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A multicentury stable isotope record from a New Caledonia coral: Interannual and decadal sea surface temperature variability in the southwest Pacific since 1657 A.D.

Terrence M. Quinn,¹ Thomas J. Crowley,² Frederick W. Taylor,³ Christian Henin,⁴ Pascale Joannot,⁵ and Yvan Join⁴

Abstract. A 335 year stable isotope record from a New Caledonia coral (22°S, 166°E) helps fill a large gap in historical climate reconstructions. Although the long-term coral $\delta^{18}\text{O}$ -based sea surface temperature (SST) trend is one of warming, there are notable decadal fluctuations, especially in the early 18th and early 19th centuries. Mean annual SSTs between 1658 and 1900 are estimated to be $\sim 0.3^\circ\text{C}$ lower than the 20th century average, with interdecadal excursions of $0.5^\circ\text{--}0.8^\circ\text{C}$. Time series analyses of the coral isotope record reveals significant concentrations of variance in the El Niño band; an interdecadal spectral peak is present, but its robustness requires additional statistical evaluation. A secular but irregular decrease in coral $\delta^{13}\text{C}$ values begins in the mid-1800s and may reflect the anthropogenic perturbation of the carbon reservoir. These and other results indicate that the New Caledonia coral isotope record is a valuable source of information on southwest Pacific climate history.

1. Introduction

Geochemical time series derived from coral skeletons are increasingly recognized as promising monitors of climate change in tropical oceans [e.g., Dunbar and Cole, 1993]. Interpretation of coralline isotopic time series usually focuses on defining the relative contribution that environmental variations in sea surface temperature (SST) and sea surface salinity (SSS) have on these records. In localities where the effect of one of these environmental variables dominates the other (e.g., SST in Galápagos) [Dunbar *et al.*, 1994, Wellington *et al.*, 1996], SST in Swain Reef, Great Barrier Reef (GBR) [Druffel and Griffin, 1993], SSS in Tarawa [Cole and Fairbanks, 1990], unambiguous records of a particular climate index may be reconstructed from stable oxygen time series alone. In localities where variations in both salinity and temperature influence the record (e.g., Vanuatu, [Quinn *et al.*, 1993, 1996a]; Pandora Reef, GBR [Gagan *et al.*, 1994]; Portland Roads Reef, GBR [Cole, 1996]) multiple proxies (e.g., combined Sr/Ca and $\delta^{18}\text{O}$) can be used to partition the effects of the different environmental factors. A variety of factors including physiological effects, growth rate, and the $\delta^{13}\text{C}$ of seawater influence the $\delta^{13}\text{C}$ of the coral skeleton [e.g., Fairbanks and Dodge, 1979; Swart, 1983; Swart *et al.*, 1996]. Hence precise environmental interpretation of coral $\delta^{13}\text{C}$ has remained elusive.

In this paper we present isotopic results from a new 335 year coral record from New Caledonia in the southwest Pacific. In addition to adding to the relatively small number of published long coral records, the New Caledonia record is of note both for being sited in a region that is relatively undersampled from the perspective of long climate series and because there is a particularly fine long environmental data set against which the coral can be calibrated. The objective of the study is to: (1) reconstruct the history of environmental variations at the sea surface, (2) determine the frequency bands in which variance in the record is concentrated, and (3) compare isotopic patterns observed at New Caledonia with other Pacific coral and climate records.

2. Study Area

New Caledonia is located in the eastern sector of the Coral Sea (Figure 1). The Loyalty Current circulates waters from the northwest to the southeast along the northeast coast and the South Tropical Current circulates waters generally from the southeast to the northwest along the southwestern coast [Rougerie, 1986]. The climatology of the southwest Pacific in the vicinity of New Caledonia is influenced by the annual variations in the relative positions of the subtropical anticyclonic belt south of New Caledonia and the South Pacific Convergence Zone (SPCZ) to the north [Morliere and Rebert, 1986]. Rainfall and air temperatures reach a maximum between December and April because of the southward displacement of the SPCZ. During the rest of the year, rainfall and air temperatures decrease in response to the northward migration of the SPCZ.

The warm phase of the El Niño-Southern Oscillation (ENSO) results in SST cooling and decreased rainfall at New Caledonia in contrast to the eastern equatorial Pacific, which experiences SST warming and increased rainfall [e.g., Rasmusson and Carpenter, 1982]. Mean monthly rainfall totals decrease by 22% for the 15 month period after the beginning of an ENSO year at New Caledonia [Morliere and Rebert, 1986]; whereas

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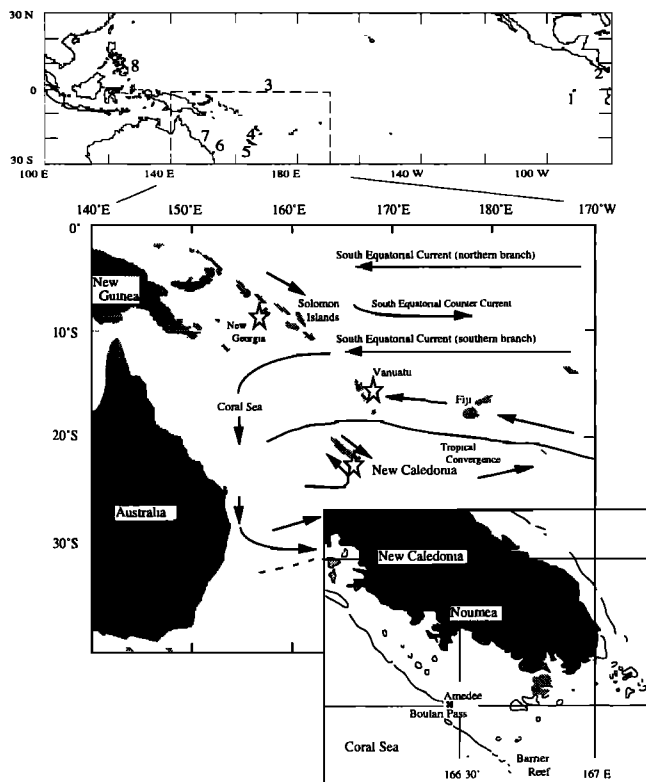


Figure 1. Map of the equatorial Pacific Ocean with an expanded view of the Coral Sea and South Pacific sector (modified from Quinn *et al.*, 1996b). The numbers denote locations of other Pacific coral sites discussed in the text: 1, Galápagos; 2, Secas Island, Panama; 3, Tarawa Atoll, Kiribati; 4, Espiritu Santo Island, Vanuatu; 5, Amédée Island, New Caledonia; 6, Abraham Reef, Great Barrier Reef (GBR); 7, GBR; and 8, Cebu, Philippines. Our coral sample came from just offshore of Amédée Island (22°29'S, 166°27'E), which is located 20 km due south of Noumea, New Caledonia, slightly landward of the barrier reef and within an ocean pass (Boulari Pass).

the average peak SST anomaly associated with ENSO is 0.2°C [Rasmusson and Carpenter, 1982]. Overall there is a weak (≤ -0.3) negative correlation for the New Caledonia region between various indices of ENSO activity (e.g., Southern Oscillation index (SOI) and rainfall [Donguy and Henin, 1980] and SST (e.g., K. Wolter, unpublished data, 1992).

3. Methods

3.1. Physical Sampling of Corals

The *Porites lutea* coral used in this study was collected in the vicinity of the Amédée Lighthouse (22° 29' S, 166° 27' E), which is located 20 km south of Noumea, New Caledonia, slightly landward of the barrier reef and within Boulari Pass (Figure 1). This locality was chosen because corals living offshore of Amédée Island are well bathed by open-ocean marine waters and because daily SST and SSS measurements have been made there since 1967 by the French research group Institut Français de Recherché Scientifique pour le Développement en

Coopération (ORSTOM). In addition, spot measurements of the oxygen isotopic composition of seawater at this locality have been made, yielding an average value of 0.52‰ for $\delta^{18}\text{O}$ (SMOW) [Beck *et al.*, 1992; J. Recy, unpublished data, 1995]. A water sampling program by ORSTOM scientists began in 1996, and these samples will be analyzed for their elemental and isotopic composition in the future.

We drilled a ~3.45 m long, 9 cm diameter core down the vertical axis of maximum growth of the *Porites lutea* coral head. This sample was collected alive in ~3 m water depth, less than a few hundred meters from the location of the daily SST and SSS measurements. The coral was slabbed to a thickness of 5 mm. X radiographs of the coral slab were taken under exposure conditions of 55 kV and 3 mA, with an exposure time of 20 s. X radiographs revealed highly regular and well-developed annual density bands (Figure 2).

Aragonite powder was extracted from the coral skeleton in a continuous fashion via a computer-driven automated sampling device [Quinn *et al.*, 1996b]. We sampled the coral at two levels of sampling density. The upper 40 years of the coral were sampled at 12 per density-band couplet (dbc) for two reasons. First, we wanted to compare the $\delta^{18}\text{O}$ record against in situ environmental measurements in the interval in which they overlap. Second, we wanted to objectively determine the minimum sampling density necessary for retrieving information on the annual cycle and lower-frequency climate variability. Quinn *et al.* [1996b] demonstrated that 4 dbc⁻¹ sampling represents an adequate sampling interval to retrieve a reproducible estimate of the annual mean, which is a standard primary target of climatological investigations. Thus the rest of the core was sampled at that density. As a further test of this sampling strategy we resampled the interval 1900-1910 at 12 dbc⁻¹. The average $\delta^{18}\text{O}$ value for the 4 dbc⁻¹ sampling and the 12/dbc sampling for the first decade of the 20th century was -4.12‰ and -4.16‰, respectively. Furthermore, the amplitude of the $\delta^{18}\text{O}$ signal in the 4 dbc⁻¹ sampling and the 12 dbc⁻¹ sampling was virtually identical.

3.2. Isotope Analysis

Stable isotopic analyses were done at the University of Michigan. Prior to isotopic analysis, powdered coralline aragonite samples were vacuum roasted for 1 hour at 200°C. Samples were reacted with anhydrous phosphoric acid at 75°C in individual reaction vessels of a CarboKiel carbonate-extraction system coupled to the inlet of a modern analogue technique (MAT) 251 mass spectrometer. Precision ($\pm 2\sigma$) was monitored by daily analyses of a powdered calcite standard (National Bureau of Standards, NBS 20) and was better than 0.08‰ for both oxygen and carbon. The average standard deviation of 46 replicate analyses of coral samples is 0.10‰ for oxygen and 0.07‰ for carbon. Values are reported in standard δ notation relative to Vienna Pee Dee belemnite (VPDB) after correction for the ¹⁷O contribution.

3.3. Time Series Analysis

Several different spectral analysis packages were used in the analysis of the coral and climate time series. The Blackman-Tukey method was applied to a variety of data sets discussed in the text using software packages including ARAND [Imbrie *et*

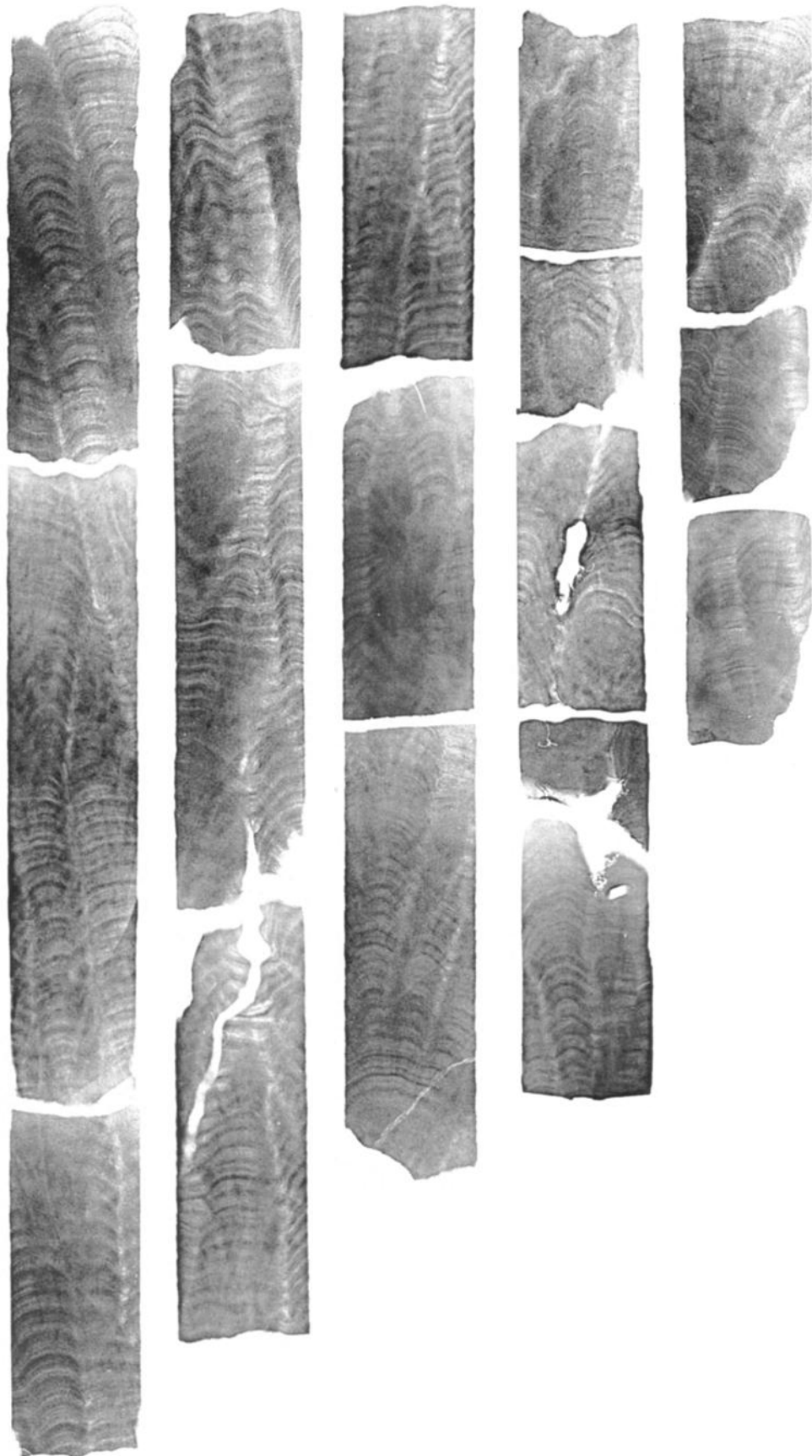


Figure 2. X radiograph of *Porites lutea*. Note the well-developed high- and low-density band couplets and the overall goodness of fit between the individual coral slabs.

al., 1989] and AnalySeries [Paillard *et al.*, 1996]. In Blackman-Tukey spectral analysis (BTSA) the autocovariance of the data is first computed, then a window is applied (in this case a Tukey window), and the Fourier transform of the covariance functions are used to compute a power spectrum. Singular Spectrum Analysis (SSA) has previously been applied to coral time series [Dunbar *et al.*, 1994; Linsley *et al.*, 1994]. This is a fully nonparametric analysis technique based on classical empirical orthogonal function (EOF) analysis applied in the time domain. SSA uses M -lagged copies of centered time series to calculate eigenvalues and eigenvectors of the covariance matrix. A detailed description of this technique and its paleoclimate application is given by Vautard and Ghil [1989] and Vautard *et al.* [1992]. However, standard SSA analysis may not take adequate account of low-frequency red noise variance due to stochastic forcing in the ocean-atmosphere system [Allen and Smith, 1996; cf. Hasselmann, 1976]. In particular, the observed spectrum of a single realization of a climate variable can differ substantially from the true underlying spectrum of the system which generates it. Careful statistical analysis is required to distinguish features which could have occurred by chance from those which indicate unambiguous departures in the spectrum of the underlying system from, for example, a red-noise background. With the assistance of M. Allen we examined the Monte Carlo SSA (MC-SSA) responses in the presence of such "colored" noise (see Allen and Smith [1996] for details).

We also analyzed the coral and climate time series using the multitaper method (MTM). MTM computes a time series as sums of sinusoids and calculates the corresponding amplitudes by a least squares procedure. Significance tests similar to those for regression analyses can be made on MTM results because of the least squares approach. MTM achieves high resolution by applying multiple tapers (i.e., optimal windowing functions in the time domain) to the time series. Tapers in MTM are discrete prolate spheroidal sequences (DPSS), which are a set of basis functions that are closely related to the Fourier transform of the function, with ones for the length of the series and zeroes elsewhere [Dettinger *et al.*, 1995]. We used an implementation of MTM that is contained in the SSA-MTM toolkit, a set of programs developed and distributed by the Atmospheric Science Department at University of California, Los Angeles (www.atmos.ucla.edu/tcd/ssa). This MTM program was developed by M. Mann and is a version of the nonparametric MTM method developed by Thomson [1982, 1990a, b; cf. Yiou *et al.*, 1991], and readily produces confidence intervals [Mann and Lees, 1996]. Details of the technique are available in the SSA-MTM toolkit documentation.

3.4. Chronostratigraphy

Conversion from the depth domain to the time domain in coral-based climate studies is commonly done by assigning a calendar year to each density-band couplet. In the absence of unambiguous density banding, or as a refinement to an initial age model based on density banding, the annual cycle in a measured geochemical variable can also prove useful in such a conversion. Counting of density-band couplets, however, remains a conventional means by which time is demarcated in

coral-based climate studies. In this study, stable isotopic values were converted from the depth domain to the time domain in two steps. For the upper part of the record (1992-1952) a calendar year was first assigned to each density-band couplet and hence to every 12 samples assuming linear subannual skeletal extension. Second, systematic variations in oxygen isotopic composition were correlated with monthly SST values by using the AnalySeries program [Paillard *et al.*, 1996] to match the peaks and troughs in the two time series.

For the lower part of the record (1657-1951), simple band counting was used to extend the chronology back to the beginning of the record, which we estimate to be 1657. Although we have no absolute dates on this coral, Crowley *et al.* [1997] recently demonstrated that some of the distinctive cool excursions in the mean annual record coincide within 1 year of known volcanic eruptions as determined from the volcano chronology of Simkin and Siebert [1994] and various ice core records. For example, some of the most characteristic cooling events are closely associated with the eruptions of Agung (1963), Krakatau (1883), Cossequina (1835), a pulse of eruptions from 1812-1822, and an unknown eruption in 1808 [Dai *et al.*, 1991]. A notable feature of the correspondence of these isotopic trends with the eruptions is that virtually no retuning of the original band counting was needed to obtain the agreement. The only slight adjustment involved part of the 20th century record in which a reexamination of the X radiograph and stable isotope data indicated that a 1 year adjustment needed to be made between the overlapping region of two separate drill paths.

Subsequent to the publication of Crowley *et al.* [1997] an error in the sequencing of the individual core segments was identified and a revised chronology had to be developed. This revision only affects parts of the 18th century section of the core. The exact change involved transposing the core segments originally defined as being from 1702.25 to 1721.25 years and 1761.75 to 1780.75. The revised chronology does not undermine the main conclusions in Crowley *et al.* [1997] about the relation between distinct interannual cooling events and volcanism, as that conclusion depended most strongly on the post-1800 section of the record (information from only one ice core was used for the comparison of volcanism and cooling in the 18th century). However the new chronology does call into question the significance of the 1752 and 1754 eruptions [Crowley *et al.* 1997] as to their climate effect.

4. Results

The results of our analysis of the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ variations in the New Caledonia coral are presented by first calibrating the coral $\delta^{18}\text{O}$ with available SST records (Table 1), then comparing the coral $\delta^{18}\text{O}$ record with observed SST fluctuations (Figures 3 and 4a) and the Southern Oscillation (Figure 4b), and finally discussing trends in the growth rate (Figure 5) and stable isotope records (Figures 6, 7, 8).

4.1. Calibration of the SST Record

Calibration between the instrument and proxy records was evaluated using two regression techniques (Table 1): standard ordinary least squares (OLS) and reduced major axis regression

Table 1. Linear, Zero-Lag, Least Square Correlation Coefficients r Between Coral $\delta^{18}\text{O}$ and Sea Surface Temperature (SST) Data at Various Sampling Densities and Time Intervals Using Ordinary Least Squares (OLS) and Reduced Major Axis (RMA) Regression Techniques

Interval (Resolution ^a)	OLS Equation ^b	RMA Equation ^b	r (r^2)
ORSTOM SST^c			
1991-1967 (mo)	SST = $-5.81(\delta^{18}\text{O}) - 2.39$	SST = $-6.64(\delta^{18}\text{O}) - 6.04$	-0.88 (0.77)
1991-1967 (qtr)	SST = $-5.53(\delta^{18}\text{O}) - 1.06$	SST = $-5.46(\delta^{18}\text{O}) - 0.85$	-0.85 (0.72)
1991-1967 (ma)	SST = $-2.67(\delta^{18}\text{O}) + 11.67$	SST = $-4.64(\delta^{18}\text{O}) + 2.89$	-0.57 (0.33)
GISST2 SST^d			
1991-1967 (mo)	SST = $-5.34(\delta^{18}\text{O}) - 0.93$	SST = $-6.05(\delta^{18}\text{O}) - 2.25$	-0.88 (0.78)
1991-1952 (mo)	SST = $-5.38(\delta^{18}\text{O}) - 0.62$	SST = $-6.17(\delta^{18}\text{O}) - 2.89$	-0.87 (0.76)
1991-1967 (qtr)	SST = $-4.75(\delta^{18}\text{O}) + 3.52$	SST = $-5.63(\delta^{18}\text{O}) - 0.39$	-0.85 (0.72)
1991-1952 (qtr)	SST = $-4.86(\delta^{18}\text{O}) + 2.96$	SST = $-5.73(\delta^{18}\text{O}) - 0.90$	-0.85 (0.72)
1991-1903 (qtr)	SST = $-3.94(\delta^{18}\text{O}) + 7.31$	SST = $-5.14(\delta^{18}\text{O}) - 2.05$	-0.77 (0.59)
1991-1967 (ma)	SST = $-2.15(\delta^{18}\text{O}) + 15.14$	SST = $-3.92(\delta^{18}\text{O}) + 7.23$	-0.55 (0.30)
1991-1952 (ma)	SST = $-1.72(\delta^{18}\text{O}) + 16.98$	SST = $-3.33(\delta^{18}\text{O}) + 9.80$	-0.52 (0.27)
1991-1903 (ma)	SST = $-1.25(\delta^{18}\text{O}) + 19.03$	SST = $-2.23(\delta^{18}\text{O}) + 14.77$	-0.56 (0.32)
1991-1903 (ma ^e)	SST = $-1.37(\delta^{18}\text{O}) + 18.53$	SST = $-1.97(\delta^{18}\text{O}) + 15.87$	-0.69 (0.48)
Best-Fit Equations			
1991-1952 (ma)	SST = $-1.72(\delta^{18}\text{O}) + 16.98$		-0.53 (0.28)
1991-1903 (ma ^e)		SST = $-1.97(\delta^{18}\text{O}) + 15.87$	-0.69 (0.48)

^a resolution: mo, monthly; qtr, quarterly/seasonal; ma, mean annual.

^b Independent variable (X) is coral $\delta^{18}\text{O}$, and the dependent variable (y) is SST.

^c SST data are from Institut Français de Recherche Scientifique pour le Développement en Coopération (ORSTROM) [C. Henin, unpublished data, 1996].

^d SST data from Global Sea Ice and Sea Surface Temperature (GISST2) gridded dataset [Parker et al., 1995].

^e Three year moving average window.

(RMA). OLS regression minimizes the deviations of the observations from the line in one direction only, either x (the independent variable) or y (the dependent variable). In contrast, RMA regression fits a line to a collection of bivariate observations that minimizes the deviations of the observations from the line in both the x and y directions simultaneously. A more complete discussion of the relative attributes of OLS and RMA is given by Davis [1986] and Sokal [1995]. We explored both regression techniques because RMA regression has been suggested as a more appropriate technique to use when comparing coral proxy and instrumental records because of the inherent errors in both records [Shen and Dunbar, 1995]. Utilizing either approach, correlation coefficients (r) between monthly and quarterly measurements of coral $\delta^{18}\text{O}$ and ORSTOM SST (1991-1967) are ≥ 0.85 ($p < 0.01$).

Although for the purpose of calibration $\delta^{18}\text{O}$ is sometimes plotted as a function of SST, in our regression we have chosen SST as the dependent variable for several reasons: (1) that is the way the equation is used to estimate paleotemperatures; (2) the approach has been utilized before, for the original calibration of molluscan $\delta^{18}\text{O}$ variations [Epstein et al., 1953], for foraminifera [Erez and Luz, 1983], and for corals [Weber and Woodhead, 1972]; (3) the larger the slope, the greater the separation of signal from noise; and (4) RMA is most appropriate for use when the errors of each index are approximately the same; that situation does not necessarily apply to the $\delta^{18}\text{O}$ /SST comparison.

The temperature/ $\delta^{18}\text{O}$ calibration determined for monthly observations is $-5.81^\circ\text{C}/\text{‰}$ using OLS regression and $-6.64^\circ\text{C}/\text{‰}$ using RMA regression. These numbers are different than the value of $-5.28^\circ\text{C}/\text{‰}$ published by Quinn et al. [1996b] and in part reflect the fact that a different interpolation scheme was used to fill in missing SST data for the ORSTOM data (SSTs were patched in from the gridded $1^\circ \times 1^\circ$ observations of Parker et al. [1995] after adjustment for the difference in mean SST values of the two data sets). Our SST- $\delta^{18}\text{O}$ relationship is also different than the $-3.31^\circ\text{C}/\text{‰}$ reported by Weber and Woodhead [1972] for *Porites lutea* and the widely recognized de facto standard values of $-4.5^\circ\text{C}/\text{‰}$ (0.22°C) first determined by Epstein et al. [1953] for calcite and $-4.2^\circ\text{C}/\text{‰}$ obtained by Grossman and Ku [1986] for aragonite. However our values are close to those determined by Gagan et al. [1994] for Great Barrier Reef *Porites lutea*.

As discussed by T. J. Crowley et al. [manuscript in preparation, 1998], comparison between a local SST record (e.g., ORSTOM, 1991-1967) and coral $\delta^{18}\text{O}$ time series is a necessary first step in the calibration of corals. Perhaps more importantly from a paleoclimate perspective, a comparison between the local coral $\delta^{18}\text{O}$ time series should also be made with the nearest grid box from a global SST data set. Even though these global data sets represent a much larger region than a coral site, comparison with the coral record is necessary because one of the aims of coral studies is to draw inferences about large-scale changes in the regional climate system, and

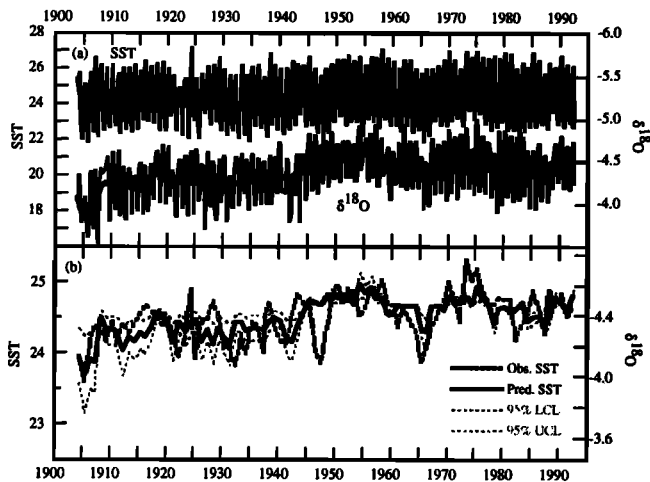


Figure 3. (a) A comparison of the seasonal variations in coral $\delta^{18}O$ and sea surface temperature (SST) from New Caledonia for the 20th century. A Stineman smoothing function (thick solid line), which applies a geometric weight to a point and $\pm 10\%$ of the data range, has been applied to the seasonal data to emphasize the low-frequency trends in these records. (b) The estimated mean annual SST changes for New Caledonia (thick solid line) based on the mean annual 1952-1991 regression equation (see Table 1 and text for discussion). The thin dashed lines represent the 95% error limits of the predicted SST values. The thick dashed line represents mean annual temperatures from observations [Parker *et al.*, 1995].

direct correlations with the large-scale data set represent the only means to test the limits of drawing such inferences. Calibration against gridded data sets also affords the opportunity to validate temperature regressions against an independent data set from the early part of the century (T. J. Crowley *et al.*, manuscript in preparation, 1998).

We therefore extracted monthly SST data from a $1^\circ \times 1^\circ$ grid centered at $22^\circ S$ and $166^\circ E$ from the Global Sea Ice and Sea Surface Temperature (GISST2) data set (1903-1994), distributed by the United Kingdom Meteorological Office [Parker *et al.*, 1995]. Monthly coral $\delta^{18}O$ and gridded SST are correlated equally as well as for the local ORSTOM data set ($r = -0.88$ and $p < 0.01$), and the temperature- $\delta^{18}O$ calibration is comparable to that determined with the 25 year comparison between monthly coral $\delta^{18}O$ and the ORSTOM SST data (Table 1). To extend the comparison between coral $\delta^{18}O$ and SST over the 20th century the GISST2 SST data was translated into a quarterly SST time series by averaging three sequential monthly values. We compared the quarterly relationship between coral $\delta^{18}O$ and SST from 1967 to 1991, 1952 to 1991, and 1903 to 1991 and determined that the correlation coefficient varied from -0.85, -0.85, and -0.77, respectively (Table 1).

The predicted SST record, derived from our regression analysis of coral $\delta^{18}O$ and SST for the period 1952-1991, was then compared (Figure 3) with the observed GISST2 SST record over the interval 1903-1950. The predicted mean annual SSTs for the verification interval, using either the subannual OLS or RMA regression equations from the calibration interval, were not a good fit to the observed mean annual SSTs over the

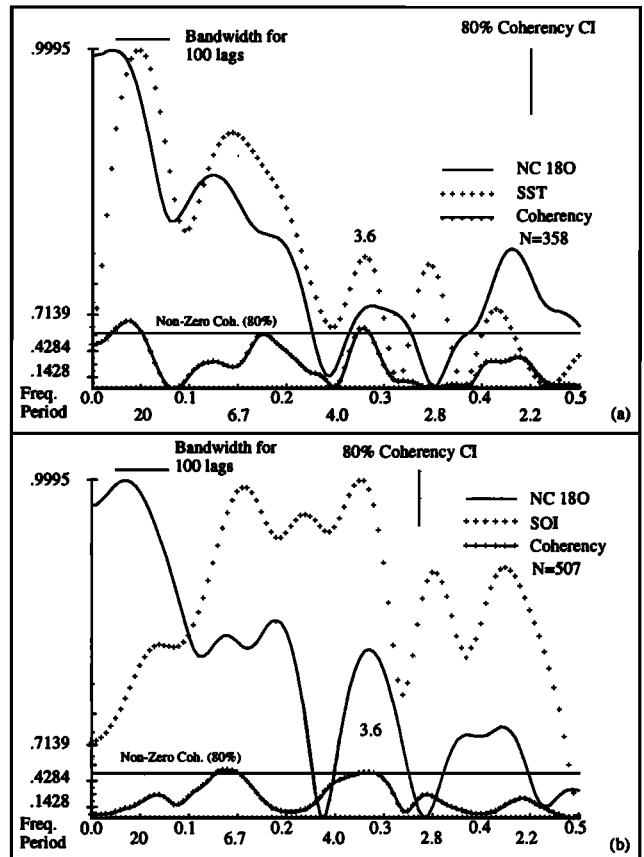


Figure 4. The results of Blackman-Tukey cross-spectral analysis (a) between seasonal variations in coral $\delta^{18}O$ and SST [Parker *et al.*, 1995] for the 20th century and (b) between seasonal variations in coral $\delta^{18}O$ and the Southern Oscillation index (SOI) for the period 1867-1991 [Ropelewski and Jones, 1987; Allan *et al.*, 1991]. The variance spectra is plotted as the normalized log of the spectral density. Spectral density equals variance divided by frequency of the coral $\delta^{18}O$ record (solid line) and of the SST record (pluses in Figure 4a) and of the SOI record (pluses in Figure 4b). Coherency, the correlation coefficient as a function of frequency, is shown as solid line with plus symbols. The most robust spectral peak occurs at 3.6 years in both analyses.

verification interval (T. J. Crowley *et al.*, manuscript in preparation, 1998). This poor fit occurred despite the high correlation between monthly and quarterly coral $\delta^{18}O$ and SST data (Table 1) over the calibration interval.

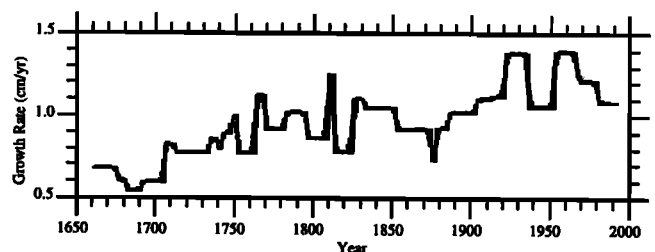


Figure 5. Variations in growth rate of the *P. lutea* coral from New Caledonia (thin line). The thick, dashed line is the growth record smoothed with a 5 year moving average window.

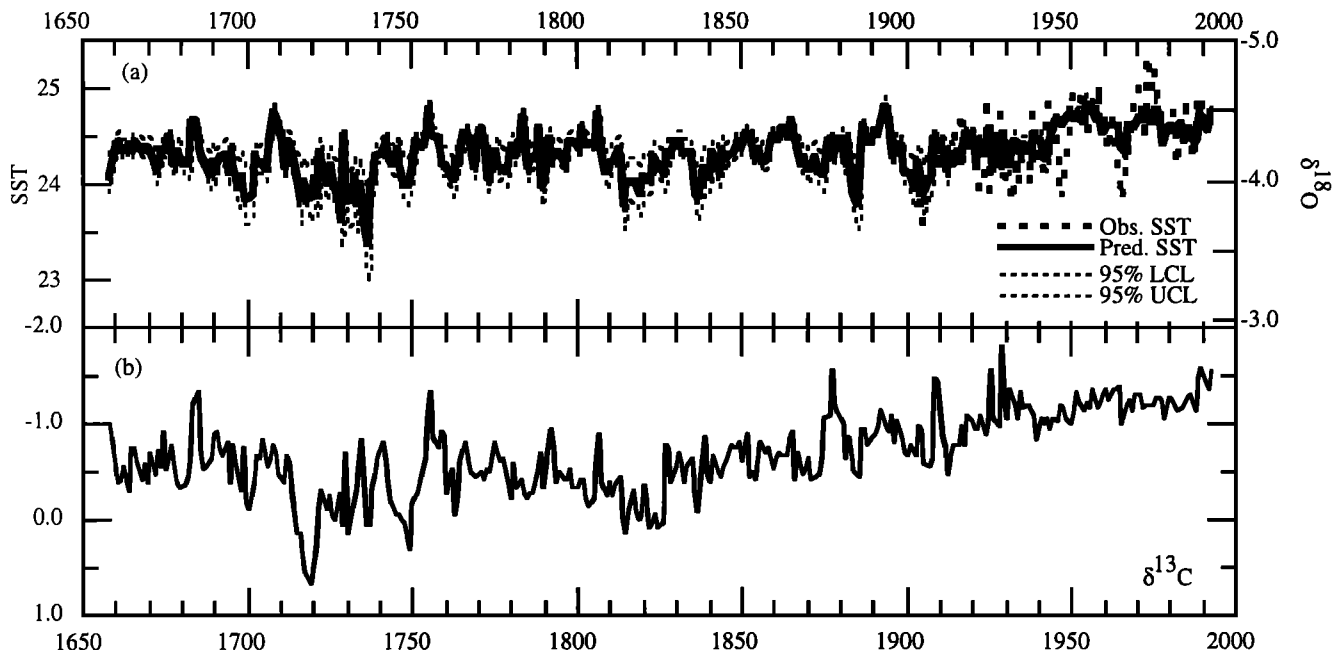


Figure 6. The 335 year record of variations in (a) mean annual coral $\delta^{18}\text{O}$ and (b) $\delta^{13}\text{C}$. Figure 6a also shows the 335 year record of predicted SST variations at New Caledonia, as inferred from the 1952–1991 $\delta^{18}\text{O}$ -SST calibration equation (see text), the 20th century record of SST observations [Parker *et al.*, 1995], and the 95% confidence levels of the SST estimates.

Despite only a moderate correlation between mean annual coral $\delta^{18}\text{O}$ and SST data over the calibration interval (Table 1), the predicted mean annual SST using the mean annual OLS equation and the observed mean annual SST match well over the verification interval (Figure 3b), with the 20th century trends and variance for estimated and observed SSTs being almost identical (T. J. Crowley *et al.*, manuscript in preparation, 1998). This calibration-verification exercise indicates that despite some theoretical reasons as to why RMA could be a more appropriate technique to use, the OLS technique provides a better estimate of SST, at least for this application. We have therefore used the OLS technique in our SST reconstruction because it is the method that works best as assessed via an independent validation exercise.

As discussed more fully by T. J. Crowley *et al.* (manuscript in preparation, 1998), the misfit between the observed and predicted SSTs in the pre-1950 period could be considered unsettling given the combination of a high-quality local environmental record of SST and SSS variations and the monthly resolved coral $\delta^{18}\text{O}$ record. Although some of this misfit may be the result of some “site-specific” effects for the New Caledonia coral, calibration differences and difficulties also occur in other coral time series. These observations lead us to conclude that the calibration difficulties we encountered in this study go beyond this particular coral isotope record and site-specific effects at New Caledonia. T. J. Crowley *et al.* (manuscript in preparation, 1998) explore further the problems and implications of predicting mean annual SST using regression equations developed from comparisons between subannual proxy and instrumental records.

Inspection of predicted coral temperatures indicates that the early 1940s portion of the coral record exhibits a step

transition from higher to lower $\delta^{18}\text{O}$ values. Although this transition has been previously identified in instrumental records [Zhang *et al.*, 1997], when compared to local air (not shown) and water temperatures (Figure 3), the New Caledonia coral transition magnifies the transition. This isotopic transition is not coincident with a sampling anomaly (e.g., core break or change in drilling location) nor is it associated with a change in growth rate (Figure 5). We speculate that the magnitude of the transition may be related to a decadal-scale shift in salinity or rainfall, which resulted in an amplification of the coral $\delta^{18}\text{O}$ -temperature signal. T. J. Crowley *et al.*, (manuscript in preparation, 1998) have suggested that the interannual changes in salinity, which may be proportionately larger (with respect to temperature) than seasonal changes [Quinn *et al.*, 1996b], could be responsible in part for some of the above discussed degradation of coral/SST correlations between monthly and mean annual regressions.

Further inspection of predicted mean annual temperatures with local air and water temperatures indicates that although the mean annual regression captures the 20th century trends, there are two notable short intervals (middle 1920s and the late 1940s) when the instrumental record indicates SST cooling and the coral $\delta^{18}\text{O}$ record shows no significant changes (Figure 3b). We explored the source of this misfit by examining the seasonal coral $\delta^{18}\text{O}$, SST, and air temperature record from Noumea, New Caledonia, for the 20th century. The SST cooling in the middle 1920s and the late 1940s is also not observed in the air temperature record. The discrepancy between the GISST2 SST estimates for the late 1940s and the coral and air temperature records may reflect spatial sampling biases in the early part of the century and the fact that the GISST2 data set eliminates the -0.5°C correction for canvas

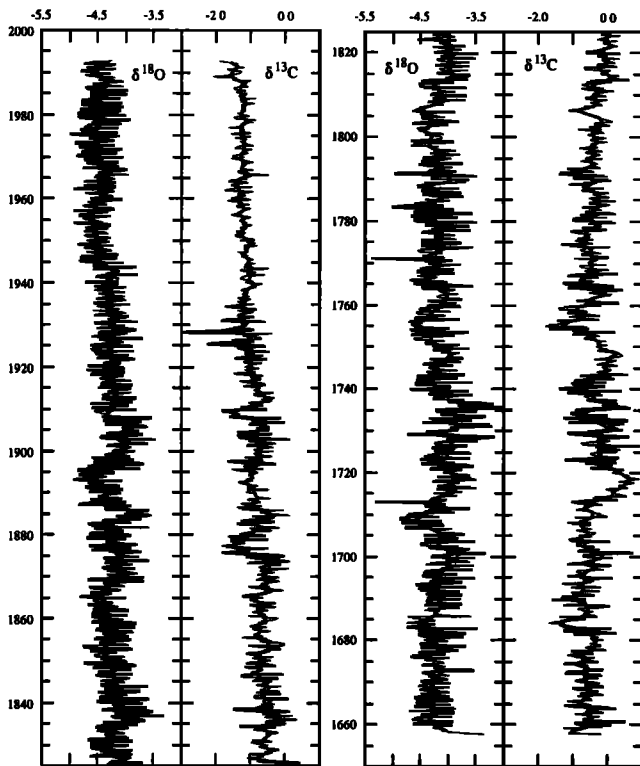


Figure 7. Seasonal (thin line) and mean annual (thick line) coral stable isotope values for the New Caledonia record. Note that the annual cycle is well represented throughout the seasonal $\delta^{18}\text{O}$ record and that there is a sharp reduction in the amplitude of seasonal $\delta^{13}\text{C}$ during the last 50 years at New Caledonia.

buckets around 1945 (P. Jones, personal communication, 1997). Yet some bucket measurements may have still been taken, especially in more remote regions such as the South Pacific.

Cross-spectral analysis of the seasonal coral $\delta^{18}\text{O}$ and GISST2 data was performed to further investigate the correspondence between the proxy and SST data sets. Only the 3.6 year peak exceeds the 80% coherence level in both time series; a weaker coherence at 5.3 years is of possible interest because it also occurs in the MTM analysis of the entire time series (Figure 8a). Intervening intervals of distinctly lower coherence suggest that in the absence of strong (ENSO) forcing the coral/SST relationship is weaker. This pattern also occurs in the time domain comparison of the predicted SSTs with observations (Figure 3) and could reflect some aliasing of coral response to high-frequency (~1 yr) environmental fluctuations [Taylor et al., 1995]. However, it is important to recall that errors in the SST reconstruction, especially for the earlier part of the 20th century, may also be responsible for some of the weaker correspondences between low-frequency variations of SST and $\delta^{18}\text{O}$.

4.2. Comparison With ENSO Indices

In addition to comparisons with local SST variations it is of interest to determine whether the New Caledonia coral records information about larger-scale Pacific Basin fluctuations. We

compared the coral record in both the time and frequency domains against Quinn's [1992] ENSO reconstruction and the Southern Oscillation index (SOI; available through the Climate Research Unit, University of East Anglia) [Ropelewski and Jones, 1987; Allan et al., 1991]. Environmental changes associated with ENSO at New Caledonia should be manifest in the coral $\delta^{18}\text{O}$ record as positive interannual excursions, because the climate response to ENSO at New Caledonia is a decrease in rainfall and SST. However, because

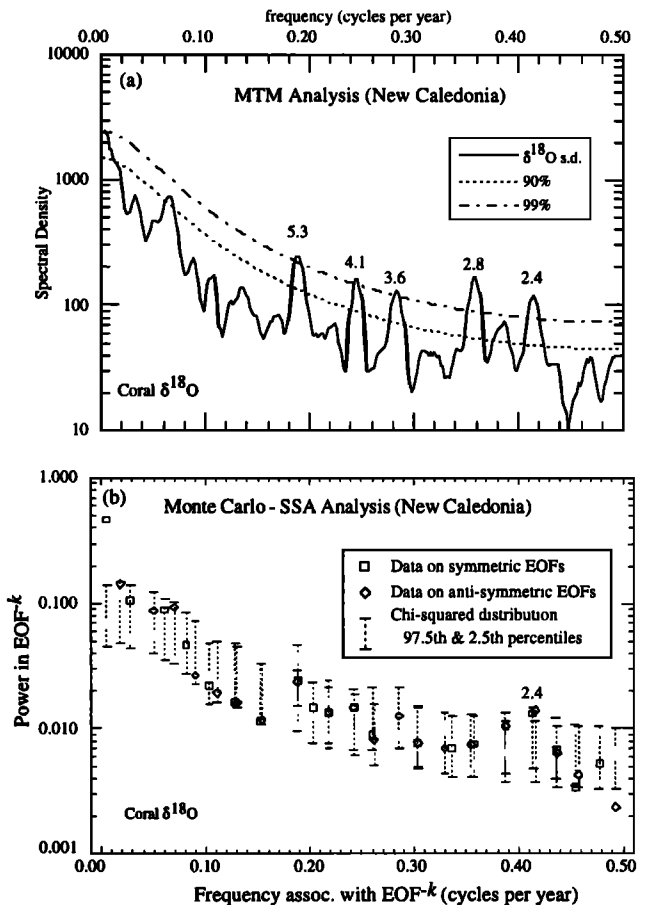


Figure 8. Frequency domain analysis of the (a) seasonal and (b) mean annual coral $\delta^{18}\text{O}$ records. Figure 8a is a plot of spectral density versus frequency as produced by the multi-taper method (MTM). Figure 8b plots power versus frequency as produced by Monte Carlo singular spectrum analysis (MC-SSA) [Allen and Smith, 1996]. The technique involves comparing the observed power (squares and diamonds) in the EOFs corresponding approximately to the frequencies shown with the expected range of power in a large ensemble of input series, taking into account the presence of the trend (>40-year variability is contained in EOFs 1 and 2). This approach provides a significance test against the "trend-plus-red-noise" null-hypothesis, for which we display the 2.5th and 97.5th percentiles. Only the ~2.4-year peak is significant at the 97.5% level (the trend is also significant against pure red noise), while an EOF with associated period 14.3 years is significant at the 94% level. The other member of this pair has presumably been lost through degeneracy. See text for further discussion of these techniques and results.

New Caledonia is located near the "hinge line" separating regions of opposite response in the Southern Oscillation [e.g., *Trenberth and Shea*, 1987], fluctuations in this region may not correlate as strongly with ENSO indices as those regions closer to the points of highest correlations (e.g., eastern equatorial Pacific, Tahiti, and Darwin).

Some of the largest isotopic excursions in the coral $\delta^{18}\text{O}$ record (e.g., 1728, 1737, 1790, 1836, 1885, and 1941) agree approximately with events listed in the El Niño time series of *Quinn* [1992]. However, the strength of ENSO event and magnitude of the isotopic excursion are not necessarily directly related. Furthermore, *Crowley et al.* [1997] demonstrated that there is a significant correlation between volcanic eruptions and some of the $\delta^{18}\text{O}$ excursions, so the precise attribution of some of the New Caledonia $\delta^{18}\text{O}$ variations is still open to question.

Cross-spectral analysis (Figure 4b) indicates that the two seasonal time series are coherent at the 80% level at periods of 3.6 and 7.2 years; however, only the 3.6 year peak occurs in the individual time series. As this peak also occurs in the $\delta^{18}\text{O}$ /SST comparison (see above), it appears to be the most robust signal of larger-scale Pacific Basin fluctuations to be recorded at the Amédée Island site. The 3.6 year peak is of considerable interest because it has also been found in another analysis of the Southern Oscillation index [*Allen and Smith*, 1996; *Brassington*, 1997], in temperature records [*Mann and Park*, 1994], atmospheric pressure records [*Tudhope et al.*, 1995], and in coral records from New Guinea [*Tudhope et al.*, 1995] and Tarawa [*Cole et al.*, 1993]. Standard SSA (not shown) also indicates that there is an increased amplitude of variance in the 3.6 year peak after a relatively quiescent period of ENSO activity from ~1940 to 1960. This pattern has also been found in ENSO records [*Quinn et al.*, 1987; *Brassington*, 1997]. This change in ENSO activity may be associated with a change in the extratropical northern hemisphere Pacific North/American (PNA) pattern identified by *Wallace et al.* [1993] and found in a coral record from the Gulf of Mexico [*Slowey and Crowley*, 1995].

4.3. Time Domain Fluctuations

We first discuss the growth rate trend. Growth rates vary from a low of 0.55 cm yr⁻¹ in the late 1600s to a high of 1.4 cm yr⁻¹ in 1930 and 1960 (Figure 5). There is a long-term increase in growth rate with decreasing age, but there are also periods when that trend is reversed (e.g., 1780-1820 and 1960-1992). Annual values of coral growth rate are negatively correlated with mean annual values of coral $\delta^{18}\text{O}$ ($r = -0.39$ and $p < 0.01$) and $\delta^{13}\text{C}$ ($r = -0.52$ and $p < 0.01$). *McConnaughey* [1989], working with corals whose growth rate varied from 4 to 8 mm yr⁻¹, suggested that there is a relation between growth rate and $\delta^{18}\text{O}$ via a kinetic isotopic fractionation effect. However, *Leder et al.* [1996], working with corals whose growth rate varied from 1 to 8 mm yr⁻¹, concluded that a kinetic isotopic fractionation effect was a minor one on coral $\delta^{18}\text{O}$. There are few indications that the New Caledonia coral $\delta^{18}\text{O}$ record is greatly influenced by growth rate changes because of the relatively low correlation, the fact that lowest growth rates in the coral (5–6 mm yr⁻¹) are significantly above the minimum values discussed by *Leder et al.* [1996], and because there is no relationship between significant decadal isotopic

oscillations and growth rate changes (e.g., the late 19th century).

The long-term coral $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ trends are of isotopic depletion toward the present (Figures 6 and 7). However, this trend is not uniform. Coral $\delta^{18}\text{O}$ values reached and began to fluctuate about the modern average value in 1944. There is also significant interannual and decadal-scale variability in the record. For example, the cooling interval in the early 1800s has been found in a number of coral records and in records of northern hemisphere temperature variations [*Bradley and Jones*, 1993; *Crowley and Kim*, 1996; *Quinn et al.*, 1996a; *Crowley et al.*, 1997]. After oscillating around some long-term mean a trend toward decreasing coral $\delta^{13}\text{C}$ values began in the mid-1800s, with a mean decrease of ~0.9‰ during this time. The long-term decrease may reflect in part the anthropogenic perturbation of the atmospheric ^{13}C reservoir by the burning of isotopically light fossil fuels (see also section 5). The amplitude of the annual cycle is also markedly reduced in the late 20th century portion of the $\delta^{13}\text{C}$ record.

As discussed above (section 4.1) we employed the 1952–1991 interval for predicting mean annual temperatures in the earlier part of the record with the OLS calibration. Other mean annual calibrations (with different smoothing or time intervals) yielded similar SST estimates, but the standard deviations varied as a function of the correlation coefficient (Table 1). Given the present uncertainties with respect to coral SST calibrations (see above), we prefer to err on the conservative side with respect to the reliability of paleo-SST estimates.

Application of the mean annual temperature calibration to the entire length of the New Caledonia time series indicates that temperatures during the primarily preanthropogenic interval of 1657-1900 are ~0.3°C less than 20th century temperatures (Figure 6a). These are large interdecadal oscillations on the order of ~0.5°–0.8°C. As discussed by *Crowley et al.* [1997], some of the interannual cooling events appear to be of volcanic origin. For example, the ~0.5°C cooling following the 1883 Krakatau eruption is comparable to or slightly larger than that following the 1991 Pinatubo eruption [*Gagan and Chivas*, 1995]. This is consistent with the different southern hemisphere aerosol loading estimate for the two events [*Crowley et al.*, 1997]. Further inspection of the temperature estimates indicates that there is only a brief interval of modest cooling in the late 17th century that correlates with the maximum "Little Ice Age" cooling in northern hemisphere temperature records [*Bradley and Jones*, 1993; *Crowley and Kim*, 1996]. However, the early 18th and 19th century patterns agree better with the northern hemisphere temperature record of *Mann et al.* [1998; cf. *Crowley and Kim*, 1996; *Crowley et al.*, 1997] and likely reflects the strong imprint of volcanism at this time [*Crowley et al.*, 1997].

4.4. Frequency Domain Fluctuations

As discussed above (section 3.3) the coral isotope data were analyzed using a variety of time series techniques. The MTM and MC-SSA approaches are illustrated in Figure 8, along with significance estimates. The results differ according to method employed. The techniques identify the peak around 2.4 years as the most robust signal in the coral $\delta^{18}\text{O}$ time series. Even

though the MTM method yields additional peaks in the ENSO band that occur as EOF pairs in ordinary SSA analysis (2.8, 3.6, 4.1, and 5.3 years), inability to reproduce these peaks with the MC-SSA analysis suggests that caution is needed in drawing conclusions about the robustness of these peaks. These EOFs are quite sensitive to variation of the series length or window width. This sensitivity can occur in genuine signals, for reasons given in Allen and Smith [1997]. Results are much more robust using the revised algorithm in Allen and Smith [1997], but in this paper we restrict attention to standard SSA.

The MC-SSA significance test indicates that only the ~2.4-year EOF pair contains more variance than we would expect from a red noise continuum background. The main reason for the apparent discrepancy between the MTM and MC-SSA results (in terms of number of significant peaks) is that MTM, as applied here, makes use of the full-resolution Fourier transform of the time series, implicitly using information from the series auto-covariance function out to lags of hundreds of years. SSA, with a 40-year window, explicitly confines attention to lags of up to 39 years. The MTM spectrum therefore appears to contain more information, but given that the physical "memory" of the ENSO phenomenon is likely much shorter than even 40 years, it is unclear whether the additional information contained in the covariance structure of the series at the longer lags is physically meaningful.

Further testing is therefore needed to determine the most appropriate window width for coral studies. At this relatively early stage in the investigation, we include both approaches in Figure 8 for the sake of comparison. MTM results on the coral $\delta^{13}\text{C}$ record (not shown) also indicate concentrations of variance in the ENSO band, but the same caveats noted for the $\delta^{18}\text{O}$ analysis apply here as well. We defer any discussion on time evolution of individual spectral peaks (evolutionary spectral analysis) until the reality of their existence can be established with greater confidence.

There is marginally significant (92-94%) evidence for an interdecadal peak around 14.3 to 15.4 years in both MTM and MC-SSA (Figure 8 and standard SSA, not shown). Analyses of other long time series (of different length) indicate an interdecadal peak in a coral from Vanuatu [Quinn et al., 1996a], instrumental data [Ghil and Vautard, 1991; Mann and Park, 1994], and a climate model [Latif and Barnett, 1994]. If verified, this peak may be relevant to understanding decadal variability in the Pacific Basin [e.g., Zhang et al., 1997]. Because of its marginal significance, more testing is needed to evaluate the reality of this feature.

5. Discussion

5.1. Interpretation of Isotope Records

5.1.1. Coral $\delta^{18}\text{O}$. The precise partitioning of the coral $\delta^{18}\text{O}$ signal into SST and SSS (via changes in seawater $\delta^{18}\text{O}$) requires the measurement of both variables at similar timescales. However, coral studies that use reef-site $\delta^{18}\text{O}_{\text{seawater}}$ determinations are still relatively rare [e.g., Wellington et al., 1996]. A more typical approach is to use existing climatological data to determine the relative contributions of salinity and temperature on the coral $\delta^{18}\text{O}$

signal. There is no presently available time series of seawater $\delta^{18}\text{O}$ time series at our coral site, although as discussed by Quinn et al. [1996b], two spot measurements of $\delta^{18}\text{O}_{\text{seawater}}$ 5 years apart are identical within the analytical error of the measurement (mean of 0.52‰). Application of an empirical equation relating salinity and $\delta^{18}\text{O}_{\text{seawater}}$ developed by Fairbanks et al. [1997] for convective regions of the western and central Pacific yields a mean value of 0.60‰ for the $\delta^{18}\text{O}_{\text{seawater}}$ at Amédée, New Caledonia. Although the amplitude of the seasonal salinity changes are small compared to SST changes [Quinn et al., 1996b], there are some indications from the salinity record for New Caledonia [Quinn et al., 1996b] of interannual changes of as much as 0.5‰. The potential impact of a 0.5‰ interannual salinity variation on $\delta^{18}\text{O}_{\text{seawater}}$ and hence on interannual coral $\delta^{18}\text{O}$ ranges from 0.14‰ [Fairbanks et al., 1997] to 0.15‰ [Craig and Gordon, 1965]. This effect could explain part of the difference in the seasonal and mean annual slopes of $\delta^{18}\text{O}/\text{SST}$ discussed above. The remaining difference could be attributed to stochastic processes (environmental and biological) that further reduce correlations (T. J. Crowley et al., manuscript in preparation, 1998). A multiyear study of variations in the $\delta^{18}\text{O}_{\text{seawater}}$ at New Caledonia, which is now in progress, is required to fully address this issue.

5.1.2. Coral $\delta^{13}\text{C}$. Variations in coral $\delta^{13}\text{C}$ are notoriously difficult to interpret in terms of an environmental signal, because questions remain regarding coral modification of the dissolved inorganic carbon pool in seawater [McConnaughey, 1989; Swart et al., 1996]. Several workers have noted a correlation between water depth and coral $\delta^{13}\text{C}$ values [e.g., Weber et al., 1976; Land et al., 1977; Swart and Coleman, 1980]. This linear decrease in coral $\delta^{13}\text{C}$ with increasing water depth has been attributed to light-level-induced changes in photosynthetic activity of the symbiotic zooxanthellae algae [e.g., Weber et al., 1976; Fairbanks and Dodge, 1979; McConnaughey, 1989]. Cloudiness changes may be important [Winter et al., 1991]. A recent study by Swart et al. [1996] concluded that there was a weak, but statistically significant, positive correlation between atmospheric pressure (light proxy) and coral $\delta^{13}\text{C}$ values.

There is an ~0.9‰ decrease in coral $\delta^{13}\text{C}$ at New Caledonia from ~1850 to the present (Figures 6 and 7). The magnitude of the depletion in coral $\delta^{13}\text{C}$ in the 20th century is 0.6‰, a value that is very close to what is expected from a box diffusion model estimate of $\delta^{13}\text{C}$ in a 75 m thick mixed layer [e.g., Druffel and Benavides, 1986]. Although these changes could reflect the anthropogenic perturbation of the ^{13}C reservoir, a number of cautionary comments are necessary. The change in $\delta^{13}\text{C}$ from 1970 to 1990 is 0.27‰, which is 32% less than mixed layer changes found by Quay et al. [1992] from $\delta^{13}\text{C}$ observations of ΣCO_2 from nearby geochemical transects. Furthermore, the $\delta^{13}\text{C}$ change occurs fairly abruptly in 1988-1989, at about the same time as a climate shift often recorded elsewhere in the Pacific [e.g., Zhang et al., 1997]. Prior to that time the $\delta^{13}\text{C}$ values change only 0.1‰ between 1950-1985.

In a general sense, atmospheric ^{13}C changes must impact the surface ocean, and without significant upwelling from great depths, coral $\delta^{13}\text{C}$ should record the anthropogenic perturbation of the ^{13}C reservoir. However, given what we know about $\delta^{13}\text{C}$ in corals, biological activity may well complicate the

oceanographic interpretation of the coral $\delta^{13}\text{C}$ record. As a result, it is not uncommon for coral $\delta^{13}\text{C}$ time series to remain unpublished despite the fact that they are generated at the same time as coral $\delta^{18}\text{O}$ time series [e.g., *Cole and Fairbanks*, 1990; *Dunbar et al.*, 1994; *Tudhope et al.*, 1995]. We believe it is important to publish and archive such data for the record.

5.2. Comparison With Other Climate Records

The 20th century portion of the New Caledonia coral $\delta^{18}\text{O}$ record contains four transitions that are especially noteworthy as they have also been identified in instrumental climate records: an abrupt warming beginning in 1940-1941, a cooling after 1958 and after 1976-1977, and a warming after 1988-1989 (Figures 6 and 9). The shift in 1988-1989 that occurs in both the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ records also occurs in a number of Pacific records [*Minobe*, 1997; *Zhang et al.*, 1997] and is associated with a step increase in global temperatures [e.g., *Jones*, 1994]. The post 1976-1977 cooling trend corresponds to a decadal-scale climate transition that is a topic of much recent interest [e.g., *Graham*, 1994, 1995; *Trenberth and Hurrell*, 1994]. As discussed in section 4.3, a shift in 1958 has also been found in a New Guinea coral [*Tudhope et al.*, 1995] and may be related to a significant change in the structure of the PNA over North America [*Wallace et al.*, 1993; *Slowe and Crowley*, 1995].

The calibration database for the 1940-1941 transition is small. However, *Zhang et al.* [1997; cf. *Minobe*, 1997] discuss a trend to cooler temperatures in a record strongly influenced by temperature variations in the eastern equatorial Pacific cold tongue. Fluctuations in the latter region are of opposite sign to what is observed in the western South Pacific during El Niño years and strongly influence global temperature records [*Graham*, 1995]. A well-known decrease/stabilization of global average temperatures also occurred around this time [e.g., *Jones*, 1994]. There are similar prominent shifts in North Pacific SSTs, North Pacific sea level pressure, and

snowmelt runoff in California at the same time [*Dettinger and Cayan*, 1995; *Zhang et al.*, 1997].

There is a 0.23‰ shift in mean $\delta^{18}\text{O}$ values associated with the 1941 transition. This value is twice as large as the 1976-1977 transition and is also substantially larger (proportionately) than the shift identified by *Zhang et al.* [1997] for low-frequency global SST changes. However, this large $\delta^{18}\text{O}$ step is consistent and a shift in the Pacific (inter)Decadal Oscillation (PDO) index (Figure 9) of *Mantua et al.* [1997] and with a large SST shift in the subpolar North Pacific identified by *Dettinger and Cayan* [1995]. The correlation coefficient between the PDO and the mean annual coral $\delta^{18}\text{O}$ record is 0.55 ($p < 0.01$). The subpolar North Pacific fluctuates approximately in parallel to the western South Pacific on decadal timescales [e.g., *Zhang et al.*, 1997]. Further investigation of the New Caledonia/North Pacific relationship might be very useful for discriminating among different models of decadal-scale variability, for these models make different predictions about the timescale required to transmit signals from the North Pacific to the tropics [e.g., *Latif and Barnett*, 1994; *McCreary and Lu*, 1994; *Gu and Philander*, 1997; *Lysne et al.*, 1997].

The pre-20th century SST reconstruction at New Caledonia, in addition to being relevant to testing coupled climate models examining the problem of decadal-scale variability [*Latif and Barnett*, 1994], are valuable for testing control runs of coupled ocean-atmosphere general circulation models used for making greenhouse predictions. For example, a critical problem for detecting greenhouse-induced warming is whether the control runs yield the correct estimate of unforced variability (the "noise" in signal/noise detection studies [e.g., *Hegerl et al.*, 1997]). The grid box in the Hamburg/Max Planck Institute ECHAM3 coupled model [*Voss et al.*, 1998] has a variability of 0.28°C (1 σ) for the $\sim 4^\circ \times 5^\circ$ grid enclosing New Caledonia. This number agrees better with that determined by the mean annual temperature equation (0.23°C, 1 σ) than it does with that obtained from the seasonal regression (0.99°C, 1 σ). The above mentioned coral aliasing at the annual cycle (section 4.1) and some evidence for forced variability in the New Caledonia record [*Crowley et al.*, 1997] are further cautionary reasons against drawing too strong a conclusion from this agreement. Nevertheless, we still consider the agreement of more than passing interest.

5.3. Comparison With Other Coral Records

There are now several century-scale coral time series from the Pacific: four from the south Pacific, two from the western Pacific, one from the central Pacific, and two from the eastern Pacific. South Pacific coral records include a quarterly $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ time series from Espiritu Santo, Vanuatu [*Quinn et al.*, 1996a], a biannual $\delta^{18}\text{O}$, $\delta^{13}\text{C}$, and $\Delta^{14}\text{C}$ time series from Abraham Reef (22°S and 153°E), the Swains Reef complex in the southwestern tip of the GBR [*Druffel and Griffin*, 1993], and a series of coral growth rate records from a number of GBR corals [*Lough et al.*, 1996]. West Pacific coral records include a mean annual $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ time series from Cebu, Philippines [*Patzold*, 1986] and a shorter seasonal $\delta^{18}\text{O}$ time series from Madang, New Guinea [*Tudhope et al.*, 1995]. The central Pacific record is the subseasonal $\delta^{18}\text{O}$ time series from Tarawa Atoll [*Cole et al.*, 1993]. The two coral records from the east-

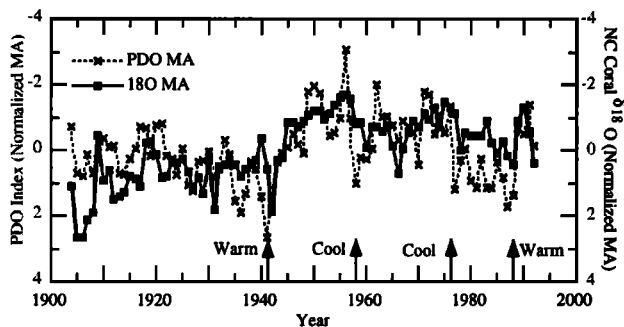


Figure 9. A comparison of 20th century mean annual $\delta^{18}\text{O}$ record from New Caledonia with the Pacific (inter)Decadal Oscillation (PDO); [*Mantua et al.*, 1997]. The PDO index is constructed from an empirical-orthogonal function (EOF) analysis of North Pacific SST anomalies. Arrows highlight climate transitions of 1940-1941 [*Zhang et al.*, 1997], 1958 [*Wallace et al.*, 1993], 1976-1977 [e.g., *Graham*, 1994, 1995; *Trenberth and Hurrell*, 1994], and 1988-1989 [e.g., *Minobe*, 1997; *Zhang et al.*, 1997] that are recorded in both the New Caledonia and PDO records.

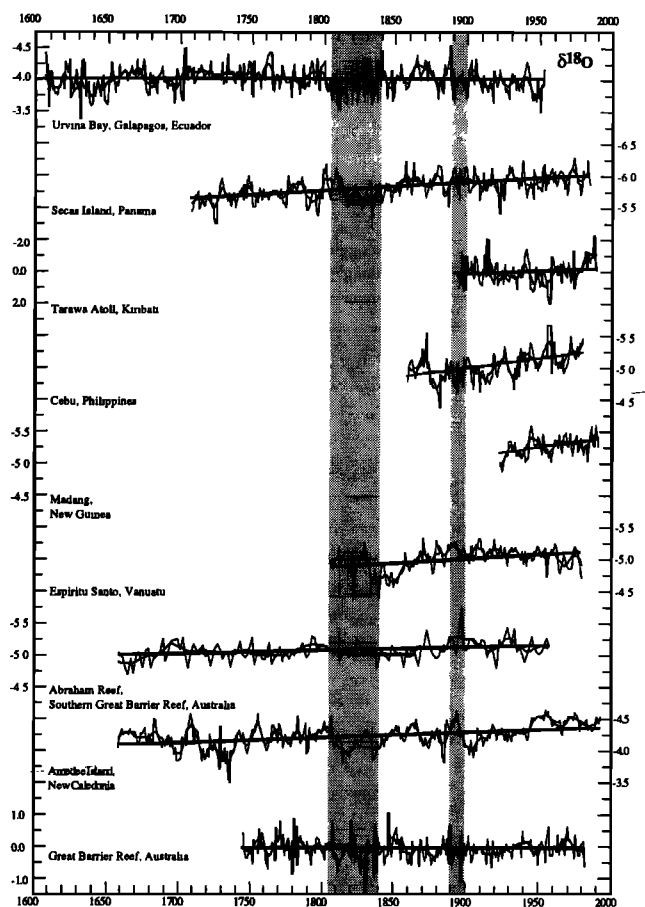


Figure 10. A comparison of Pan-Pacific coral time series. All records are plotted as mean annual time series (including a 5 year moving average) and with a least squares linear fit through the mean annual time series. Individual time series include: mean annual $\delta^{18}\text{O}$ record from Urvina Bay, Galápagos (UBG) [Dunbar et al., 1994], subseasonal $\delta^{18}\text{O}$ record from Secas Island, Panama (SIP) [Linsley et al., 1994]; subseasonal $\delta^{18}\text{O}$ record from Tarawa Atoll, Kiribati (TAK) [Cole et al., 1993]; mean annual $\delta^{18}\text{O}$ record from Cebu, Philippines (CBP) [Patzold, 1986]; seasonal $\delta^{18}\text{O}$ record from Madang, New Guinea (MDG) [Tudhope et al., 1995]; seasonal $\delta^{18}\text{O}$ record from Espiritu Santo, Vanuatu (ESV) [Quinn et al., 1996a]; biannual $\delta^{18}\text{O}$ record from Abraham Reef, southern GBR, Australia (ARA) [Druffel and Griffin, 1993]; seasonal $\delta^{18}\text{O}$ record from Amédée, New Caledonia (ANC) [Quinn et al., this study]; and mean annual coral growth rate record from 10 GBR corals (GR GBR) [Lough et al., 1996]. Shaded bars represent intervals discussed in text.

ern Pacific are the annual $\delta^{18}\text{O}$ time series from Urvina Bay, Galápagos Islands [Dunbar et al., 1994] and the subseasonal $\delta^{18}\text{O}$ time series from Secas Island, Panama [Linsley et al., 1994]. As discussed by Crowley et al., [1997] the detailed chronology of some of these records may be in error by a small number of years.

A time domain comparison of the mean annual versions of the various coral time series (Figure 10 and Table 2) indicates that whereas each time series contains interannual and decadal-

scale variability, it is a rare occurrence when this variability is shared amongst the various records. On the largest scale, there is at best only modest evidence for any long-term trends in most of the records. Other than Secas Island and Cebu, the records have either no trend (Galápagos, Great Barrier Reef growth rate record, and Abraham Reef), flat or negative trends in the latter part of the records (Vanuatu, New Caledonia), or a departure at the end of the record associated with the 1976 climate transition that dominates the entire trend (Tarawa). A number of the records have a warm-cold-warm-cold pattern in the early 19th century (1805-1840) that can also be found in northern hemisphere records [Bradley and Jones, 1993]. The larger scale of coherence may be related to the major pulse of volcanism that occurred during this time Crowley et al. [1997]. There is also some evidence for a decadal-scale warming in the last decade of the 19th century in the three western South Pacific isotope records.

A cross spectral comparison of the New Caledonia record with the various Pacific coral time series (not shown) indicates little coherent trends among the coral records — coherent peaks that occur in some comparisons do not occur in others. It is not clear whether this lack of large scale coherence represents a basic limitation of the coral records or whether there are real differences in response between different parts of the ocean basins. Two reasons to consider the latter possibility involve, for example, a study by Tudhope et al. [1995] for the western equatorial Pacific showing differences in response of the local sea level pressure pattern from the main Tahiti-Darwin Southern Oscillation Index. Also, M. Latif [personal communication, 1997] compared the interdecadal oscillations in a coupled ocean-atmosphere model [Latif and Barnett, 1996] at sites coincident with the coral sites and, despite the existence of an interdecadal mode in the model, the correlations from point to point were quite low in the model.

Table 2. Linear, Zero-Lag Ordinary Least Square Correlation Coefficients Between Eight Pan-Pacific Coral Time Series

	UBG	SIP	TAK	ESV	ANC	ARA	GBR	CBP	MDG
UBG	1.00								
SIP	(-0.39)	1.00							
TAK	0.41	0.04	1.00						
ESV	-0.05	(0.37)	-0.17	1.00					
ANC	-0.20	(0.56)	-0.22	(0.29)	1.00				
ARA	-0.19	(0.43)	-0.22	(0.49)	0.10	1.00			
GBR	-0.04	-0.04	-0.04	(0.28)	-0.16	-0.22	1.00		
CBP	-0.24	0.10	-0.30	0.24	(0.50)	0.24	(0.44)	1.00	
MDG	(-0.59)	0.23	0.26	-0.31	(0.45)	-0.07	-0.30	0.44	1.00

The abbreviations are defined as follows:

UBG, coral $\delta^{18}\text{O}$, Urvina Bay, Galápagos [Dunbar et al., 1994]; SIP, coral $\delta^{18}\text{O}$, Secas Island, Panama [Linsley et al., 1994]; TAK, coral $\delta^{18}\text{O}$, Tarawa Atoll, Kiribati [Cole et al., 1993]; ESV, coral $\delta^{18}\text{O}$, Espiritu Santo, Vanuatu [Quinn et al., 1996a]; ANC, coral $\delta^{18}\text{O}$, Amédée, New Caledonia [Quinn et al., this study]; ARA, coral $\delta^{18}\text{O}$, Abraham Reef, Great Barrier Reef, Australia [Druffel and Griffin, 1993]; GBR, average coral growth rate, 10 corals, Great Barrier Reef, Australia [Lough et al., 1996]; CBP, coral $\delta^{18}\text{O}$, Cebu, Philippines [Patzold, 1986]; MDG, coral $\delta^{18}\text{O}$, Madang, Philippines [Tudhope et al., 1995]. Values in parentheses are significant at $p < 0.01$.

6. Conclusions

1. A 335 year coral record (1657-1992) has been collected near Amédée, New Caledonia, and sampled at the seasonal timescale. A prior sampling study and new data from the early 20th century indicate that quarterly sampling yields estimates of mean annual $\delta^{18}\text{O}$ very similar to 12 yr⁻¹ sampling. Comparison of distinctive interannual scale cooling events in the record with a volcano time series suggests that the coral chronology may be accurate to within a year over much of its length.

2. Although the seasonal $\delta^{18}\text{O}$ record is mainly controlled by SST, salinity may play a proportionately larger role in mean annual $\delta^{18}\text{O}$ changes. This factor may contribute to a weaker correlation between SST and $\delta^{18}\text{O}$ on the annual timescale relative to the seasonal timescale, leading to a different $\delta^{18}\text{O}/\text{SST}$ calibration for each timescale. Only the mean annual calibration provides the correct estimate for early 20th century SSTs.

3. A comparison of $\delta^{18}\text{O}$ with observations (Southern Oscillation and the regionally gridded SST field) indicates coherence in the interannual band. Significant decadal-scale changes in 1940-1941, 1976-1977, and 1988-1989 agree with other instrumental observations from the Pacific Basin.

4. Application of the annual $\delta^{18}\text{O}/\text{SST}$ calibration indicates that temperatures between 1657 and 1900 are $\sim 0.3^\circ\text{C}$ lower than the 20th century average, with temperatures hovering around a near-constant value for the last 50 years. Interdecadal oscillations were as large as $0.5^\circ\text{--}0.8^\circ\text{C}$. The standard deviation of the estimated mean annual temperatures over this interval (0.23°C) is comparable to that predicted by a control run of a coupled ocean-atmosphere model. Although early 19th century cooling is comparable to that which occurred in northern hemisphere temperature records, the late 17th century cooling at New Caledonia (1695-1702) is barely distinguishable from the early 20th century cooling from this region. A strong early 18th century cooling at New Caledonia does not compare well with northern hemisphere fluctuations.

5. The $\delta^{13}\text{C}$ shows a 0.9‰ decrease from ~ 1850 to the present which is consistent with the perturbation of the atmospheric ^{13}C reservoir by fossil fuel burning. However, there are prominent periods of stasis (1950-1985) and abrupt change (1988-1989) that parallel some of the shifts in the $\delta^{18}\text{O}$ record, suggesting that the link between atmospheric ^{13}C , coral $\delta^{13}\text{C}$, and ocean dynamics is complex.

6. Time series analysis of the New Caledonia coral $\delta^{18}\text{O}$ record documents concentrations of variance in the El Niño band and a marginally significant interdecadal oscillation. The statistical robustness of some of these peaks requires continued evaluation.

7. Comparisons with multicentury coral records from the Pacific indicate only modest levels of agreement between the sites in the time domain. There is some evidence for a packet of warm-cold-warm-cold decadal oscillations in the early 1800s that may be forced from a pulse of volcanism at that time. Three western South Pacific isotope records also show evidence for a decadal-scale warming in the last decade of the 19th century. There is relatively little coherent structure in the frequency domain among the different coral records in the interdecadal band.

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References

- Allan, R.J., N. Nicholls, P.D. Jones, and I.J. Butterworth, A further extension of the Tahiti-Darwin SOL, early ENSO events and Darwin pressure, *J. Clim.*, **4**, 743-749, 1991.
- Allen, M.R. and L.A. Smith, Monte Carlo SSA: Detecting irregular oscillations in the presence of colored noise, *J. Clim.*, **9**, 3373-3404, 1996.
- Allen, M.R. and L.A. Smith, Optimal filtering in Singular Spectrum Analysis, *Physics Lett.*, **234**, 419-428, 1997.
- Beck, J.W., R.L. Edwards, E. Ito, F.W. Taylor, J. Recy, F. Rougerie, P. Joannot, and C. Henin, Sea surface temperature from skeletal Sr/Ca ratios, *Science*, **257**, 585-716, 1992.
- Bradley, R.S., and P.D. Jones, "Little Ice Age" summer temperature variations: Their nature and relevance to recent global warming trends, *Holocene*, **3**, 367-376, 1993.
- Brassington, G.B., The modal evolution of the Southern Oscillation, *J. Clim.*, **10**, 1021-1034, 1997.
- Cole, J.E., Coral records of climate change: Understanding past variability in the tropical ocean-atmosphere, in *Climatic Variations and Forcing Mechanisms of the Last 2000 Years*, edited by P.D. Jones, R.S. Bradley, and J. Jouzel, pp. 331-354, Springer-Verlag, New York, 1996.
- Cole, J.E., and R.G. Fairbanks, The Southern Oscillation recorded in the $\delta^{18}\text{O}$ of corals from Tarawa Atoll, *Paleoceanography*, **5**, 669-683, 1990.
- Cole, J.E., R.G. Fairbanks, and G.T. Shen, Recent variability in the Southern Oscillation: Isotopic results from a Tarawa Atoll coral, *Science*, **260**, 1790-1793, 1993.
- Craig, H., and L.I. Gordon, Isotopic oceanography: Deuterium and oxygen-18 variations in the ocean and marine atmosphere, in: *Marine Geochemistry*, edited by D.R. Schink and J.T. Corless, pp. 227-234, Univ. of R. I. Press, Narragansett, 1965.
- Crowley, T.J., and K.-Y. Kim, Comparison of proxy records of climate change and solar forcing, *Geophys. Res. Lett.*, **23**, 359-362, 1996.
- Crowley, T.J., T.M. Quinn, and F.W. Taylor, Evidence for a volcanic cooling signal in a 335 year coral record from New Caledonia, *Paleoceanography*, **12**, 633-639, 1997.
- Dai, J., E. Mosley-Thompson, and L.G. Thompson, Ice core evidence for an explosive tropical volcanic eruption 6 years preceding Tambora, *J. Geophys. Res.*, **96**, 17361-17366, 1991.
- Davis, J.C., *Statistics and Data Analysis in Geology*, 2nd ed., 646 pp., Wiley and Sons, New York, 1986.
- Dettinger, M.D., and D.R. Cayan, Large-scale atmospheric forcing of recent trends toward early snowmelt runoff in California, *J. Clim.*, **8**, 606-623, 1995.
- Dettinger, M.D., M. Ghil, C.M. Strong, W. Weibel, and P. Yiou, Software expedites singular-spectrum analysis of noisy time series, *Eos, Trans. AGU*, **76(2)**, 12, 14, 21, 1995.
- Donguy, J.R., and C. Henin, Climatic teleconnections in the western South Pacific with the El Niño phenomena, *J. Phys. Oceanogr.*, **10**, 1952-1958, 1980.
- Druffel, E.R.M., and L.M. Benavides, Input of excess CO_2 to the surface ocean based on $^{13}\text{C}/^{12}\text{C}$ ratios in a banded Jamaican sclerosponge, *Nature*, **321**, 58-61, 1986.
- Druffel, E.R.M., and S. Griffin, Large variations of surface ocean radiocarbon: Evidence of circulation changes in the southwestern Pacific, *J. Geophys. Res.*, **98**, 20,249-20,259, 1993.
- Dunbar, R.B., and J.E. Cole, Coral records of ocean-atmosphere variability, National Oceanic and Atmos. Admin. Clim. and Global Change Program, *Spec. Rep. 10*, 38 pp., Boulder, CO, 1993.
- Dunbar, R.B., G.M. Wellington, M.W. Colgan, and P.W. Glynn, Eastern Pacific sea surface

- temperature since 1600 A.D.: The $\delta^{18}\text{O}$ record of climate variability in Galápagos corals, *Paleoceanography*, 9, 291-315, 1994.
- Epstein, S., R. Buchsbaum, H.A. Lowenstam, and H.C. Urey, Revised carbonate-water isotopic temperature scale, *Geol. Soc. Am. Bull.*, 64, 1315-1326, 1953.
- Erez, J., and B. Luz, Experimental paleotemperature equation for planktonic foraminifera, *Geochim. Cosmochim. Acta*, 47, 1025-1031, 1983.
- Fairbanks, R.G., and R.E. Dodge, Annual periodicity of the $^{18}\text{O}/^{16}\text{O}$ and $^{13}\text{C}/^{12}\text{C}$ ratios in the coral *Montastrea annularis*, *Geochim. Cosmochim. Acta*, 43, 1009-1020, 1979.
- Fairbanks, R.G., M.N. Evans, J.L. Rubenstone, R.A. Mortlock, K. Broad, M.D. Moore, and C.D. Charles, Evaluating climate indices and their geochemical proxies measured in corals, in *Proceedings of the 8th International Coral Reef Symposium, Panama*, edited by H.A. Lessios and I.G. Macintyre, 107-116, Smithsonian Tropical Research Institute, Balboa, Panama, 1997.
- Gagan, M.K., and A.R. Chivas, Oxygen isotopes in western Australian coral reveal Pinatubo aerosol-induced cooling in the Western Pacific Warm Pool, *Geophys. Res. Lett.*, 22, 1069-1072, 1995.
- Gagan, M.K., A.R. Chivas, and P.J. Isdale, High-resolution isotopic records from corals using ocean temperature and mass spawning chronometers, *Earth Planet. Sci. Lett.*, 121, 549-558, 1994.
- Ghil, M., and R. Vautard, Interdecadal oscillations and the warming trend in global temperature time series, *Nature*, 350, 324-327, 1991.
- Graham, N.E., Decadal-scale climate variability in the tropical and North Pacific during the 1970s and 1980s: Observations and model results, *Clim. Dyn.* 10, 135-162, 1994.
- Graham, N.E., Simulation of recent global temperature trends, *Science*, 267, 666-671, 1995.
- Grossman, E. L., and T.-L. Ku, Oxygen and carbon isotope fractionation in biogenic aragonite: temperature effects, *Chem. Geol.*, 59, 59-74, 1986.
- Gu, D., and S.G.H. Philander, Interdecadal climate fluctuations that depend on exchanges between the tropics and extratropics, *Science*, 275, 805-807, 1997.
- Hasselmann, K., Stochastic climate models, I, Theory, *Tellus*, 28, 473-485, 1976.
- Hegerl, G.C., K. Hasselmann, U. Cubasch, J.F.B. Mitchell, E. Roeckner, R. Voss, J. Waszkewitz, Multi-fingerprint detection and attribution analysis of greenhouse gas, greenhouse gas-plus-aerosol and solar forced climate change, *Clim. Dyn.*, 13, 613-634, 1997.
- Imbrie, J., A. McIntyre, and A. Mix, Oceanic response to orbital forcing in the late Quaternary: Observational and experimental strategies, in *Climate and Geosciences*, edited by J. C. Duplessy et al., pp. 121-164, D. Reidel, Norwell, Mass, 1989.
- Jones, P.D., Hemispheric surface air temperatures: A reanalysis and update to 1993, *J. Clim.*, 7, 1794-1802, 1994.
- Land, L.S., J.C. Lang, and D.J. Barnes, On the stable carbon and oxygen isotopic composition of some shallow water, ahermatypic, scleractinian coral skeletons, *Geochim. Cosmochim. Acta*, 40, 31-39, 1977.
- Latif, M., and T.P. Barnett, Causes of decadal climate variability over the North Pacific and North America, *Science*, 266, 634-637, 1994.
- Leder, J.J., P.K. Swart, A. Szmant, and R.E. Dodge, The origin of variations in the isotopic record of scleractinian corals, I, Oxygen, *Geochim. Cosmochim. Acta*, 60, 2857-2870, 1996.
- Linsley, B.K., R.B. Dunbar, G.M. Wellington, and D.A. Mucciarone, A coral-based reconstruction of intertropical convergence zone variability over Central America since 1707, *J. Geophys. Res.*, 99, 9977-9994, 1994.
- Lough, J.M., D.J. Barnes, and R.B. Taylor, The potential of massive corals for the study of high-resolution climate variation in the past millennium, in *Climatic Variations and Forcing Mechanisms of the Last 2000 Years*, edited by P.D. Jones et al., NATO ASI Ser., 141, 355-371, 1996.
- Lysne, J., P. Chang, and B. Giese, Impact of extra-tropical Pacific on equatorial variability, *Geophys. Res. Lett.*, 24, 2589-2592, 1997.
- Mann, M. E., and J. M. Lees, Robust estimation of background-noise and signal-detection in climatic time-series, *Clim. Change*, 33, 409-445, 1996.
- Mann, M.E., and J. Park, Global-scale modes of surface temperature variability on interannual to century timescales, *J. Geophys. Res.*, 99, 25,819-25,833, 1994.
- Mann, M.E., R.S. Bradley, and M.K. Hughes, Global scale temperature patterns and climate forcing over the past six centuries, *Nature*, 392, 779-787, 1998.
- Mantua, N.J., S.R. Hare, Y. Zhang, J.M. Wallace and R.C. Francis, A Pacific interdecadal climate oscillation with impacts on salmon production, *Bull. Am. Met. Soc.*, 78, 1069-1079, 1997.
- McConaughy, T., ^{13}C and ^{18}O isotopic disequilibrium in biological carbonates, I, Patterns, *Geochim. Cosmochim. Acta*, 53, 151-162, 1989.
- McCreary, J.P., and P. Lu, Interaction between the subtropical and equatorial ocean circulations: The subtropical cell, *J. Phys. Oceanogr.*, 25, 3076-3088, 1994.
- Minobe, S., A 50-70 climatic oscillation over the North Pacific and North America, *Geophys. Res. Lett.*, 24, 683-686, 1997.
- Morlier, A., and J.P. Rebert, Rainfall shortage and El Niño-Southern Oscillation in New Caledonia, southwestern Pacific, *Mon. Weather Rev.*, 114, 1131-1137, 1986.
- Paillard, D., L. Labeyrie, and P. Yiou, Macintosh program performs time-series analysis, *Eos Trans. AGU*, 77 (39), 379, 1996.
- Parker, D.E., M. Jackson, and E.B. Horton, *The 1961-1990 GISS2.2 Sea Surface Temperature and Sea Ice Climatology*, *Clim. Res. Tech. Note 63 (CRTN63)*, Hadley Centre for Clim. Pred. and Res., Meteorol. Off., Bracknell, Berkshire, Eng., U.K., 1995.
- Patzold, J., Temperature and CO_2 changes in tropical surface waters of the Philippines during the past 120 years: Record in the stable isotopes of hermatypic corals, *Berichte (Rep.)*, 12, 82 pp., Geol.-Paläontol. Inst., Univ. of Kiel, Kiel, Germany, 1986.
- Quay, P.D., B. Tilbrook, and C.S. Wong, Oceanic uptake of fossil fuel CO_2 : Carbon-13 evidence, *Science*, 256, 74-79, 1992.
- Quinn, T.M., F.W. Taylor, and T.J. Crowley, A 173 year stable isotope record from a tropical South Pacific coral, *Quat. Sci. Rev.*, 12, 407-418, 1993.
- Quinn, T.M., T.J. Crowley, and F.W. Taylor, New stable isotope results from a 173-year coral record from Espiritu Santo, Vanuatu, *Geophys. Res. Lett.*, 23, 3413-3416, 1996a.
- Quinn, T.M., F.W. Taylor, T. J. Crowley, and S.M. Link, Evaluation of sampling resolution in coral stable isotope records: A case study using monthly stable isotope records from New Caledonia and Tarawa, *Paleoceanography*, 11, 529-542, 1996b.
- Quinn, W.H., A study of Southern Oscillation-related climatic activity for A.D. 622-1990 incorporating Nile River flood data, in *El Niño: Historical and Paleoclimatic Aspects of the Southern Oscillation*, edited by Diaz, H.F. and V. Markgraf, pp. 119-149, Cambridge Univ. Press, New York, 1992.
- Quinn, W.H., V. Neal, and S. Antunez de Mayolo, El Niño occurrences over the past four and a half centuries, *J. Geophys. Res.*, 92, 14,449-14,461, 1987.
- Rasmusson, E.M., and T.H. Carpenter, Variations in tropical sea surface temperature and surface wind fields associated with the Southern Oscillation/El Niño, *Mon. Weather Rev.*, 110, 354-384, 1982.
- Ropelewski, C.F., and P.D. Jones, An extension of the Tahiti-Darwin Southern Oscillation Index, *Mon. Wea. Rev.*, 115, 2161-2165, 1987.
- Rougerie, F., Le lagon sud-ouest de Nouvelle-Calédonie: Specificité hydrologique, dynamique et productivité, Ed. de l'ORSTOM, Ins. Français de Rech. Sci. Pour le Dev. en Coop., Collect. et études et thèses, 234 pp., New Caledonia, 1986.
- Shen, G.T., and R.B. Dunbar, Environmental controls on uranium in reef corals, *Geochim. Cosmochim. Acta*, 59, 2009-2024, 1995.
- Simkin, T., and L. Siebert, *Volcanoes of the World*, 2nd ed., 349 pp., Geosci. Press, Tucson, Ariz., 1994.
- Slowey, N.R., and T.J. Crowley, Interdecadal variability of northern hemisphere circulation recorded by Gulf of Mexico corals, *Geophys. Res. Lett.*, 22, 2345-2348, 1995.
- Sokal, R. R., *Biometry: The Principles and Practice of Statistics in Biological Research*, 3rd ed., 887 pp., W.H. Freeman, New York, 1995.
- Swart, P.K., Carbon and oxygen isotope fractionation in scleractinian corals: A review, *Earth Sci. Rev.*, 19, 51-80, 1983.
- Swart, P.K., and M.L. Coleman, Isotopic data for scleractinian corals explain their paleotemperature uncertainty, *Nature*, 283, 557-559, 1980.
- Swart, P.K., J.J. Leder, A.M. Szmant, and R.E. Dodge, The origin of variations in the isotopic record of scleractinian corals: II, Carbon, *Geochim. Cosmochim. Acta*, 60, 2871-2885, 1996.
- Taylor, R.B., D.J. Barnes, and J.M. Lough, On the inclusion of trace materials into massive coral skeletons, I, Materials occurring in the environment in short pulses, *J. Exp. Mar. Biol. Ecol.*, 185, 255-278, 1995.
- Thomson, D.J., Spectrum estimation and harmonic analysis, *Proc. IEEE*, 70, 1055-1096, 1982.
- Thomson, D.J., Time series analysis of Holocene climate data, *Philos. Trans. R. Soc. London, Ser. A*, 330, 601-616, 1990a.
- Thomson, D.J., Quadratic-inverse spectrum estimates: Applications to paleoclimatology, *Philos. Trans. R. Soc. London A*, 332, 593-597, 1990b.
- Trenberth, K.E., and J.W. Hurrell, Decadal atmosphere-ocean variations in the Pacific, *Clim. Dyn.* 9, 303-319, 1994.
- Trenberth, K.E., and D.J. Shea, On the evolution of the Southern Oscillation, *Mon. Weather Rev.*, 115, 3078-3096, 1987.
- Tudhope, A.W., G.B. Shimmield, C.P. Chilcott, M. Jebb, A.E. Fallick, and A.N. Dalgleish, Recent changes in climate in the far western equatorial Pacific and their relationship to the Southern Oscillation: Oxygen isotope records from massive corals, Papua New Guinea, *Earth Planet. Sci. Lett.*, 136, 575-590, 1995.
- Vautard, R., and M. Ghil, Singular spectrum analysis in nonlinear dynamics, with applications to paleoclimatic time series, *Physica D*, 35, 395-424, 1989.
- Vautard, R., P. Yiou, and M. Ghil, Singular-spectrum analysis: A toolkit for short, noisy chaotic signals, *Physica D*, 58, 95-126, 1992.
- Voss, R., R. Sausen, and U. Cubasch, Periodically synchronous coupled integrations with the atmosphere-ocean general circulation model ECHAM3/LSG. Part I: Simulations of the present day climate, *Clim. Dyn.*, 14, 249-266, 1998.
- Wallace, J.M., Y. Zhang, and K.H. Lau, Structure and seasonality of interannual and interdecadal variability of the geopotential height and temperature fields in the northern hemisphere troposphere, *J. Clim.* 11, 2063-2082, 1993.
- Weber, J.N., and P.M. Woodhead, Temperature dependence of oxygen-18 concentration in reef

- coral carbonates, *J. Geophys. Res.*, **77**, 463-473, 1972.
- Weber, J.N., P. Deines, P.H. Weber, and P.A. Baker, Depth-related changes in the C-13/C-12 ratio of skeletal carbonate deposited by Caribbean reef-frame building coral *Montastrea annularis*, *Geochim. Cosmochim. Acta*, **40**, 31-39, 1976.
- Wellington, G.M., R.B. Dunbar, and G. Merlen, Calibration of stable oxygen isotope signatures in Galápagos corals, *Paleoceanography*, **11**, 467-480, 1996.
- Winter, A., C. Goenaga, and G.A. Maul, Carbon and oxygen isotope time series from an 18 year Caribbean reef coral, *J. Geophys. Res.*, **96**, 16,673-16,678, 1991.
- Yiou, P., C. Genthon, M. Ghil, J. Jouzel, H. LeTreut, J.M. Barnola, C. Lorius, and Y.N. Korotkevitch, High-frequency paleovariability in climate and CO₂ levels from Vostok ice-core records, *J. Geophys. Res.*, **96**, 20,365-20,378, 1991.
- Zhang, Y., J.M. Wallace, and D.S. Battisti, ENSO-like decade-to-century scale variability: 1900-93, *J. Clim.*, **10**, 1004-1020, 1997.
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