The CLIMAP simulation, in contrast, falls in the warmer group of mixed-layer models (and is essentially equivalent to the GENESIS PMIP experiment). Vegetation–based estimates of LGM cooling over tropical land as expressed by the mean temperature of the coldest month (MTCO) ranges from 2 to 6°C (Ferrera et al., 1999). Cooling of MTCO in the tropics is 3.2, 5.4, and 5.6°C, respectively, for the CLIMAP, EXP2, and E2PYC experiments.

Table 5
Annual average precipitation (mm d⁻¹) as simulated in the model experiments

<table>
<thead>
<tr>
<th></th>
<th>Global</th>
<th>Global land</th>
<th>Global ocean</th>
<th>Tropical land</th>
<th>Tropical ocean</th>
<th>NH land</th>
<th>NH ocean</th>
<th>NH ice</th>
<th>SH land</th>
<th>SH ocean</th>
<th>SH ice</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>3.1</td>
<td>2.4</td>
<td>3.5</td>
<td>3.3</td>
<td>4.1</td>
<td>2.2</td>
<td>3.9</td>
<td>1.0</td>
<td>2.8</td>
<td>0.5</td>
<td>0.1</td>
</tr>
<tr>
<td>CLIMAP</td>
<td>2.8</td>
<td>1.9</td>
<td>3.5</td>
<td>2.6</td>
<td>4.2</td>
<td>1.5</td>
<td>3.7</td>
<td>0.9</td>
<td>2.8</td>
<td>0.6</td>
<td>1.3</td>
</tr>
<tr>
<td>EXP2</td>
<td>2.6</td>
<td>2.2</td>
<td>3.1</td>
<td>3.0</td>
<td>3.6</td>
<td>1.7</td>
<td>3.3</td>
<td>0.7</td>
<td>3.2</td>
<td>0.6</td>
<td>1.1</td>
</tr>
<tr>
<td>E2PYC</td>
<td>2.6</td>
<td>2.2</td>
<td>3.0</td>
<td>3.1</td>
<td>3.6</td>
<td>1.8</td>
<td>3.3</td>
<td>0.7</td>
<td>3.1</td>
<td>0.6</td>
<td>1.0</td>
</tr>
</tbody>
</table>
All three LGM simulations yielded reduced average precipitation rates in the tropics and Northern Hemisphere (but not necessarily reduced net moisture, P–E) relative to the control (Tables 5 and 6). In contrast, all three LGM simulations yielded precipitation rates that exceed the control when averaged over the Southern Hemisphere. The range of precipitation reduction, 0.1–0.4 mm d\(^{-1}\), is comparable to the range of 0.1–0.9 mm d\(^{-1}\) for the PMIP mixed-layer models, which display considerable spatial variation in the patterns of simulated precipitation.
4.2. Data-model comparisons

The SST field used in the E2PYC experiment is oceanographically more realistic than that of the EXP2 experiment because cooling in the warm pool region is linked with (reduced) cooling in the subtropics, thereby preserving observed SST gradients. Regions in the baseline reconstruction where LGM SSTs exceed modern values are also adjusted to be equal to or less than CLIMAP values in the E2PYC SST field.

Hereafter we evaluate the modeled temperature responses by comparing the E2PYC and CLIMAP simulations to each other and to data-based estimates of annual temperature change from various geologic records (e.g., ice cores, ground water, pollen) and vegetation-based estimates of changes in the mean temperature of the coldest month (MTCO) as published by Farrera et al. (1999). MTCO is a primary control of plant distribution that varies by latitude and elevation. We did not apply any topographic adjustments to simulated air temperatures to
account for differences in the GCM grids and actual elevations, nor did we attempt to stratify the model output into elevation bands.

The broad patterns of mean–annual global cooling in the E2PYC and CLIMAP simulations (Fig. 10) are typical of most LGM simulations in response to ice age atmospheric CO₂ levels, the distribution of continental ice sheets, and colder SSTs. The regions of greatest cooling are located over the high latitudes and polar regions where the continental ice sheets and expanded sea ice impart a strong climatic effect. The E2PYC simulation is globally colder than CLIMAP in response to our SST changes in the tropics and subtropics. Coherent patterns of colder temperature are simulated over the continents, with marked cooling over much of the Laurentide Ice sheet. Similar global patterns and model–to–model differences are displayed in the changes in MTCO (Fig. 11). The magnitude of the changes in MTCO are, in general, greater than those of annual mean temperature indicating asymmetry in cooling over the annual cycle. This may reflect a of the need for specifying changes in the LGM annual cycle of SST, particularly in the tropics.

In response to global cooling, global precipitation during the ice age is reduced in both the E2PYC and CLIMAP (Fig. 12, Table 4) model runs, with the greatest drying located in the tropics, low latitudes and over the ice sheets. The global distribution of precipitation changes are similar in the two simulations; however, cooling the warm pool and modifying the magnitude of subtropical SST gradients redistributes atmospheric pressure at the surface and aloft (not shown), which in turn modify wind patterns and regions of convection, suggesting the possibility of associated changes in sea surface salinity over the tropics and subtropics. Over land, the largest responses are over central and northern South America, central Africa, and in Southwest Asia, Indonesia, and Australia. The LGM patterns of net moisture (Fig. 13) are similar to those of the precipitation fields with additional temperature–related effects on evaporation and evapotranspiration.

Cooling in the warm pool and subtropical SSTs results in ice-age MTCOs in the low latitudes that are clearly colder in the EP2PYC simulation than those of the CLIMAP simulation (Fig. 14a,b). In Fig. 14, the simulated temperatures are distance-weighted interpolations of the land-based model grid cells (four in most cases) around each data site. Interpolation was used to avoid point-sampling along steep gradients. T-tests (null hypothesis is that the means are equal) of the simulated and reconstructed temperature anomalies indicate that the mean glacial–interglacial anomalies for the E2PYC simulation are statistically equivalent (P = 0.05) (1) to the reconstructed MTCOs, (2) to the globally distributed mean annual air temperature changes, (3) to the pooled MTCO and mean annual data and (4) to five PMIP regions. The means of the
CLIMAP anomalies are statistically equivalent ($P = 0.05$) only to the globally distributed mean annual air temperature changes and the to the five PMIP regions (Fig. 14d). As is indicated by the inset box–and–whisker plots, the EP2PYC simulation produces slightly better agreement with the land data than does the CLIMAP simulation, particularly over the PMIP regions where the CLIMAP temperatures display a warm bias.

We evaluate simulated changes of LGM moisture using inferences of plant available moisture (PAM) and $P–E$ (Fig. 15). Inferred PAM values are compared to changes in precipitation at the data site location. The majority of the PAM inferences are drier than present, reflecting in part the low-latitude focus of the data sites. Overall, 70% of the CLIMAP and 65% of the E2PYC precipitation anomalies agree in sign with the PAM data. This comparison must be viewed with caution, however, because responses of LGM vegetation that appear to have been controlled by moisture availability may in fact have been the result of lowered CO$_2$ levels on plant physiology and associated transpiration efficiency (Jolly and Haxeltine, 1997). Agreement in sign for the simulated and inferred $P–E$ changes (Fig. 11a) are similar to those of the PAM comparison; 65% for both the EP2PYC and CLIMAP simulations.

4.3. Ice sheet responses

Changes in tropical and subtropical SSTs induce responses in the simulated mass balance of Northern and Southern hemisphere ice sheets and ice caps (as they are specified in the model land mask, Fig. 11). The largest responses, located over the Laurentide ice sheet, are the result of competing influences in the mass balance from changes in moisture and temperature. Annual temperatures are lower over the entire ice sheet (Fig. 10) which reduce already low precipitation rates (Fig. 11) and yields a more negative mass balance, particularly over the interior of the ice sheet (Fig. 16). Along the ice margin, however, colder temperatures during the summer ablation season reduce both the amount of ablation and associated runoff, yielding a more positive mass balance. Reduced mass loss in the EP2PYC simulations offsets more negative mass balance in the interior, resulting in a net positive mass balance over the Laurentide of 50 mm; the CLIMAP simulation produced a total mass balance of $-32$ mm, a value comparable to previous applications of GENESIS (Pollard and Thompson, 1997). The apparent direct linkage between tropical and subtropical SSTs and the mass balances of the Northern Hemisphere ice sheets is consistent with other modeling studies (e.g., Rodgers et al., 2003) and adds support to hypotheses that tropical and subtropical ocean temperatures may have played a role in triggering glacial changes at higher latitudes (Cane, 1998; Stott et al., 2002).

5. Discussion

Our adjusted SST fields attempt to address the long-standing controversy about the magnitude of glacial cooling estimated by CLIMAP for the tropics and subtropics, and the western Pacific warm pool in particular, and to evaluate the sensitivity of global climate systems to relatively small adjustments in SST in these regions. We have focused on preserving realistic temperature gradients and distributions by basing changes to the
faunal temperature estimates on likely biases in the method and by employing oceanographically plausible mechanisms in the bias adjustments. In doing so, our adjusted SST field is in better agreement with recent geochemical inferences than are some of the original CLIMAP estimates.

When used as a boundary condition in the climate model, our SST field clearly induces substantial global and regional responses over both land and oceans, including the Northern Hemisphere ice sheets. A central issue of the model application is whether or not the new SSTs lead to an “improved” simulation of LGM climatology. Our data-model comparisons suggest that, relative to the CLIMAP SSTs, the bias-adjusted SSTs result in a colder modeled LGM climate over land with greater cooling in the northern hemisphere and tropics and less cooling in the southern hemisphere. The spatial pattern of MTCO appears to provide a better match with the data (Fig. 11). As measured by air temperature changes, there is statistical improvement over CLIMAP, and our SST fields are in better agreement with recent marine data. Improvement in simulated precipitation is not statistically significant. Problems with plant-based moisture inferences notwithstanding, our results are consistent with many other modeling experiments in which matching reconstructed moisture levels with model simulations has proven to be a challenge (e.g., Pinot et al., 1999, Bonfils et al., 2004; Ruter et al., 2004).

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References

Lea, D.W., Pak, D.K., Paradis, G., 1999. Influence of volcanic shards on foraminiferal Mg/Ca in a core from the Galapagos Region. Geochemistry, Geophysics, Geosystems 6, Q11P04.


Morey, A.E., Mix, A.C., Pisias, N.G., 2005. Planktonic foraminiferal assemblages preserved in surface sediments correspond to multiple environmental variables. Quaternary Science Reviews 24, 925–950.


Ortiz, J.D., Mix, A.C., 1997. Comparison of Imbrie-Kipp transfer function and modern analog temperature estimates using sediment trap and core top foraminiferal faunas. Paleoceanography 12, 175–190.


