Tropical Temperature Variations Since 20,000 Years Ago: Modulating Interhemispheric Climate Change

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Tropical sea surface temperatures (SSTs), as thermodynamically recorded in Barbados corals, were 5°C colder than present values 19,000 years ago. Variable tropical SSTs may explain the interhemispheric synchroneity of global climate change as recorded in ice cores, snowline reconstructions, and vegetation records. Radiative changes due to cloud type and cloud cover are plausible mechanisms for maintaining cooler tropical SSTs in the past.

Estimates of global temperature change for a doubled-CO_2 world range from 1° to 4°C, in large part depending on the simulated change in tropical sea surface temperatures (SSTs) (*1*). Simulations of tropical SSTs range from +1° to +3°C and are strongly dependent on poorly constrained cloud parameterizations and feedbacks in global circulation models (GCMs) (*1, 2*). Whether tropical SSTs are thermostated at modern values is a central issue in climate change forecasts. In support of the thermostat concept, it has been argued that tropical SSTs have remained relatively constant over most of the Cenozoic (*3*) even during intervals of much warmer and colder polar temperatures.

However, a discrepancy between continental proxy records of tropical temperatures and those based from oceanic recorders exists for the Last Glacial Maximum (LGM) (*4*). Specifically, CLIMAP's statistically based SST estimates from planktonic microfossils indicate that global SSTs averaged 1.7°C cooler in August and 1.4°C cooler in February compared with modern SSTs, but with little to no change in the tropics (*5*). In contrast, records of tropical continental climate from snowline reconstructions and vegetation changes indicate that the tropics were 5° to 6°C colder during the LGM (*4, 6*). The lack of change in tropical SSTs as reconstructed by CLIMAP has proved to be an obstacle for some GCM simulations of the LGM (*7, 8*).

In this report, we determine tropical SSTs between 18,000 and 8,000 years ago by applying two independent methods, oxygen isotope thermometry (*9, 10*) (Fig. 1A) and Sr/Ca thermometry (*11–13*) (Fig. 1A), to a well-dated sequence of submerged coral reefs drilled off the south coast of Barbados. Barbados corals monitor the temperature of the deep surface mixed layer of the western equatorial Atlantic, the Atlantic "warm pool." Cores recovered from the Barbados offshore reefs constitute a nearly continuous sequence of coral reef growth spanning the last astronomical climate cycle (*14, 15*). These cores contain abundant specimens of the reef crest coral *Acropora palmata*, which are ideal for measuring paleo-SSTs with great accuracy and precision. The same reef crest coral samples have been dated with the $^{230}Th/^{234}U$ and ^{14}C techniques (*14–16*). This record allows a precise determination of paleo–sea level and thus the means to remove explicitly the component attributed to global ice volume from the measured $\delta^{18}O$ value of the coral specimens. The residual signal ($\Delta\delta^{18}O$) is controlled by temperature and local changes in the $\delta^{18}O$ value of seawater (*17, 18*).

Oxygen isotope values (*19*) of *A. palmata* exhibit a glacial to interglacial amplitude of 2.27 per mil (Fig. 1B). Approximately 1.25 per mil of that signal is a result of global ice volume; thus, the residual change is ~1.0 per mil (Fig. 1C) (*20*). If explicitly a function of temperature, this residual indicates that waters were ~5°C colder than present 18,000 to 19,000 years ago. The residual $\Delta\delta^{18}O$ time series reveals that nearly half of the glacial-interglacial temperature change occurred between 18,000 and 14,700 years ago. Between 14,700 and 13,300 years ago, the $\Delta\delta^{18}O$ signal increased by nearly 0.4 per mil; this change indicates that western tropical Atlantic temperatures became temporarily cooler. This time interval coincides with the release of a large pulse of meltwater from the Northern Hemisphere ice sheet, designated MWP1A (*14*), and marks a period of intense ice sheet disintegration that caused sea level to rise at rates exceeding 4 cm/year. The most prominent step in both the $\delta^{18}O$ and $\Delta\delta^{18}O$ records occurred 11,100 years ago, just after MWP1B. Local changes in seawater $\delta^{18}O$ values could have resulted from changes in the regional evaporation or precipitation pattern or from variable influence of the Amazon River.

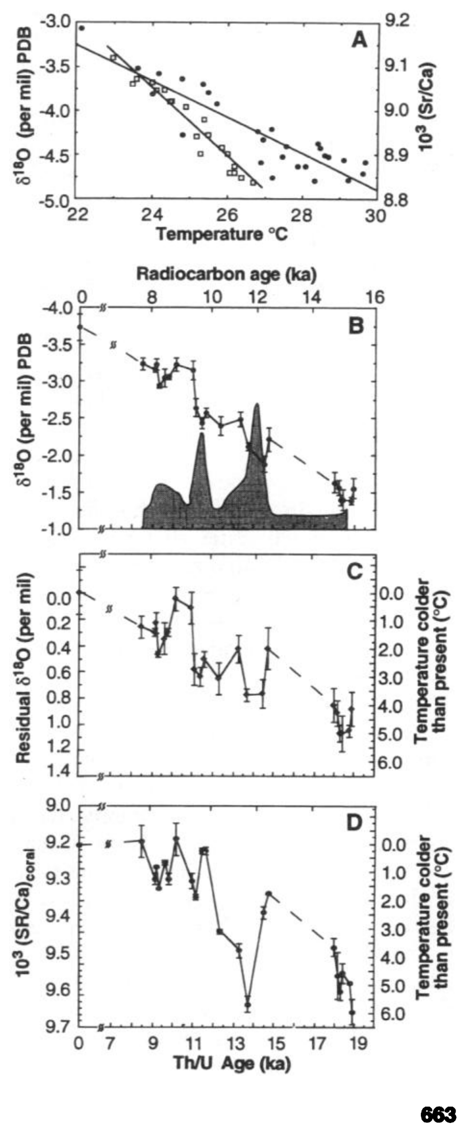

Fig. 1. (**A**) The linear relationship between $\delta^{18}O$ value and temperature in *Acroporid* corals (*35*) (●) and the [Sr/Ca] ratio in coral skeletons as measured by thermal ionization mass spectrometry (TIMS) on Hawaiian *Porites lobata* (*13*) (□). (**B**) Oxygen isotope data from *Acropora palmata* with respect to absolute (lower axis) and radiocarbon (upper axis) time. The record of ice-sheet meltwater discharge, as calculated from the first derivative of the Barbados Sea Level record (*15*) and as discussed in the text (shaded curve). Each sample was composed of powder routed from an average of three transects across coral growth bands encompassing a minimum of 2 years of growth. Error bars are one standard deviation of the average of the transects. (**C**) The oxygen isotope residual after correction for ice volume (*20*) with respect to time as in (B). Conversion to temperature is by the empirical fractionation relationship (*35*). Error bars are the same as in (B). (**D**) $[Sr/Ca]_{atomic}$ determined by TIMS, multiplied by 1000, with respect to time as in (B). Error bars constitute the range of two transects across several annual growth bands (*21*). Conversion to temperature is by the temperature–Sr/Ca relationship for *M. verrucosa* (*11*) as described in the text.

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Although variation in local $\delta^{18}O$ values of seawater is believed to be small, it requires confirmation. For this reason, it is especially important to confirm the $\Delta\delta^{18}O$ paleotemperature estimates with an independent thermometer.

The Sr/Ca ratio in coral skeletons is linearly related to temperature (*11*). When measurements are made by thermal ionization mass spectrometry, this method has a reported precision of better than ±0.1°C (*12*). More recent work has shown that the Sr/Ca ratio in coral skeletons exhibits a slight dependence on calcification or growth rate (*13*). However, as the two thermometers ($\delta^{18}O$ value and Sr/Ca ratio) do not share the same sources of error, these two measurements provide useful cross checks.

The Sr/Ca ratio time series (*21*) is similar to the $\Delta\delta^{18}O$ record (Fig. 1D), although there is a more pronounced increase in the Sr/Ca ratio that reaches glacial values at 13,700 years ago. In order to use the Sr/Ca ratio in coral skeletons for paleothermometry, the Sr/Ca ratio in seawater, $[Sr/Ca]_{sw}$, must either be known or have remained similar to present values. The $[Sr/Ca]_{sw}$ has probably remained constant on time scales approaching the residence times of strontium [2.5 million years (*22*)] and calcium [0.6 to 1.1 million years (*23*)] in the ocean (*24*). Changes in the calcium budget resulting from movement of the carbonate compensation depth during glacial-interglacial cycles, which can change the fraction of $CaCO_3$ buried, are also likely to be minimal (*12*). Thus, the measured ratio in coral skeletons should record only temperature during the last 20,000 years. Fractionation of Sr by the coral skeletons also seems to depend slightly on growth rate, and this effect results in an offset in the Sr/Ca ratio between the top versus sides of columnar coral specimens (*13*). The maximum effect of this offset is 1° to 2°C for growth rates that varied by a factor of 2 from 12 to 6 mm per year. This does not affect the precision of measured temperature changes as long as specimens with similar growth rates are used. All of our analyzed samples have similar linear extension rates (*25*).

An explicit relationship between temperature and Sr/Ca ratio for *A. palmata* is not available because the present-day distribution of *A. palmata* geographically is restricted to the tropical Atlantic Ocean and Caribbean Sea, a region that has minimal range in annual average SSTs (*26*). In lieu of an explicit calibration line, we used previously determined temperature–Sr/Ca relations (*11, 12, 13*) to constrain our Sr/Ca paleotemperature estimates. The slopes of the relations between coral Sr/Ca and temperature for the four genera studied are remarkably similar. For our Sr/Ca temperature record we used the relation derived for *Montipora verrucosa*, which is in the same family as *A. palmata* (*27*).

Both data sets indicate that SSTs at Barbados were 5°C cooler 18,000 to 19,000 years ago. It is evident that there was a significant warming preceding the first meltwater pulse, MWP1A, although the absolute timing is unresolved. The Sr/Ca paleotemperature estimate indicates a cooling of 4°C during the early phase of deglaciation ~13,700 years ago, which is greater than the cooling inferred from the $\delta^{18}O$ record of ~2°C. By 11,400 years ago the Sr/Ca temperature estimate is similar to present temperatures, and then varies 1° to 2°C cooler than present temperatures until the end of the time series 8,450 years ago. A similar story unfolds for the $\delta^{18}O$ record except during the times of catastrophic disintegration of the ice sheets (MWP1A and MWP1B) (*14*).

Our Barbados LGM SST estimates are significantly cooler than those reported on the basis of planktonic foraminiferal assemblages for the western equatorial Atlantic (*5, 28*). We believe that this discrepancy results from the hydrographic control of the seasonality and vertical distribution of foraminifera in the tropics, in conjunction with the wide temperature range of the two planktonic species which dominate the deep mixed layer assemblage (*29*). Provided a deep mixed layer persists in the western equatorial Atlantic during the LGM, the same foraminiferal species could dominate the assemblage there (*29*).

Radiative changes must be involved in producing the tropical SST changes. An equatorward displacement of the highly reflective mid-latitude cloud decks could cool tropical SSTs and amplify northern hemisphere glaciation (*30*) as the tropics are in near radiative balance (*30, 31*). This radiative imbalance should be transmitted into the tropics through the mean meridional circulation and affect the type and amount of cloud cover in the tropics and indirectly the amount of water vapor in the atmosphere. In turn, less water vapor reduces the longwave forcing of the atmospheric column (*32*). Changes in low-level cloudiness (*33*) and feedbacks affecting the mean meridional circulation (*34*) are additional means for modifying tropical and extratropical climate.

The 5°C cooling measured at Barbados supports the tropicwide paleotemperature estimates from continents based on snowline and vegetation changes. The results from Barbados indicate that the western tropical Atlantic is sensitive to climate change on glacial-interglacial time scales and is capable of changing quite rapidly, as evidenced by the 4°C shift between 13,700 and 12,000 years ago. In addition to changes in thermohaline circulation and atmospheric CO_2 levels, variable tropical SSTs provide an explanation for the synchroneity of interhemispheric climate change observed in many climate proxy records through changes in the export of energy and water vapor. Our measured variation in tropical SSTs bears directly on modeling future climate with respect to the sensitivity of the tropics to changes in the radiative balance and the influence of the tropics on global climate. Reproducibility of the spatial variation in temperature as reconstructed by CLIMAP (*4*) is used as a benchmark test for many GCMs and coupled ocean-atmosphere GCMs (*1*). This benchmark test must be reevaluated in light of our results.

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per mil (1σ). In an effort to reduce intracoral variability we only sampled the upward-growing portion of coral fronds. To test the variability of our $\delta^{18}O$ measurements we sampled individual transects from the upper surface all the way around to the shaded or bottom side. No significant $\delta^{18}O$ differences were found even though there was a reduction in extension rate from the upper to bottom surface of greater than 50%.
20. $\Delta\delta^{18}O = \delta^{18}O_m - \delta^{18}O_{present} - 0.011UpZ$, where $\delta^{18}O_m$ is the measured value of paleosamples, $\delta^{18}O_{present}$ is the value of recent *A. palmata* collected from Barbados, *UpZ* is the uplift corrected depth using Th/U years, and 0.011 is the maximum ice volume–sea level relationship (*18*). $\Delta T = \Delta\delta^{18}O \cdot \partial T/\partial\delta^{18}O$, where $\Delta\delta^{18}O$ is as defined previously and $\partial T/\partial\delta^{18}O$ is the temperature-dependent oxygen isotopic fractionation, which is 0.21 per mil per degree Celsius for *Acroporids* (*35*).
21. Two individual transects encompassing 2 to 4 years of growth from the upward-growing portion of coral fronds, identical to those used in the $\delta^{18}O$ analyses, were prepared as follows. Samples, approximately 100 to 200 μg, were directly dissolved in a ^{84}Sr-^{43}Ca acid spike solution. One microliter of the solution was then loaded onto degassed W filaments and analyzed on a VG Sector Thermal Ionization Mass Spectrometer in static faraday mode. In determining the Sr concentration 120 ratios were measured whereas 150 ratios were measured for Ca. Mass fractionation correction during analysis assumed a natural $^{86}Sr/^{88}Sr$ = 0.1194 (*36*), and $^{42}Ca/^{44}Ca$ = 0.31221 (*37*).
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Lake-Atmosphere Feedbacks Associated with Paleolakes Bonneville and Lahontan

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A high-resolution, regional climate model nested within a general circulation model was used to study the interactions between the atmosphere and the large Pleistocene lakes in the Great Basin of the United States. Simulations for January and July 18,000 years ago indicate that moisture provided by synoptic-scale atmospheric circulation features was the primary component of the hydrologic budgets of Lakes Lahontan and Bonneville. In addition, lake-generated precipitation was a substantial component of the hydrologic budget of Lake Bonneville at that time. This local lake-atmosphere interaction may help explain differences in the relative sizes of these lakes 18,000 years ago.

At the last glacial maximum 18,000 years ago (18 ka), paleolakes Bonneville and Lahontan existed in the Great Basin of the western United States (Fig. 1). The surface area of Lake Bonneville (47,800 km^2) was about eight times that of the present Great Salt Lake (6200 km^2), and the surface area of Lake Lahontan (14,700 km^2) was about six times that of the present lakes in the Lahontan basin (2500 km^2) (*1–3*). Lake-level chronologies (*2*) indicate that at 18 ka the surface area of Lake Bonneville was nearly as large as the postglacial maximum that occurred at ~14.5 ka, whereas the surface area of Lake Lahontan was about two-thirds of its postglacial maximum, which was at 13.5 ka.

The presence of these lakes from 30 to 12 ka has been linked to atmospheric circulation changes, in particular changes in the mean position of the polar jet stream and associated storm track (*4, 5*). Simulations with general circulation models (GCMs) have shown that the jet stream was split by the Laurentide Ice Sheet, with the southern branch of the jet displaced to the south (*6–9*). Hydrological modeling (*10*) of the Lahontan basin showed that variations of lake levels were consistent with the regional climatic response to these circulation changes. The large sizes of lakes Bonneville and Lahontan suggest that lake-atmosphere interactions (thermal effects and evaporation) within the lake basins may have generated or enhanced precipitation (*5, 11, 12*); thus, to some extent, the large lakes may have been self-maintaining.

To investigate the hierarchy of controls of the pluvial climates of the western United States, we used a regional climate model (RegCM) (*13, 14*). Coastlines and mountainous terrain that exert substantial forcings on the climate of the region were resolved in this model (*15*). An interactive lake model that can simulate lake temperature, evaporation, and ice cover was also included (*16*). We conducted a series of 90-day simulations with RegCM to describe the climate of 18 ka and to separate large-scale circulation changes from small-scale lake-atmosphere feedbacks. The simulations were conducted for perpetual January and July conditions; initial and lateral

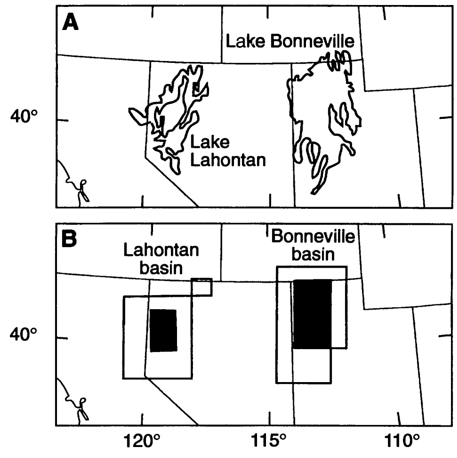

Fig. 1. (**A**) Locations and sizes of lakes Lahontan and Bonneville. Lake extents are for maximum lake sizes. (**B**) Model representation of the 18-ka lake surface areas and the lake basins.

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