

The thermal oceanographic signal of El Niño reconstructed from a Kiritimati Island coral

Michael N. Evans,¹ Richard G. Fairbanks,¹ and James L. Rubenstone

Lamont-Doherty Earth Observatory, Columbia University, Palisades, New York

Abstract. Central equatorial Pacific sea surface temperature (SST) anomaly is a critical predictor of basin-wide oceanographic and atmospheric effects of the El Niño-Southern Oscillation (ENSO) phenomenon. We employ two geochemical thermometers measured on coralline aragonite to reconstruct an independent proxy-based measure of central equatorial Pacific sea surface temperature anomaly. In addition, we assess the observational error associated with extraction of large-scale SST anomalies from $\delta^{18}\text{O}$ and Sr/Ca measurements. On the basis of paired data for the 1981–1987 period, RMS error for the estimation of SST from Kiritimati coral $\delta^{18}\text{O}$ is about 0.4°C, assuming no seawater $\delta^{18}\text{O}$ influence; for Sr/Ca, the observational error is about 0.5°C. Singular spectrum analysis of the $\delta^{18}\text{O}$ time series suggests that 1/3 of the variance is explained by SST anomaly and that this variance may be separated from other signals in the frequency domain. The interannual component of the $\delta^{18}\text{O}$ record shares 70% variance with the interannual component of local SST anomaly estimates ($\rho=-0.84$) and correlates as highly with NINO3 region (150°W–90°W, 5°N–5°S) SST anomaly estimates; comparison of annual averages of the $\delta^{18}\text{O}$ data with analyzed SST for 55 years suggests that the error in deriving annually averaged SST anomalies at Kiritimati is about 0.4°C. A global SST correlation analysis suggests significant correlation with the rest of the Pacific Basin and the tropical Indian and Atlantic sectors. The ocean-wide level of correlation achieved using the coral data is indistinguishable from that achieved using the NINO3 SST anomaly index and suggests that a few well-located coral reconstructions with low observational error may be sufficient to reconstruct the global SST anomaly field associated with ENSO activity.

1. Introduction

The world's oceans have recently been shown to possess global-scale modes of spatial variability [Weare *et al.*, 1976; Latif and Barnett, 1994; Latif *et al.*, 1995; Smith *et al.*, 1994; Kaplan *et al.*, 1997, 1998; Tourre and White, 1995, 1997]. These patterns, which redistribute atmospheric heat and water vapor, play an important

role in interseasonal to interdecadal climate variability [Wright, 1977; Barnston and Livezey, 1987; Webster, 1994; Bell and Halpert, 1995]. Yet much of the world's oceans are not well observed, especially prior to 1950 [Bottomley *et al.*, 1990]. In particular, data are scarce from the remote tropical and high latitude southern oceans, but these regions play a important role in climatic processes [Horel and Wallace, 1981; Kumar *et al.*, 1994; White and Peterson, 1996; Picaut *et al.*, 1996]. As an example from the tropics, sea surface temperature anomaly averaged over the NINO3 region (150°W–90°W, 5°N–5°S) is commonly used as an index of historical El Niño-Southern Oscillation (ENSO)

¹Also at Department of Earth and Environmental Sciences, Columbia University, Palisades, New York.

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activity. However, during both World Wars, observations are almost absent over the entire box and full observational coverage, especially in the western part of the box, is not achieved until the mid-1960s [Fairbanks *et al.*, 1997]. In particular, during ENSO warm phase events, sea surface temperature (SST) is highest in the central equatorial Pacific west of the NINO3 box, potentially making it the most sensitive region for the transfer of heat and moisture anomalies to the extratropics [Pan and Oort, 1983; Gadgil *et al.*, 1984; Graham and Barnett, 1987]. In this critical part of the central equatorial Pacific, including the NINO3 box, SST observational coverage may be inadequate for the study of tropical-extratropical climate interaction.

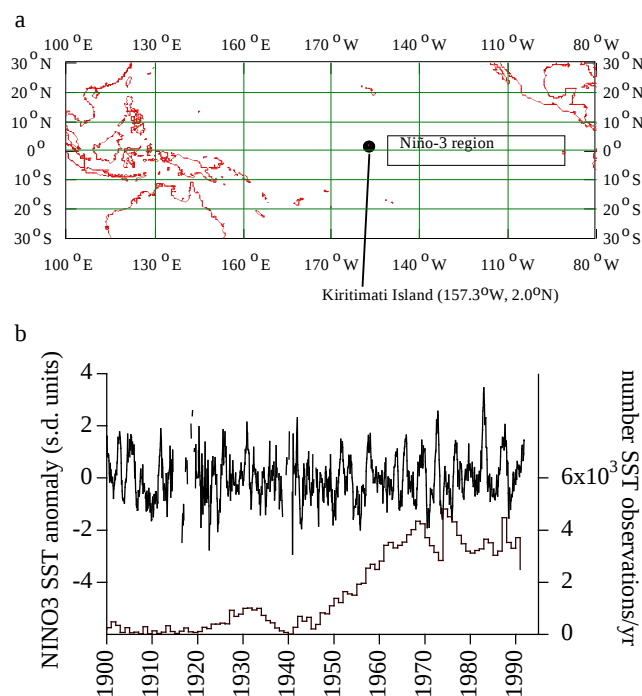


Figure 1a. Map of the tropical Pacific region. NINO3 averaging region and location of Kiritimati Island are as indicated.

Figure 1b. Left axis, monthly time series of NINO3 sea surface temperature (SST) anomaly, 1900–1993, from Global Ocean Surface Temperature Atlas (GOSTA) data [Bottomley *et al.*, 1990], standardized to 1951–1980 mean and standard deviations; right axis, number of SST observations per year in the NINO3 region.

Proxy climate data derived from the isotopic composition and chemistry of aragonite formed by massive reef corals provide a supplementary data set for the study of the tropical near-surface climate [Cole *et al.*, 1993; Dunbar *et al.*, 1994; Linsley *et al.*, 1994; Tudhope *et al.*,

1995; Charles *et al.*, 1997]. Given constant seawater $\delta^{18}\text{O}$ and biological processes, variability in the oxygen isotopic composition ($\delta^{18}\text{O}$) of coralline aragonite is a function of the sea surface temperature in which the coral secreted its skeleton [Epstein *et al.*, 1953; Weber and Woodhead, 1972; McConnaughey, 1989]. However, coral $\delta^{18}\text{O}$ also records changes in the $\delta^{18}\text{O}$ of seawater [Cole *et al.*, 1993; Linsley *et al.*, 1994; McCulloch *et al.*, 1994] if freshwater flux variance is significant.

The strontium:calcium (Sr/Ca) ratio in corals has been shown to be a function of sea surface temperature as well, with minimal seawater Sr/Ca influence [Weber, 1973; Schneider and Smith, 1982; Beck *et al.*, 1992; DeVilliers *et al.*, 1994; Shen *et al.*, 1996]. Application of these two proxy measurements in tandem has been used to distinguish seawater $\delta^{18}\text{O}$ and sea surface temperature anomalies recorded on the Great Barrier Reef during the 1982–1983 El Niño–Southern Oscillation event [McCulloch *et al.*, 1994].

When the proxy data span both sparse and rich observational periods (e.g., in the equatorial Pacific, before and after 1960), an opportunity is provided to cross-validate coral and instrumental data. Potential biases in a coral-based record of SST variability (age uncertainties, sampling artifacts, or other non-climatic information) are independent of those in the instrumental record (instrument calibration, precision, accuracy, measurement method, frequency and spacing of sampling, and analysis technique). Thus coral-based data can provide an independent assessment of the robustness of historical instrumental indices of tropical Pacific climate. Conversely, we can use modern SST data to estimate the observational error expected when extracting SST variability from the coral $\delta^{18}\text{O}$ and Sr/Ca data. In either case, if the instrumental and coral data describe climatic phenomena, then they should agree. Such intercomparison efforts are essential if prehistorical or paleoclimatic information is to be derived from proxy measurements on corals.

We made measurements of stable isotopic composition ($\delta^{18}\text{O}$ and $\delta^{13}\text{C}$) and Sr/Ca on a coral collected live from Kiritimati (Christmas) Island (157.3°W, 2°N) in the Republic of Kiribati. Kiritimati lies at the eastern edge of the western Pacific warm pool and just west of the NINO3 box (Figure 1). The seasonal range of sea surface temperature is less than 1.2°C. Because the seasonal cycle is modest, ENSO-induced warm and cold phase sea surface temperature anomalies of $\pm 3^\circ\text{C}$ are prominent [Bjerknes, 1969; Wyrtki, 1975, 1985; Philander, 1990]. By contrast, ENSO-driven precipitation anomalies are moderate in magnitude and brief in du-

ration, creating a minor change in seawater $\delta^{18}\text{O}$ due to enhanced rainfall [Ropelewski and Halpert, 1987] and eastward advection of lower-salinity water [Picaut et al., 1996]. Thus Kiritimati is an ideal location from which to monitor the thermal signal associated with the full ENSO cycle and should record features similar to those in the NINO3 region. In this paper, we estimate the observational error associated with constructing the SST anomaly record from a Kiritimati Island coral, and we show that not only local but also basin-scale information on the thermal oceanographic ENSO signal is captured by the proxy measurements at this location. This signal is consistent with that recovered by large-scale instrumental indices of interannual SST variability such as NINO3.

2. Data and Methods

2.1. Stable Isotope Data

Coral PP7-3 was collected live and cored at South West Point, Kiritimati Island, at 9 meters depth in March 1994. The coral has been identified as *Porites* sp., probably *P. australiensis*, possibly *P. lobata* (D. Potts, University of California, Santa Cruz, personal communication, 1995). The core was slabbbed along the growth axis and X radiographed for densitometry study; slabs were then ultrasonically cleaned in deionized water and dried at 50°C prior to sampling. Samples for oxygen and carbon isotope analyses were drilled from the slab at 0.5 mm intervals starting from just below the most recent growth and extending to a maximum depth of 1004 mm. A precise age model was recoverable for the upper 839.5 mm. Samples were drilled from lines chosen parallel to the axis of maximum growth to avoid nontemporal (i.e., potential differential, growth related) isotopic effects [McConnaughey, 1989]. X ray diffraction analyses of drilled and chipped coral aragonite verify that samples were 100% aragonite after drilling. Sequential isotopic analyses were made using a Finnigan MAT-251 gas source mass spectrometer coupled to a Carousel-48 automated sample preparation device. Measurement precisions (1σ) were $\pm 0.06\text{‰}$ and $\pm 0.04\text{‰}$ for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ analyses, respectively, based on analysis of a laboratory standard; sample replication over the 1981–1987 period (see section 2.2) gives an external precision of $\pm 0.09\text{‰}$ in the $\delta^{18}\text{O}$ time series shown in Figure 2. All stable isotope measurements are reported relative to the Pee Dee belemnite (PDB) standard.

We use the strongly periodic $\delta^{13}\text{C}$ record to provide an age model for time series analysis of $\delta^{18}\text{O}$ data. The

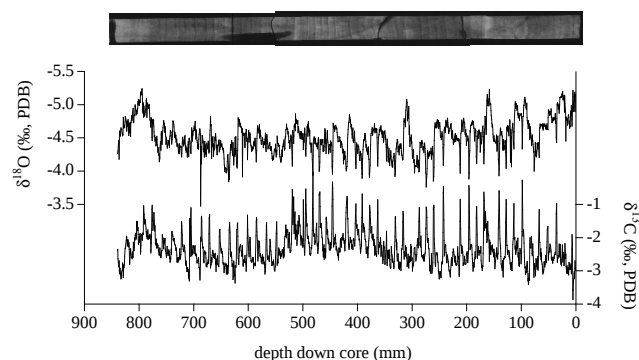


Figure 2. Stable isotopic results from a coral collected at Kiritimati Island. (top) X radiograph negative of the sampled slab from the drill core. Light features correspond to high-density coralline aragonite; dark features correspond to low-density aragonite. Left axis shows oxygen isotopic composition (per mil); right axis shows carbon isotopic composition (per mil). Isotopic results are reported relative to the Pee Dee belemnite (PDB) standard. Horizontal axis shows depth down core (millimeters). Samples were taken at 0.5 mm intervals.

carbon isotopic composition of coralline aragonite varies with seasonal changes in insolation, reproductive activity, and changes in the $\delta^{13}\text{C}$ of ΣCO_2 of seawater [Fairbanks and Dodge, 1979; Patzold, 1984; Gagan et al., 1994, 1996]. The carbon isotope record from coral PP7-3 (Figure 2) shows a series of regular $\delta^{13}\text{C}$ enrichments, which correspond to thin, low density bands. These enrichments are probably too large to be explained entirely by variability of $\delta^{13}\text{C}$ of seawater ΣCO_2 [Swart et al., 1996] or to be solely related to seasonal insolation variability [Fairbanks and Dodge, 1979]. However, a model that invokes the drawdown of ^{12}C within the coral polyp by production of gametes during a brief period of each year, a process called mass spawning [Babcock et al., 1986; Gagan et al., 1994, 1996], may explain the annual enrichments we observe in the carbon isotopic record. There are no published accounts of the timing of mass spawning events in the central Pacific. We assume that such events take place in April of each year, at the onset of the April–September insolation maximum at Kiritimati. Intermittently in the PP7-3 record, single years do not display a $\delta^{13}\text{C}$ enrichment, perhaps due to suboptimal coral reproductive conditions. In these cases we use the same growth rate as the previous and subsequent years to continue the age model, an assumption consistent with an age model based on densitometric observations (Figure 2, top) and relatively constant extension rate (Figure 4). By these criteria, the length of record extends from September

1993 back to April 1938 (Figures 2 and 4). The high (0.5 mm) sampling resolution enabled us to obtain 25–35 samples per annual band and achieve a chronology with biweekly to monthly resolution. The chronology we construct is in good agreement with annual age determinations based on X radiography of the sampled slab and with an age model based on a constant growth rate assumption (Figure 4).

Uncertainty in such an age model might be expected to vary with time before present; therefore a single error estimate for the entire time period is unsatisfactory. Since the timing of mass spawning at Kiritimati has not been observed, we conservatively assume an initial age model error of $\pm 1/2$ year, and we estimate the age model error as a function of the absence of the annual $\delta^{13}\text{C}$ chronometer. The mean extension rate is 14 mm/yr, with a standard deviation of 1.5 mm/yr (11%). Therefore we estimate age model uncertainty for each year within the preceding 5 decades as $(0.5 + 2 \times n \times \sigma_{\text{growth rate}})$ years, where n is the number of years in that decade in which the $\delta^{13}\text{C}$ enrichment is absent. These errors range from ± 0.5 to ± 1 year per year of age model (Figure 4). We believe that uncertainty in the age model is well represented by these estimates. Initially, we derived cumulative age model error estimates, but it was clear by this model that these error estimates were too large in the earlier half of the record and too small in the more recent half. A better error representation was obtained with the error model below, which acknowledges that periods of consistently periodic $\delta^{13}\text{C}$ pulses provide better age control than periods with more intermittent reproductive activity. Subsequent intercomparison of the Kiritimati $\delta^{18}\text{O}$ record with the historical SST analysis by *Kaplan et al.* [1998] suggests our age model estimates are probably appropriate.

2.2. Sr/Ca Data

Sr/Ca may be the preferred proxy for the purpose of SST reconstructions, since, unlike $\delta^{18}\text{O}$, Sr/Ca variability is solely a function of SST. However, since analysis of $\delta^{18}\text{O}$ is more easily automated, requires less sample preparation, and is 10–20 times faster than measurement of Sr/Ca by thermal ionization mass spectrometry (TIMS), we use limited, high-precision TIMS Sr/Ca measurements to confirm the interpretation of the Kiritimati $\delta^{18}\text{O}$ time series as primarily driven by regional SST anomaly. On the basis of the carbon isotope age model, we resampled the 1981–1987 period (Figure 3a). Samples were drilled at 1 mm intervals in a transect parallel to the line chosen for the high-resolution stable

isotopic measurements described above. The samples created were split into two portions; one for replicate stable isotope analysis and one for Sr/Ca analysis. Stable isotope and Sr/Ca sample sizes were 70–120 and 130–250 μg , respectively. The replicate stable isotope measurements were used to verify the chronology for the Sr/Ca data and to assess the reproducibility of the oxygen isotope anomaly record. Stable isotope measurements were made using the previously described procedures. Sr and Ca concentrations were measured using the isotope dilution technique on a VG Sector thermal ionization mass spectrometer, using enriched ^{84}Sr and ^{43}Ca spikes. In-run isotopic fractionation was corrected using measured $^{86}\text{Sr}/^{88}\text{Sr}$ and $^{42}\text{Ca}/^{44}\text{Ca}$ and an exponential fractionation law [*Russell et al.*, 1978]. Duplicate or triplicate runs were made on each Sr/Ca sample and averaged. The standard error (2σ) on the Sr/Ca measurements is less than 0.03 mmol/mol, based on internal precision, duplicate analyses, and replicates; this translates into an SST error estimate of less than 0.5°C .

2.3. SST Observational Error Estimates

We compared estimates of SST variability extracted from the Sr/Ca and stable isotope data to local estimates of instrumental SST variability to estimate the observational error in the two geochemical thermometers. We computed the linear correlation coefficient r between monthly interpolated $\delta^{18}\text{O}$ and Sr/Ca data and SST anomaly estimates for the $2^\circ\text{x}2^\circ$ grid box enclosing Kiritimati, from the analysis of *Reynolds and Smith* [1994], for the period 1981–1987. The least squares regression equation can be rearranged to form the following relationship: $\sigma_{\delta^{18}\text{O}}(\text{SST anomaly}) \approx \sigma_{\text{SST anomaly}} \times (1 - r_{\text{SST, proxy}}^2)^{1/2}$. We can construct a more robust estimate of the observational error in this manner, using annual averages of coral $\delta^{18}\text{O}$ anomaly for the full record length (1938–1993) and SST anomaly from a reanalysis of the U.K. Meteorological Office Historical $5^\circ\text{x}5^\circ$ gridded data set (MOHSST5) [*Bottomley et al.*, 1990; *Kaplan et al.*, 1998] (also see Figure 6). The *Kaplan et al.* [1998] analysis uses observed modern spatial SST covariance patterns and a temporal first-order autoregressive model to estimate the SST field for locations and periods for which observations are sparse, minimizing the variance-weighted residuals between the analysis and the covariance patterns and between the analysis and observations [*Kaplan et al.*, 1998].

2.4. Time Series Analysis of the Stable Isotope Time Series

We applied singular spectrum analysis (SSA) to the oxygen isotopic data to extract the interannual component of the $\delta^{18}\text{O}$ record. SSA is a form of empirical orthogonal analysis applied in the time domain, which looks for autocovariance patterns in a matrix of lagged copies of a single time series [Vautard and Ghil, 1989; Vautard et al., 1992; Ghil and Mo, 1991]. The advantage of SSA over traditional Fourier analysis in the study of time series is its ability to separate secular variability from quasi-oscillatory signals in a time series. In addition, reconstructed components (RCs) can be formed, representing the signal of interest in the time domain without loss of record length. Summing all RCs returns the original time series. SSA has been successfully applied to climatic data series to extract signals with timescales of seasons to millennia [Ghil and Vautard, 1991; Penland et al., 1991; Vautard et al., 1992; Schlesinger and Ramankutty, 1994; Yiou et al., 1994; Linsley et al., 1994; Cook et al., 1996].

To interpret the signals extracted from the data series, we have examined the Fourier transforms of the individual periodic reconstructed modes (results not shown). We expect quasi-periodic RCs with periods of 2–10 years to be associated with ENSO-driven SST anomalies. From inspection of Figure 2, we also expect to see a time-varying annual signal and secular or low-frequency (20–30 year period) variability.

3. Results

The results of paired $\delta^{18}\text{O}$ and Sr/Ca measurements are shown in Figure 3a. The time series spans 1981–1987, a period of significant ENSO activity. Also shown are observational uncertainties for both proxies and SST anomaly estimates for the $2^\circ \times 2^\circ$ grid box enclosing Kiritimati, from the analysis of Reynolds and Smith [1994]. The three ordinate axes shown have been scaled to equivalent ranges, based on literature estimates of the dependence of the two proxies on SST [Weber and Woodhead, 1972; DeVilliers et al., 1994]: 1.6‰ $\delta^{18}\text{O}$ anomaly ≈ 0.6 mmol/mol Sr/Ca anomaly $\approx 8^\circ\text{C}$. Both $\delta^{18}\text{O}$ and Sr/Ca anomaly track the local ($2^\circ \times 2^\circ$ average) SST anomaly through this period, which includes the 1982–1983 ENSO warm phase event, return to cool conditions in 1984–1985, and the onset of the 1986–1987 warm phase event. Figure 3a also shows that during this period there are $\delta^{18}\text{O}$ enrichments of 0.4‰ amplitude and 1–2 month duration, which are matched by neither the Sr/Ca anomaly record nor the SST anomaly record.

Weekly SST time series from the National Oceanic and Atmospheric Administration (NOAA) buoy at (155°W , 2°N) suggest that this feature is not SST related. Time series of seawater $\delta^{18}\text{O}$ measurements from Kiritimati for 1995 and 1996 show that this feature is not due to seawater $\delta^{18}\text{O}$ variability [Fairbanks et al., 1997]. This annual $\delta^{18}\text{O}$ signal is also evident throughout the 1938–1993 $\delta^{18}\text{O}$ time series and is coincident with annual $\delta^{13}\text{C}$ enrichments described above (Figures 2 and 4).

Error estimates for the extraction of SST anomaly from these two proxies are somewhat larger than the analytical uncertainties described above. On the basis of potential variability in the Sr/Ca ratio of seawater and uncertainty in the Sr/Ca–SST relationship (slope and intercept), root mean-square (RMS) error in application of Sr/Ca thermometry has been shown to be closer to $\pm 0.7^\circ\text{C}$ (DeVilliers et al. [1994] and Shen et al. [1996], ± 0.06 mmol/mol). Excluding the $\delta^{18}\text{O}$ data for the suspected spawning periods noted above (Figure 3a), we compared the paired Kiritimati $\delta^{18}\text{O}$ and Sr/Ca to $2^\circ \times 2^\circ$ gridded estimates of SST from the analysis of Reynolds and Smith [1994]. The following relationships are observed between anomaly measurements (Figures 3b and 3c):

$$\delta^{18}\text{O} = -0.067(\pm 0.018) + [-0.18(\pm 0.016)][\text{SST}] \quad r = -0.84$$

$$\text{Sr/Ca} = 0.010(\pm 0.008) + [-0.0614(\pm 0.007)][\text{SST}] \quad r = -0.77$$

The squared linear correlation coefficient of the $\delta^{18}\text{O}$ anomaly–SST anomaly regression is $r^2 = 0.7$ (Figure 3b); for Sr/Ca anomaly–SST anomaly, $r^2 = 0.6$ (Figure 3c). The standard deviation of the monthly sea surface temperature anomaly estimates for this location is 1.2°C . Using the relation between r , $\sigma_{\text{SST anomaly}}$ and observational RMS error noted above, the levels of correlation indicated suggest that RMS errors in SST anomaly estimates from $\delta^{18}\text{O}$ and Sr/Ca anomalies are roughly $\pm 0.4^\circ\text{C}$ and $\pm 0.5^\circ\text{C}$, respectively, similar to those achieved by others [Beck et al., 1992; DeVilliers et al., 1994; Shen et al., 1996]. These uncertainty estimates are plotted in Figure 3a. By comparison, error estimates for monthly SST analyses for the Kiritimati region are less than 0.1°C over the same time period [Kaplan et al., 1998; Reynolds, 1988]. The complete $\delta^{18}\text{O}$ time series, on which further statistical analyses are performed, is shown in Figure 4. The data have been normalized to the 1951–1980 mean and standard deviation of -4.51‰ and 0.25‰ ; the $\delta^{18}\text{O}$ observational error based on the $\delta^{18}\text{O}$ anomaly–SST anomaly regression is also plotted in standard deviation units and has an amplitude of ± 0.4 RMS.

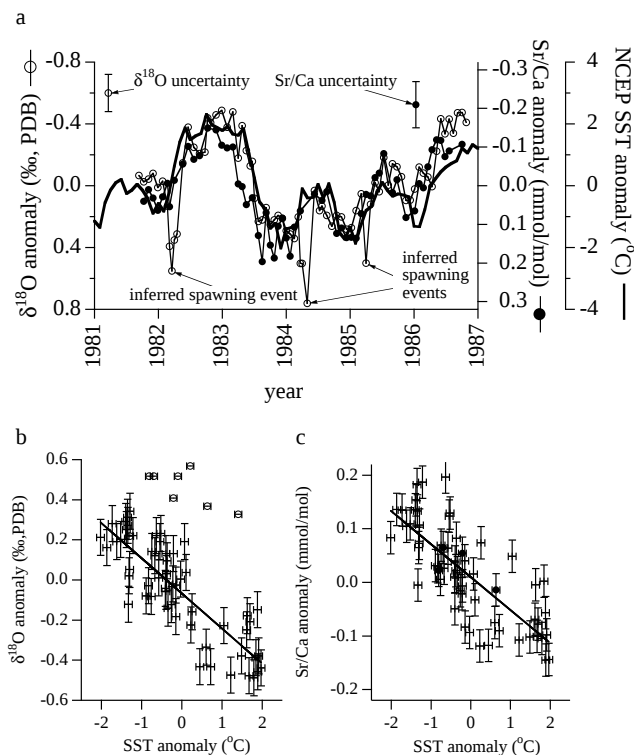


Figure 3a. Kiritimati coral $\delta^{18}\text{O}$ and Sr/Ca anomalies relative to the mean for the 1981–1987 period. Left axis shows $\delta^{18}\text{O}$ anomaly (per mil, PDB). Right axis shows Sr/Ca anomaly (mmol/mol). Observational errors for each proxy are as indicated, based on the analyses shown in Figures 3b and 3c. Also plotted are sea surface temperature anomaly estimates (far right axis, solid line) from the optimal interpolation analysis of Reynolds and Smith [1994] for the $2^\circ \times 2^\circ$ grid box encompassing Kiritimati. The data have been plotted on a common range so that direct comparisons of the different measurements may be made: $8^\circ\text{C} \approx 1.6\text{‰}$ in $\delta^{18}\text{O}$ ≈ 0.6 mmol/mol in Sr/Ca. Note that the relationships between $\delta^{18}\text{O}$ and SST and Sr/Ca and SST are negative and linear; light $\delta^{18}\text{O}$ values correspond to high SST, and low Sr/Ca ratios correspond to high SST. On the basis of comparison with Sr/Ca and SST anomalies and seawater $\delta^{18}\text{O}$ variability (see text), $\delta^{18}\text{O}$ anomalies not attributable to SST anomalies nor precipitation influences have been labeled “inferred spawning events” to denote periods of suspected coralline reproductive activity.

Note that these functional relationships and error calculations do not represent “calibrations” of the $\delta^{18}\text{O}$ and Sr/Ca paleothermometers in the coral skeleton, as they are not specifically compared to the local condi-

Figure 3b. Paired $\delta^{18}\text{O}$ anomaly versus SST anomaly for approximately monthly resolution data. The $\delta^{18}\text{O}$ measurement precision error bars are as plotted. Solid line is a least squares regression line. Inferred spawning event data (open symbols) have not been used to compute the regression.

Figure 3c. Paired Sr/Ca anomaly versus SST anomaly for approximately monthly resolution data. Sr/Ca measurement precision error bars are as plotted. Solid line is a least squares regression line. Sr/Ca in samples that show anomalous inferred “spawning event” $\delta^{18}\text{O}$ are shown as solid points but are indistinguishable from all other Sr/Ca values.

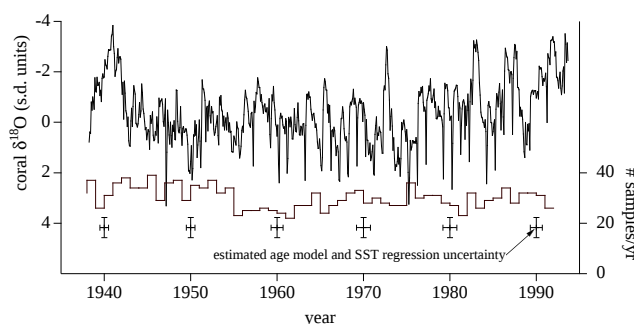


Figure 4. Age-modeled $\delta^{18}\text{O}$ results. Left axis shows $\delta^{18}\text{O}$ data, normalized by the 1951–1980 mean (-4.51‰) and standard deviation (0.26‰). Right axis shows number of samples analyzed per year. Also plotted are 10 year averages of the age model uncertainty; $\delta^{18}\text{O}$ amplitude uncertainty estimates are the RMS error of the regression in Figure 3b.

tions in which the coral grew. Rather, they measure the ability of the composition of our Kiritimati Atoll coral to predict the regional SST anomaly within the grid point surrounding Kiritimati in the central equatorial Pacific. This is consistent with our purpose of extracting large-scale modes of climate variability from the coral data.

The results of the singular spectrum analysis are shown in Figure 5. We choose a trajectory matrix dimension of 60 months; however, similar decompositions are obtained for trajectory matrix dimensions of 24 and 96 months. Similar results are also obtained using low-pass ($f \leq 0.2$ cycles/yr), band-pass ($0.2 < f \leq 0.5$ cycles/yr) and high-pass ($f > 0.5$ cycles/yr) filters in the frequency domain. The first 10 modes of the decomposition explain about 80% of the variance in the data series (Figure 5a); the remainder is interpreted here as noise. The first mode of the $\delta^{18}\text{O}$ data (Figure 5b) is

a close approximation of a 5-year running mean. This mode is suggestive of two extended warming periods or, alternatively, two periods of high ENSO frequency: the first, during the 1939–1942 period, and the second, during the post-1975 era. The annual signal is reconstructed from modes 6–10 (Figure 5d). We find the corresponding RCs are periodic with annual frequencies but have time-varying amplitudes. The annual reconstructed components explain about 15% of the total variance in the $\delta^{18}\text{O}$ data.

The SSA-extracted interannual reconstructed component of the coral $\delta^{18}\text{O}$ record is shown in Figure 5c. It is composed of two pairs of quasi-periodic oscillatory modes, each described by a pair of periodic RCs phase offset by 90° relative to each other. We interpret these RCs summed together as representing the low-frequency (≈ 5 –9 year period) and biennial modes of ENSO, respectively. The decomposition of ENSO into two characteristic frequency bands is similar to that of Barnett [1991], when he considered the interaction of the low-frequency and biennial components of ENSO in SST and sea level pressure fields.

We take the interannual component of the $\delta^{18}\text{O}$ data as a measure of the ENSO signal recorded at Kiritimati, and we compare it to local, regional, and global measures of the ENSO thermal oceanographic signature. Figure 6 shows the coherency ρ between the interannual component of the $\delta^{18}\text{O}$ data derived by the singular spectrum analysis, and the $5^\circ \times 5^\circ$ gridded SST anomaly [Kaplan *et al.*, 1998] centered on 157.5°W , 2.5°N , filtered using the same SSA parameters used to construct the Kiritimati $\delta^{18}\text{O}$ time series. Also plotted are 10 year averages of the squared coherence between the two time series. Shared coherence is maximal during decades with strong ENSO variability and varies between 50 and 85%; the overall squared coherence between the two time series is 70% ($\rho = -0.84$).

The linear correlation of annual averages of the $\delta^{18}\text{O}$ data and the global SST anomaly field from Kaplan *et al.* [1998] is plotted in Figure 7a. Plotted in Figure 7b is the same statistic, but for the correlation with annual averages of NINO3 SST anomaly from the Kaplan *et al.* [1998] analysis. Correlations of ≥ 0.3 are significant at the 90% level for a time series of ≥ 45 independent data points. Figure 7c shows the difference between these two correlation maps. The correlation patterns are very similar, especially for the Pacific basin, the equatorial Atlantic, and the North Indian Ocean regions.

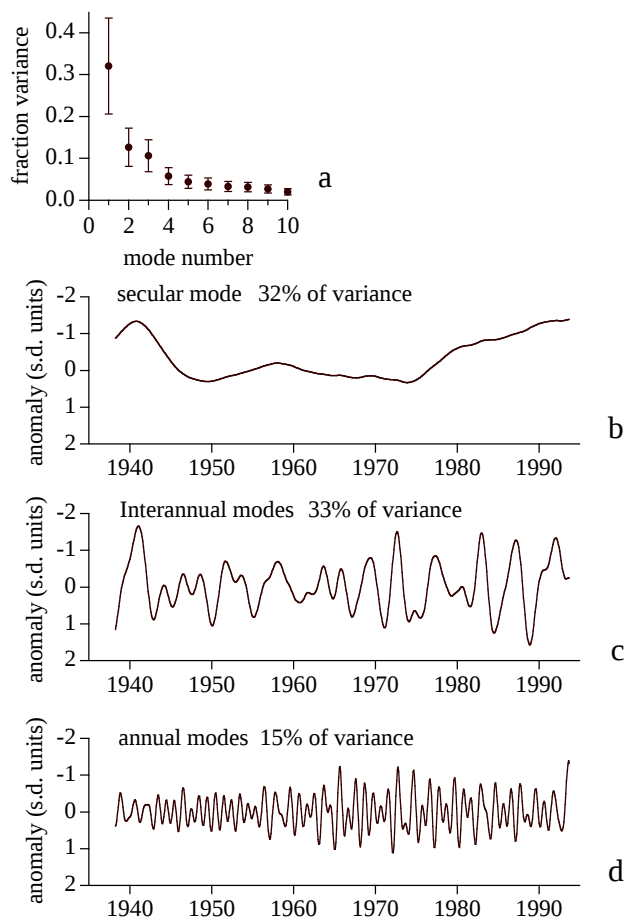


Figure 5. Frequency domain extraction of the climate signal, using the singular spectrum analysis technique. (a) Eigenvalues and estimated uncertainty [Vautard *et al.*, 1992] for the first 10 eigenvectors. Total variance explained by the first 10 modes is about 80%. Lag matrix dimension is 60 months for the analysis shown. (b) First reconstructed mode (RC) of the $\delta^{18}\text{O}$ data; 32% of total variance is explained. This mode closely resembles a 60 month running mean. (c) Second through fifth summed RCs; 33% of variance is explained. These summed modes describe SST anomalies associated with ENSO. (d) Sixth through tenth summed RCs; 15% of variance is explained. These modes have annual periodicity and are interpreted here as influenced by the coral’s annual reproductive processes (Figure 3; also see text for discussion).

4. Discussion

4.1. Observational Error in the Coral Data

Oxygen isotopic composition of coralline aragonite reflects three influences: sea surface temperature varia-

bility, seawater $\delta^{18}\text{O}$ variability, and biological (metabolic or reproductive) processes [Weber and Woodhead, 1972; Fairbanks and Dodge, 1979; Cole et al., 1993; Dunbar et al., 1994; McCulloch et al., 1994; McConnaughey, 1989; Gagan et al., 1994, 1996; Fairbanks et al., 1997]. Therefore extraction of a climate signal from coral data must be accompanied by an estimate of the observational error. In practice, this means separating any biological and non-SST climatic influences from the derived record of SST variability as well as acknowledging measurement reproducibility error.

Results from Figure 3 suggest that interannual $\delta^{18}\text{O}$ variability at Kiritimati is strongly a function of SST anomaly, with some seawater $\delta^{18}\text{O}$ influences. Comparison of the Sr/Ca and $\delta^{18}\text{O}$ data indicates a moderate seawater $\delta^{18}\text{O}$ anomaly toward the end of the 1982–1983 ENSO warm phase event (Figure 3a); this is consistent with surface salinity and seawater $\delta^{18}\text{O}$ anomalies that we observed at Kiritimati Atoll during the 1997 warm phase event (data not shown). The observations suggest that Kiritimati saw a significant seawater $\delta^{18}\text{O}$ anomaly due to heavy precipitation or eastward advection of low-salinity waters toward the middle of this event and some months subsequent to the arrival of the maximal SST anomaly, with amplitude approximately consistent with the amplitude of the seawater $\delta^{18}\text{O}$ anomaly inferred from the paired proxy data for late in the 1982–1983 ENSO warm phase event (Figure 3a). The importance of seawater $\delta^{18}\text{O}$ anomalies in the Kiritimati coral $\delta^{18}\text{O}$ record is probably greatest for large ENSO warm phase events, as during these local SST may exceed the convective threshold (28–28.5°C [Pan and Oort, 1983; Graham and Barnett, 1987]). For events of lesser amplitude, the $\delta^{18}\text{O}$ anomaly will be relatively smaller. Since seawater $\delta^{18}\text{O}$ anomaly covaries with SST anomaly at Kiritimati via enhanced local rainfall during ENSO events, the interannual coral $\delta^{18}\text{O}$ record synthesizes a convective atmospheric as well as thermal oceanographic ENSO signal. Hence the coral $\delta^{18}\text{O}$ may make a better climate monitor than either SST or seawater $\delta^{18}\text{O}$ anomaly alone.

Given these caveats, the observational error in SST anomaly estimated from the Sr/Ca and $\delta^{18}\text{O}$ coral data (Figures 3a–c) is roughly ± 0.4 – 0.5°C equivalent or about ± 0.4 standard deviation unit. Using the correlation between annual averages of the complete coral $\delta^{18}\text{O}$ time series and estimates of local SST anomaly, we obtain similar observational error estimates. We have assumed in this analysis that the RMS difference between the coral and SST data is due only to error in the coral estimate; it includes any uncertainty in the instrumental

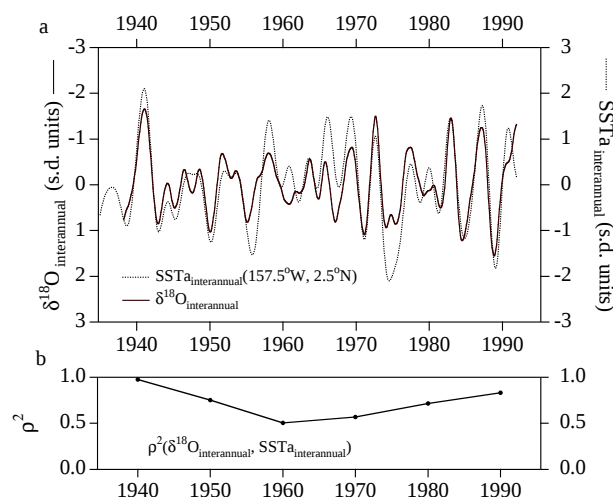


Figure 6. (a) Comparison of the interannual component of the $\delta^{18}\text{O}$ data with the SST anomaly from the analysis of Kaplan et al. [1998] for the (157.5°W, 2.5°N) $5^\circ \times 5^\circ$ grid box. SST anomaly data have been filtered using the SSA technique with the same lag window as for analysis of the coral $\delta^{18}\text{O}$ data shown in Figure 5c. Left axis shows summed interannual modes of $\delta^{18}\text{O}$ data (solid line, standard deviation units). Right axis shows SST anomaly (dashed line, standard deviation units). (b) Squared coherence ρ between interannual component of the $\delta^{18}\text{O}$ data and Kaplan SST anomaly, for 10 year nonoverlapping periods. The Kiritimati $\delta^{18}\text{O}$ data and Kaplan et al. [1998] SST anomaly share 50–80% of interannual coherency; overall coherence is 0.84.

SST data and in the assumption that coral $\delta^{18}\text{O}$ anomaly is solely a function of SST anomaly. Thus the observational error we estimate here should be considered an upper limit for the Kiritimati coral paleothermometer.

4.2. Interpretation of the Oxygen Isotope Time Series

SSA results (Figure 5) suggest that the climate signal-to-noise ratio in the Kiritimati $\delta^{18}\text{O}$ data is frequency dependent. The interannual component of the $\delta^{18}\text{O}$ record (Figure 5c), consisting of the summed second through fifth reconstructed components (Figure 5a), is most closely associated with ENSO-related SST variability and explains about one third of the variance in the data series. We focus on the nature and quality of the information contained in this component of the $\delta^{18}\text{O}$ data in section 4.3. In the remainder of this section, we discuss the origin of the other frequency components we observe in the $\delta^{18}\text{O}$ time series.

On the basis of intercomparison with analyzed historical SST data, annual and secular-band components of the $\delta^{18}\text{O}$ data series likely have large proportions of nonclimatic information (Figures 3 and 5). Figure 3a shows that there is an annually occurring pulse in the $\delta^{18}\text{O}$ data, which is not seen in the Sr/Ca data, seawater $\delta^{18}\text{O}$ data, buoy SST data, or gridded SST data. The annual $\delta^{18}\text{O}$ pulse extracted by SSA (Figure 5d) is contemporaneous with a similar annual feature in the $\delta^{13}\text{C}$ record (Figure 2), which has been suggested by *Gagan et al.* [1994, 1996] to be associated with the production of gametes during the mass spawning reproductive process [*Babcock et al.*, 1986]. On the basis of this evidence we interpret this feature as part of the coral's annual reproductive cycle, and do not interpret it as a climatically driven signal. While this feature overprints the climatic annual signal in the $\delta^{18}\text{O}$ record, we note that its counterpart in the $\delta^{13}\text{C}$ record enabled construction of a precise chronology for time series analysis of interannual, climatic $\delta^{18}\text{O}$ variability (Figures 2 and 4). However, the contemporaneous $\delta^{18}\text{O}$ pulse has not been observed in other coral isotopic records in which mass spawning effects are observed [*Gagan et al.*, 1994, 1996], perhaps owing to overwhelming seawater $\delta^{18}\text{O}$ anomalies introduced by riverine freshwater input to the barrier or nearshore reefs studied. In addition, the mass spawning studies cited above report a range of annual spawning dates for Australian corals (October/November and March) whose timing is not completely understood [*Gagan et al.*, 1996]. Mass spawning timing may be shifted from these periods for locales with different mean climates and seasonal cycles. Further study of the timing and mechanism of the isotopic signal attributed to mass spawning is needed to fully validate the mass spawning chronometer hypothesis.

The first SSA mode of the $\delta^{18}\text{O}$ data (Figure 5b) may be compared to secular variability in the SST anomaly reconstruction of *Kaplan et al.* [1998], extracted by performing SSA on SST anomaly data from the Kiritimati grid box with the same lag window. There is no similar secular component in the SST anomaly data. Available rain gauge data for the central equatorial Pacific is sparse, but unusually heavy rainfall in 1939–1941 and more frequent episodes of heavy rainfall associated with ENSO activity in the 1980s and 1990s are qualitatively consistent with lower coral $\delta^{18}\text{O}$ values during these periods in both this data set and in $\delta^{18}\text{O}$ data from a Tarawa Atoll coral [*Cole et al.*, 1993]. This hypothesis can be examined via additional application of paired Sr/Ca- $\delta^{18}\text{O}$ analyses on the 1939–1941 and 1987–1993 time intervals. However, as the secular variability

in the coral $\delta^{18}\text{O}$ record requires significant changes in the salinity of surface waters on decadal timescales, we cannot yet rule out nonclimatic coral $\delta^{18}\text{O}$ influences for this mode. The reproducibility of secular variability observed in additional coral colonies collected at Kiritimati or nearby atolls should be determined before this frequency component is interpreted as climatically driven.

4.3. Comparison of Proxy and Instrumental Indices of the Thermal Oceanographic ENSO Signal

The interannual component of the $\delta^{18}\text{O}$ time series captures much of the thermal oceanographic signature of the ENSO phenomenon, sharing 50–85% variance with local SST anomaly (Figure 6). Correlation is highest during periods of strong ENSO activity, such as the post-1971 period. Lower correlations are found for periods when ENSO frequency and SST anomaly amplitudes are small, and, consequently, the coral observational error is relatively large. The most prominent feature of the record is the strong, extended warm phase of 1939–1942, a particularly sparse observational period during which estimated SST analysis error approaches 0.35°C , about the observational error level of the $\delta^{18}\text{O}$ data [*Bottomley et al.*, 1990; *Kaplan et al.*, 1998]. Thus, the coral data provide an independent confirmation of the magnitude and duration of this singular event with comparable uncertainty. The high coherency suggests that the Kiritimati $\delta^{18}\text{O}$ data capture a large fraction of the variance associated with the thermal oceanographic signature of ENSO. Furthermore, the global pattern correlation is nearly identical to that derived by correlation of NINO3 with global SST anomaly (Figure 7). Strong positive correlations exist for much of the central and eastern equatorial Pacific, while the western equatorial Pacific and North Pacific subtropical gyres are regions of negative correlation. Weaker covariance is found in the North Indian Ocean and western equatorial Atlantic, but both of these regions also have weak correlation with NINO3. These observations suggest that between 10 and 60% of global SST anomaly variability may be captured in proxy records from only a few selected sites such as Kiritimati.

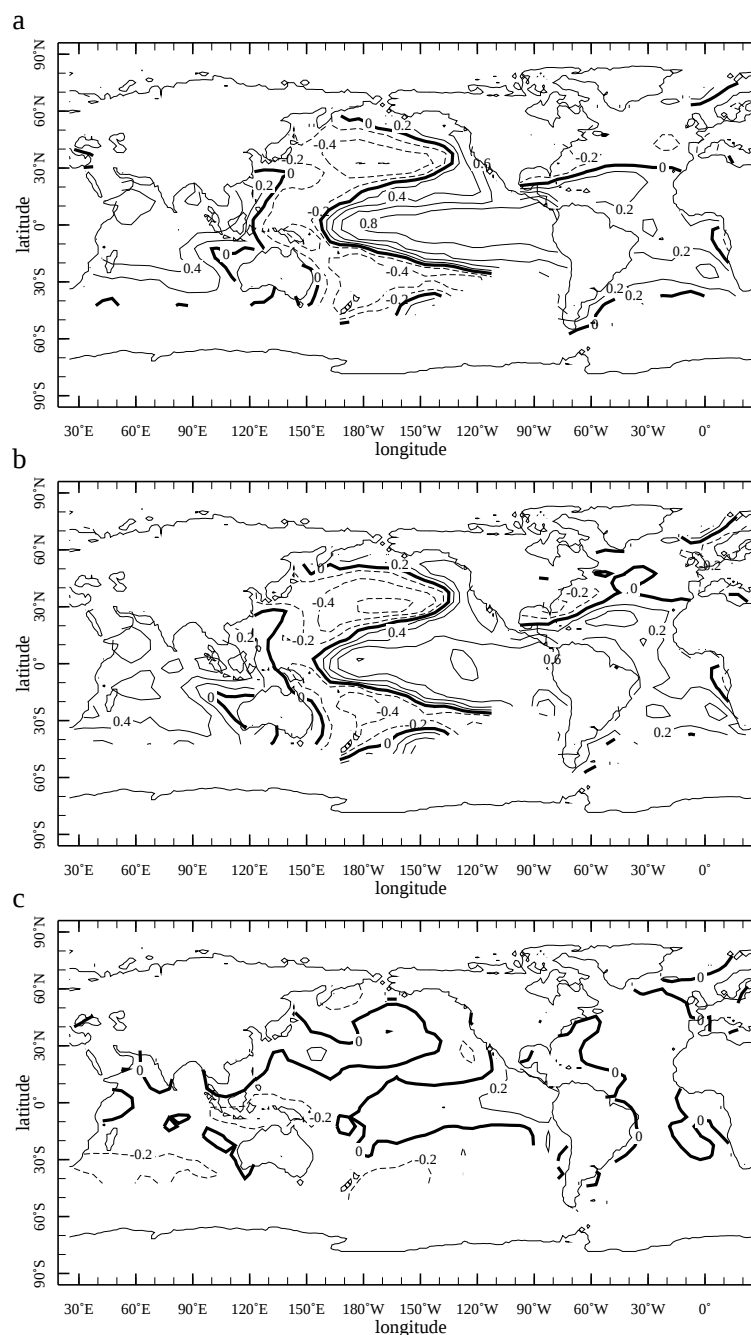


Figure 7. Comparison of the regional and global climate signal captured in the *Kaplan et al.* [1998] analysis NINO3 SST anomaly index with that captured in the Kiritimati $\delta^{18}\text{O}$ time series. We plot the correlation coefficient between the time series of annually averaged (April–March) SST anomaly at each ocean grid point and annual averages of each of the two climate indices. (a) Correlation map of NINO3 SST anomaly with global SST anomaly. (b) Correlation map of $\delta^{18}\text{O}$ anomaly with global SST anomaly. Map is multiplied by -1. (c) Difference between maps in Figures 7a and 7b. The thermal oceanographic signal of ENSO reconstructed from the Kiritimati $\delta^{18}\text{O}$ data is virtually indistinguishable from that reconstructed from NINO3.

5. Conclusions

In order to reconstruct SST anomaly in a critical region of the equatorial Pacific, we have extracted a measure of interannual sea surface temperature variability from proxy measurements on a Kiritimati Island coral. Paired stable isotopic and Sr/Ca data suggest that at Kiritimati, coralline $\delta^{18}\text{O}$ responds to interannual SST anomalies associated with both the warm and cold phases of ENSO, with minor seawater $\delta^{18}\text{O}$ influence. The observational error involved in reconstructing interannual SST anomalies from the Kiritimati $\delta^{18}\text{O}$ thermometer is equivalent to 0.4°C or about 30% of local SST variance. The interannual record of SST variability is overprinted by nonclimatic annual and interdecadal influences, which can be effectively separated from the climate signal using singular spectrum analysis or annual averaging.

Correlation of the interannual component of the $\delta^{18}\text{O}$ record with the global SST anomaly field is almost indistinguishable from correlation of the SST field with NINO3 SST anomaly. This result suggests that coral-based proxy measures of SST may be used to construct both local and nonlocal estimates of prehistorical SST variability, using high-quality SST anomaly reconstructions from a sparse network of well-located sites.

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- M. N. Evans, R. G. Fairbanks, and J. L. Rubenstone, Lamont-Doherty Earth Observatory, Columbia University, RT 9W, Palisades, NY 10964-8000. (email: mevans@ldeo.columbia.edu; fairbank@ldeo.columbia.edu; jimr@ldeo.columbia.edu) January 14, 1998; revised August 10, 1998; accepted November 2, 1998.
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