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A Coral Window on Western Tropical Pacific Climate during the Pleistocene

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A CORAL WINDOW ON WESTERN TROPICAL PACIFIC CLIMATE DURING THE PLEISTOCENE

by

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A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science College of Marine Science University of South Florida

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ABSTRACT

Monthly $\delta^{18}O$ and Sr/Ca records generated from modern and fossil corals from Southwestern Pacific Ocean sites in the Republic of Vanuatu are used to assess the differences in mean climate state, seasonality, and interannual variability between a glacial and interglacial period.

The modern coral contains a well-defined annual signal in $\delta^{18}O$ and Sr/Ca. The top 40 cm of the coral used in this study has a mean $\delta^{18}O$ value of $-4.99\pm0.13\%e$ VPDB (2$\sigma$) and a mean Sr/Ca value of $8.691\pm0.015$ mmol/mol (2$\sigma$). El Niño–Southern Oscillation (ENSO) events are characterized by positive $\delta^{18}O$ and Sr/Ca anomalies, consistent with cooler temperatures and reduced rainfall that typifies ENSO at Vanuatu.

The ~12 cm long fossil coral is dated to 346 ka $\pm 25, - 9$, based on uranium-series analysis and stratigraphic forward modeling, indicating that the fossil coral grew during MIS10 – a glacial period. X-ray diffraction, petrographic inspection, SEM analysis, and geochemical considerations indicate excellent preservation. The mean $\delta^{18}O$ value is enriched by 0.74‰, and the mean Sr/Ca value is equivalent, compared to the modern coral. Mathematical modeling of Pleistocene mean SST and SSS results in temperature estimates up to ~2°C warmer and salinity up to ~2 psu saltier than present-day conditions, if seawater Sr/Ca were 1-2% higher in MIS10. Our fossil coral data and modeling results preclude colder SST and lower SSS at Vanuatu during MIS10. Accurate estimates of past values of seawater Sr/Ca remain the largest obstacle to accurately reconstructing past tropical SST using pristine fossil corals.

The fossil coral Sr/Ca annual range is similar to the modern range, indicating that seasonal SST ranges were similar, whereas the $\delta^{18}O$ annual range is about half that of the modern coral, indicating weaker past seasonal salinity variations. The reduced seasonal
SSS variations and increased SSTs near Vanuatu are interpreted as evidence that the SPCZ was displaced from its present location while the fossil coral lived.

The geochemical response to El Niño events in the modern coral is observed twice in the fossil coral record, indicating that ENSO–like processes are not unique to interglacial time periods, but characterize the tropical Pacific at least back to MIS 10.
1. INTRODUCTION

The El Niño – Southern Oscillation (ENSO) dominates interannual variations in the tropical climate system. During warm phase events (El Niño events), the thermocline in the western Pacific shoals while deepening in the eastern Pacific, causing warmer sea surface temperature (SST) anomalies in the eastern Pacific and cooler SST anomalies around the Western Pacific Warm Pool (WPWP; defined by the 28 °C mean SST isotherm). The South Pacific Convergence Zone (SPCZ; a region of atmospheric convergence in the South Western Pacific Ocean) also weakens and moves northward to merge with the Intertropical Convergence Zone (ITCZ, an equatorial region of atmospheric convergence) during El Niño events, which causes positive sea surface salinity (SSS) anomalies in the region normally beneath the SPCZ (Gouriou and Delcroix, in press). Climate variability associated with ENSO variations gives rise to significant rainfall and temperature anomalies that have large societal impacts. Intensive investigations for over 20 years concerning the ENSO phenomenon have yielded significant insights into the coupled ocean-atmosphere dynamics that give rise to this mode of climate variability in the modern and in the recent past.

Computer simulations of ENSO (Zebiak and Cane 1987; Schopf and Suarez, 1988, Battisti and Hirst, 1989) indicate that the amplitude, frequency, and regularity of ENSO events are sensitive to climate-related model parameters, suggesting that interannual variability is sensitive to background climate state (Enfield and Cid, 1991). A change in the frequency and intensity of El Niño events since the ‘climate regime shift’ of the mid-1970’s illustrated the possibility of a nonstationary ENSO system (Quinn and Neal, 1984; 1985; Nitta and Yamada, 1989; Trenberth and Hurrel 1994; Minobe 1997; 1999; An and Wang 2000; Stephans et al. 2001).

Evaluating modulations in ENSO during different background climate states requires records of El Niño events from times of different background climate. Instrumental records that could be used to reconstruct ENSO, such as SST in the eastern
equatorial Pacific Ocean or the Southern Oscillation Index, do not extend much further back in time than the end of the 19th century, if that far. Instrumental SST records of ENSO are more complete both spatially and temporally post-1950, and instrumental SSS records of ENSO are much shorter in length relative to the SST records. If available at all, SSS records extend back to the 1970’s (e.g., Lagerloef and Delcroix, 1999; Gouriou and Delcroix, in press).

Indirectly measured records of climate, often referred to as proxy records (e.g. tree-rings, ice cores, and corals; Dunbar and Cole 1999 for an overview), are required to investigate changes in ENSO over long time periods. Proxy records that overlap with, and extend beyond, the instrumental period offer a unique perspective on ENSO variability and can be used to address tropical climate variability on decadal to centennial and millennial time-scales. Proxy climate records also provide a data set from the tropical oceans that can be used to test model output and to provide constraints on climate model input.

Corals are ideal for reconstructing seasonal to interannual variations in tropical surface oceans because they are long–lived (many survive for centuries), their skeletons have annual density bands that provide chronological control, they are widely distributed, and their skeletons are amenable to geochemical analyses. Two of the most widely used geochemical determinations of coral aragonite are stable oxygen isotope ratios and Sr/Ca ratios. Coral oxygen isotopic ratios (\(\delta^{18}O\)) reflect both temperature and the oxygen isotopic composition of the seawater at the time of carbonate precipitation (\(\delta^{18}O_{\text{sw}}\); e.g., Epstein et al., 1953; Weber and Woodhead, 1972; Fairbanks et al., 1997). In the modern ocean, the oxygen isotopic composition of seawater responds to salinity changes brought about by evaporation or precipitation and, in coastal zones, fluvial input (Craig and Gordon, 1965; Bigg and Rohling, 2000). Sr/Ca ratios in the modern ocean are largely invariant and the Sr/Ca ratios in corals vary in response to temperature and the seawater Sr/Ca ratio (e.g., Beck et al., 1992). Paired analyses of Sr/Ca and \(\delta^{18}O\) on a single coral sample can result in the unique solution of SST and SSS (McCulloch et al., 1996; Gagan et al., 1998).

Paleoclimate records based on coral skeletons are particularly well suited to study the past and present behavior of ENSO. For example, Dunbar et al. (1994) used a coral
from the Galapagos Islands that extended back to 1587 to illustrate that spectral power within the ENSO frequency band shifted on decadal time scales and postulated that solar variability may modulate interannual and decadal climate variability in the tropics. In another study, two Holocene coral records (8920 yr BP and 7375 yr BP) indicated mean surface temperatures in Papua New Guinea were cooler than they are today (McCulloch et al., 1996). Higher Sr/Ca variance in the same fossil corals was interpreted as representing increased frequency of high amplitude ENSO events (McCulloch et al., 1996).

Oxygen isotopic variations in a last interglacial coral (~124 ka) from Indonesia exhibit behavior much like that observed in modern corals from the same area, indicating that ENSO activity at that time was similar to today (Hughen et al., 1999).

Modeling results suggest that tropical climate variability is orbitally controlled by seasonal insolation changes arising from the precession of the equinoxes (Clement et al., 1999; 2000; Kukla et al., 2002). Tudhope et al. (2001) used the record of geochemical variations in a suite of fossil corals from New Guinea, ranging in age from modern to 130 ka, to demonstrate that ENSO activity is a feature of both the interglacial and glacial ocean-atmosphere system. The corals from glacial times have weaker variance at the ENSO frequency bands in accordance with the model, however more paleoclimate data are needed to assess the predictions of the model with confidence.

Climatological interpretation of geochemical data from fossil coral samples must also account for any changes in seawater chemistry that occur on glacial-interglacial time scales. The stable oxygen isotopic composition of seawater is known to change in response to sea-level changes such that during glacial times the seawater $\delta^{18}O$ values increase as $^{16}$O is preferentially sequestered in ice sheets. The current best estimate for the change in seawater $\delta^{18}O$ during the last glacial maximum (18ka) is $1.0 \pm 0.1\%e$ (Schrag et al., 2002). Seawater Sr/Ca ratios, long thought invariant on glacial-interglacial time scales, likely vary in response to changes in sea level (Stoll and Schrag, 1998). These authors used a geochemical box model to demonstrate that the diagenetic alteration of Sr-rich shelf carbonates during sea-level low stands results in a flux of Sr to the global oceans that raises the seawater Sr/Ca ratio. Hence, glacial intervals of lower sea level have a higher seawater Sr/Ca ratio than do interglacial intervals.
Here we present $\delta^8$O and Sr/Ca data from an exceptionally preserved fossil coral drilled on Bougainville Guyot (16°00.56’S, 166°40.34’E) near Espiritu Santo Island in the Republic of Vanuatu (Fig. 1). The fossil coral is ~350 ka old, an age that places coral growth during deep-sea isotope stage 10 - a glacial interval. In this study, we compare geochemical variations in the fossil coral with those from a modern coral from near Espiritu Santo Island (15.7S, 167.2E; Fig. 1) to assess the differences and similarities in mean climate state, seasonality, and interannual variability between a glacial and interglacial period as recorded in coral $\delta^{18}$O and Sr/Ca.
2. OCEAN-ATMOSPHERIC INTERACTIONS AT VANUATU

Ocean-atmosphere interactions in the Vanuatu region of the South Pacific are dominated by spatial and temporal variability in the linkages between warm SST associated with the Western Pacific Warm Pool (WPWP) and atmospheric convection associated with the South Pacific Convergence Zone (SPCZ). This study will address this variability on intra-annual and interannual time scales in the modern and glacial ocean.

2.1. South Pacific Convergence Zone

The SPCZ is a zone of atmospheric convergence of the trade winds, and can be thought of as an extension of the ITCZ that stretches roughly from Papua New Guinea east to 120°W, 30°S (Fig. 2, Vincent, 1994). The SPCZ can be defined by several variables: sea-level pressure, convective activity, outgoing longwave radiation, or surface wind-field convergence. There are four leading hypotheses that explain the existence of the SPCZ (Vincent, 1994). One hypothesis states that SST gradients around the WPWP (Fig. 2) cause surface pressure gradients that drive low-level wind, resulting in moisture convergence. A second hypothesis states that SPCZ strength is determined by the heating and circulation patterns over Australia. Model results testing the second hypothesis indicate that removing Australia from the model weakened the strength of the SPCZ, although the SPCZ remained in the same location (Kiladis et al. 1989). The third hypothesis suggests that the strength of the SPCZ is dependent on forcing from the monsoon systems of the Southern Hemisphere, including the Australian Monsoon and Indonesian Monsoon (Davidson and Hendon, 1989). According to Vincent (1994), the most viable hypothesis is that both tropical and extra-tropical forcing affects the strength and location of the SPCZ in a variety of different ways. Surface temperatures, mid-latitude wave activity and tropical convection are all likely to be involved in SPCZ activity and variability.
The spatial extent of the SPCZ changes on seasonal, interannual and multidecadal time scales in response to changing wind fields and SST. The SPCZ is stronger in the austral summer than in the austral winter. Summertime warm SSTs facilitate upward convection in the atmosphere and northeasterly trade winds merge into the southeasterly trades along the SPCZ (Vincent, 1994). The combination of warm SSTs and atmospheric convergence are associated with low atmospheric pressure and a persistent cloud band over the SPCZ. In the austral winter, the Western Pacific Warm Pool (WPWP) shifts to the north as northern-hemisphere summer progresses. A shift in the wind field moves the focus of low-level tropospheric convergence to the ITCZ in the northern hemisphere in the austral winter (Vincent, 1994). The locus of convection moves northward with the warm water and the winds, resulting in the SPCZ’s apparent northward movement into the ITCZ.

An east-west oriented circulation cell over the equatorial Pacific Ocean is known as Walker circulation and is directly analogous to the north-south oriented Hadley circulation. Walker circulation occurs all along the equator (Bigg, 1996), but the circulation cell over the Pacific Ocean is most important to this study. The upward limb of the Pacific Ocean Walker circulation cell is the Indo-Pacific convective region, including the SPCZ, and the downward limb is the eastern equatorial region of high pressure. Connecting the low-pressure (upward limb) and high-pressure (downward limb) is the Trade Wind Belt that acts as the westward limb of the Pacific Walker circulation. Walker circulation is only one component affecting the strength of the trade winds; other components are also involved (e.g., Hadley Cell interactions with the earth’s rotation). When atmospheric convection decreases along the SPCZ, the above-described connections can cause weaker Walker circulation and a concomitant easterly trade wind anomaly. The surface expression of this type of variability is known as the Southern Oscillation. Thus interannual SPCZ variability is directly linked to ENSO.

The strength and position of the SPCZ also varies on multidecadal time scales according to the phase of the Interdecadal Pacific Oscillation (IPO; a Pacific-wide manifestation of the Pacific Decadal Oscillation; Folland et al., 2002). SPCZ movements in response to the phase of the IPO are quasi-independent of, but of the same magnitude as, SPCZ movements in response to ENSO (Folland et al., 2002). In general, a positive phase of
the IPO is related to a northeast shift in the SPCZ (the case from 1977-present), and a negative phase of the IPO is related to a southwest shift in the SPCZ (the situation from the mid-1940’s to the mid-1970’s). These types of movements become important when interpreting a climate proxy record because the proxy has been recording a Eulerian measurement of a moving system. In other words, both the corals remain in the same location but surface ocean currents and atmospheric phenomena may change their position with time.

2.2. Physical Oceanographic Setting

Ocean conditions in the western tropical Pacific both respond to and force atmospheric phenomena. The WPWP is the largest oceanic heat source in the world and strongly affects global climate on a variety of time scales, especially at interannual timescales (ENSO).

The WPWP changes shape and position seasonally depending on solar heating. Its center shifts towards the summer hemisphere such that it is in its most northerly position in August and September, and most southerly in January and February (Yan et al., 1997). Interannually, the WPWP shifts to the east because of anomalous eastward advection in El Niño years (Philander, 1990). SST in the warm pool region decrease slightly during warm events, but areas on the edge of the WPWP experience much greater temperature anomalies because of the strong thermal gradients surrounding the WPWP.

Low average SSS spatially correlates with the WPWP, where warm SSTs encourage convection and precipitation (Delcroix, 1998). A salinity minimum lies in a “V” pattern with one arm along about 5˚N associated with the ITCZ and another arm directed southeast along the SPCZ. In the crook of the “V” lies a strong salinity front (~175˚W) that advects eastward during ENSO events (Picaut et al., 1996; 1997; 2001; Vialard and Delecluse, 1998). Another strong salinity gradient lies on the southern edge of the WPWP, under the SPCZ. Salinity variations in this region are related to ENSO-related advection changes and E-P changes associated with SPCZ movements (Gouriou and Delcroix, in press).

In the southwestern tropical Pacific, the South Equatorial Current (SEC)
dominates surface-ocean circulation. The SEC is a westward flowing current that bifurcates down stream of ~150°W in the Pacific basin. The northernmost branch of the SEC is centered at 5°S while the southern branch of the SEC is centered at 15°S (Reverdin et al., 1994). The South Equatorial Counter Current (SECC), which is centered near 10°S, occupies the region between the two branches of the SEC on a seasonal basis. The SEC splits west of Vanuatu; the northern portion merges with northward flowing currents that link the Coral Sea with the Solomon Sea, and eventually merge with the Equatorial Counter Current (Qu and Lindstrom, 2002). The southward branch contributes to the East Australian Current (Qu and Lindstrom, 2002).

2.3. El Niño – Southern Oscillation

Atmospheric and oceanic changes during ENSO events have been alluded to above and require further explanation to set the stage for interpreting ENSO variability from coral geochemical records. During a warm phase event (negative values of the Southern Oscillation Index, SOI), an enormous number of interrelated changes occur in the Pacific that affect and are affected by SSS and SST. Walker circulation decreases, coincident with weaker convergence in the SPCZ and decreased precipitation under the SPCZ. The SPCZ axis shifts to the northeast (Folland et al., 2002) and salinity increases along the southern edge of the SPCZ by almost an order of magnitude more than on seasonal cycles. Drought conditions prevail, and the salinity front separating the fresh warm pool waters from the salty subtropical waters is displaced away from the region (Gouriou and Delcroix, in press). Eastward trade wind anomalies are also associated with decreased Walker circulation. The trade wind anomalies require that the tropical Pacific adjust its equilibrium state and equatorial upwelling is reduced all across the Pacific (Bigg, 1996). An equatorially trapped Kelvin wave displaces warm water from the WPWP across the Pacific in response to changes in the wind-stress field. Warm waters (from Kelvin waves and decreased upwelling) encourage convergence and precipitation in the central Pacific. Meanwhile the WPWP has lost heat energy, and local temperatures in the WPWP decrease. Temperature anomalies are strongest along the edges of the WPWP, where the thermal gradient is the largest. The net result is that the western Pacific in general is cooler and saltier during an El Niño event as warm water
and centers of precipitation move eastward.

Mean conditions in the tropical Pacific Ocean are also punctuated by events commonly called La Niña events or cold phase events because they are characterized by anomalies in the opposite direction from El Niño events (warm phase events). During a La Niña year Walker circulation increases and concomitantly, SPCZ convergence and rainfall increases. The westward trade winds are strong, and warm water ‘piles up’ in the western Pacific. Surface temperatures are slightly warmer than usual and salinity decreases under the convergence zones in response to enormous amounts of rain (e.g., meters per year).

2.4. Vanuatu as a Local Site to Monitor Regional Climate

Vanuatu is well situated to study changes in tropical climate variability because it is positioned beneath the SPCZ and along the southern edge of the WPWP, and is affected by spatial and thermal variations in the warm pool. Thus, ENSO forcing is strong at Vanuatu and surface-ocean conditions there reflect the influence of important climate forcing.

Sea-surface conditions in Vanuatu are characterized by a clear, 2-3˚C, seasonal temperature cycle, and a highly variable seasonal salinity cycle that is small compared with the interannual SSS variations. The average SST for a 1° x 1° grid box centered on 166.5E and 15.5S is 27.69˚C and the average salinity is 34.87, whereas the annual range in SST and SSS is 2.5˚C and 0.36psu, respectively (Levitus and Boyer, 1994). In an average year, SST and SSS changes are out of phase so that the maximum SST coincides with the minimum SSS (Fig. 3; Gouriou and Delcroix, in press; Levitus and Boyer, 1994). Inverse phasing between SST and SSS is significant because it produces a constructive signal in the oxygen isotope geochemistry of a coral skeleton. Movements of the SPCZ control seasonal SSS changes in this region (Gouriou and Delcroix, in press).

The SPCZ lies to the northwest of Vanuatu, towards the equator, during the austral winter. In the austral summer the SPCZ moves southeast over Vanuatu, bringing with it enormous amounts of rain that contributes to regional SSS anomalies (Gouriou and Delcroix, in press). The importance of rainfall in the region is evidenced by
precipitation data from Luganville, Vanuatu, located on the coastal plain just beside Malo Channel. The climatological range in precipitation is large, from ~ 100 mm per month to ~ 350 mm per month (Fig. 4). As explained above, heavy rainfall in the winter to early spring months is due to strong atmospheric convergence, and is correlated with low salinity in the region. Important regional patterns are also clear in the annual rainfall totals over the past half-century. El Niño and La Niña years can be differentiated as especially dry and wet years respectively, and the climate regime shift of the mid-1970s is readily discernable.

SSS and SST anomalies (SSSA and SSTA, respectively) in the Vanuatu region are strongly affected by ENSO activity. During warm phase (cool phase) years, the SPCZ weakens (strengthens) and Vanuatu is dry (wet). Salinity increases sharply during warm phases due to advection of salty water from south of the SPCZ and due to the local E-P balance as described above. SST’s decrease (increase) during warm phase (cool phase) years, though the deviation is not as large as with SSS. As with the seasonal cycle, interannual variations of SSS and SST are inversely related and cause an additive response in the coral δ¹⁸O signal. The maximum salinity in warm phase years is 0.28 psu saltier than the average salinity maximum for non-warm phase years for the period 1980-1993 (calculated with data from Gouriou and Delcroix, in press). This salinity value of 0.28 psu is 78% of the entire range of the annual SSS cycle. SST data (GISST2.36; Parker et al., 1995) can be used to demonstrate that winter minima in SST during warm phase years are 0.76˚C cooler than the minima in non-warm phase years for the period 1980-1993. The difference between normal years and warm phase years is 30% of the average annual range over the same period.

SST and SSS values quoted above give the correct sense of the response; however it is important to note that not all ENSO events manifest in the exact same manner at Vanuatu (Fig. 5). There are some warm phase years where a small SSTA is associated with a strong SSSA (1983), while other years have strong SSTA and weak SSSA (1987). Still other “events” last for multiple years, such as the mild El Niño of the early 1990’s (Trenberth and Hoar, 1996; 1997). The key to recognizing ENSO from temperature and salinity records in the Vanuatu region is determining the sense and duration of the SSSA and SSTA.
For the purpose of this study ENSO events must be defined in terms recognizable in a proxy record, faithful to the local modern instrument record, and consistent with conventional definitions. In a fashion similar to Trenberth (1997), the 5-month running mean of surface anomalies are used to define an ENSO event. A simultaneous SSTA decrease (increase) and SSSA increase (decrease), of a magnitude greater than one standard deviation of the running mean for more than 6 months of either SSSA or SSTA (δ¹⁸O or Sr/Ca anomalies in proxy situations), indicates a warm (cool) event (Fig. 6).

The definition above differs from the definition of Trenberth (1997) in two ways. Surface salinity anomalies are included in the definition in addition to SSTA because the local response to El Niño at Vanuatu can be either strong salinity anomalies or strong temperature anomalies, and is often both. Another difference is the use of a fractional standard deviation threshold as opposed to a specific temperature (0.4°C) as in Trenberth (1997). The fractional standard deviation threshold captures anomalies that are proportionally large compared to the variance in the specific record. When considering records of past SSTA and SSSA that may have a smaller or larger amount of variance, the standard deviation threshold is a more robust indicator of significant change relative to the mean.
3. METHODS

1.1. Physical Sampling of Corals

A living *Porites lutea* head was cored in 3 m of water within Malo Channel, which is located between Malo Island and Espiritu Santo Island in the Republic of Vanuatu (Fig. 1) in 1992. The site was chosen because of its exposure to open-ocean conditions, minimizing any local microclimate effects. A 5 mm-wide slab was removed from the core (henceforth referred to as MCB) and subjected to X-radiography, revealing distinct density band couplets (Fig. 7). The slab is composed of 8 individual pieces, which fit together well and ensure stratigraphic completeness. Samples of the coral slab were removed as powder using a computer controlled drilling stage and drill mount. Powder samples were generated by routing a 3 mm wide swath, ~1mm deep and collecting the powder every 1.33 mm down the slab.

The fossil *Porites sp.* sample was recovered from the carbonate sequence of Bougainville Guyot during ODP Leg 134 at Site 831B in 1992 (831B-18R-1, 110-125; Taylor et al., 1994). The sample (henceforth referred to as BG831) is 15 cm long and was recovered from 238.3m below the sea floor in 1066m water depth. Similarly to the modern coral, BG831 was cut into a slab 5 mm wide, X-radiographed to reveal density bands, and drilled for aragonite powder. Because the fossil coral grew more slowly than the modern coral, individual samples from BG831 were extracted about every 0.667 mm.

Representative samples from the modern and fossil coral were subjected to a variety of tests to determine the preservation state of the fossil coral. Initially BG831 was subjected to X-ray diffractometry to test for the presence of calcite. A thin section of the same coral was examined under a petrographic microscope and both MBC and BG831 were examined under a scanning electron microscope to look for secondary aragonite or calcite.

1.2. Age Determination of the Fossil Coral

The age of the fossil coral sample was determined by a combination of radiometric (uranium series and Sr isotopes) and stratigraphic modeling techniques. The
U-series determinations were performed at the University of Minnesota using thermal ionization mass spectrometry (TIMS) and the results have been previously published (Taylor et al., 1994). The Sr isotope determinations were performed at the University of Michigan radiogenic chemistry laboratory and the results have been previously published (Quinn et al., 1994).

Stratigraphic modeling was performed using a computer spreadsheet with the following input parameters: present-day depth of sample, subsidence rate (Taylor et al., 1994; F.W. Taylor, pers. comm., 2002) and Quaternary sea-level curves (SPECMAP Stack; Imbrie et al., 1984; and sea level from Shackleton 2000). Coral growth position errors of ±10m were used in all of the calculations. The elevation of the sample was calculated through time using the subsidence rate and the elevation of the sample in the sediments. The intersection of the sample elevation with a sea level curve dictates the age of the sample.

1.3. Stable Isotopes and Elemental Ratios

Stable isotopic determinations were made on coral powder samples at the University of South Florida College of Marine Science using a ThermoFinnigan Delta Plus XL dual inlet mass spectrometer with an attached Kiel Carbonate Preparation Device. Carbon and oxygen isotope data have a precision of 0.04 ‰ and 0.08 ‰ respectively (1σ), as calculated from NBS-19 standard material measured with every run.

Sr/Ca determinations were made on coral powder samples at the University of South Florida College of Marine Science using a Perkin Elmer Optima 4300DV dual view ICP-OES, following the drift-correction methods of Schrag (1999). Each coral sample (~250 mg) is dissolved in 10 ml of 2 % (v/v) trace metal-grade nitric acid diluted with super-deionized water (resistance equals 18 MΩ). Gravimetric standards were made by combining Mg, Ca, Sr, U and Ba from SPEX ultra-pure solution standards in proportions to match that of a typical coral. Bulk coral powder from a modern Porites lutea sample from Vanuatu was also dissolved in 2 % (v/v) trace metal-grade nitric acid to make a "coral standard" solution. The coral standard solution was routinely analyzed as part of each run. Blanks of acid and deionized water are routinely monitored for contamination. Analytical precision (2σ) for Sr/Ca determinations used in this study is
estimated to be 0.15 % (0.013 mmol/mol), based on 86 determinations of the coral standard solution.

Instrumental parameters of the ICP-OES are important for obtaining accurate results. The samples were introduced to the plasma at a rate of 1.0 ml/min using a Perkin Elmer AS93 plus auto sampler with a Meinhard TR-50 nebulizer and a cyclonic spray chamber. The sample probe was washed for 30 seconds in 2% trace metal-grade nitric acid between samples. Concentrations of strontium and calcium for the Sr/Ca ratios were measured on the ICP-OES by the intensity of two spectral lines: Sr (421.552 nm) and Ca (422.673 nm). The Perkin Elmer Optima 4300DV has two different detectors, each with its own high frequency noise (time period of seconds). The spectral lines used in this study were chosen because they are both measured on the same detector, thus minimizing the effect of high frequency noise.

Geochemical data were converted from the depth domain to the time domain by matching the maxima (minima) in Sr/Ca to minima (maxima) in SST data and linearly interpolating between, using the software package AnalySeries (Paillard et al., 1996). In the modern coral, SST data as extracted from a 1° by 1° grid box (GISST 2.36; Parker et al., 1995) was matched to the coral Sr/Ca, and the first tie point on the top of the core corresponds to the sampling date. For BG831, the monthly climatological average SST served as a substitute for monthly SST data. Sr/Ca was chosen over δ¹⁸O to create the age model because coral δ¹⁸O is affected by seawater temperature and δ¹⁸O variations, whereas coral Sr/Ca ratios are only affected by seawater temperature variations on short time scales. Checking the age models visually against annual banding evident in the X-radiographs ensured accuracy of our depth to time conversions.
2. RESULTS

4.1. Age of BG831

Sample BG831 (831B-18R-1, 110-125; 238.3 meters below seafloor; mbsf) has been dated using uranium-series techniques with thermal ionization mass spectrometry (Taylor et al., 1994). Sample BG831 has a $^{230}$Th age of 450 ka ± 20, a $^{234}$U/$^{238}$U age of 612 ka ± 15 and initial $\delta^{234}$U value of 96 ±7. The lack of concordance between the $^{230}$Th and $^{234}$U/$^{238}$U ages is probably caused by leakage of $^{234}$U related to alpha-recoil processes (Edwards et al., 1991). Alpha-recoil processes would yield calculated ages that are older than true ages so that the calculated ages of sample BG831 should be considered maximum ages (Taylor et al., 1994). The initial $\delta^{234}$U value of 96 ±7 for BG831 is substantially lower than the initial $\delta^{234}$U values modern coral samples (~150±10; Edwards et al., 1986; Bar-Matthews et al., 1993; Henderson et al., 1993) and indicative of open-system behavior with respect to uranium. The non-retention of modern initial $\delta^{234}$U values is a sign of diagenetic alteration of the uranium system in a coral (Bar-Matthews et al., 1993; Stirling et al., 1995).

Additional constraints on the age of sample BG831 comes from Sr isotopic and uranium-series analyses on other coral samples recovered in the core. A coral sample from 59 cm deeper in the core than sample BG831 (831B-18R-2, 44-46; 238.89 mbsf) has an apparent age of 270 ka, a minimum age of 180 ka and a maximum age of 370 ka based on Sr isotopes (Quinn et al., 1994). A coral sample from 85.22 m deeper in the core than sample BG831 (831B-27R-CC, 12-14; 323.52 mbsf) has nearly concordant ages of 371 ± 9 ka ($^{230}$Th) and 393 ± 11 ka ($^{234}$U/$^{238}$U) and an initial $\delta^{234}$U value of 159 ± 7. These data on other coral samples from Hole 831B constrain the age of sample BG831 to be younger than ~ 400 ka.

Stratigraphic forward modeling was performed to further constrain the age of sample BG831. The model predicts the age of a coral sample based on its stratigraphic
position in a core (1304.3 meters below sea level, mbsl; Taylor et al., 1994), tectonic subsidence rate from the coral sample with concordant ages (3.5 ±0.2,-0.1 mm/yr; Taylor et al., 1994) and Pleistocene sea-level history (SPECMAP Stack, Imbrie et al., 1984). Model results indicate that most likely age of coral sample BG831 is 346 ka + 25, - 9 (Fig. 8).

4.2. Assessment of Diagenetic Alteration

Stratigraphic forward modeling can be used to demonstrate that coral sample BG831 was quickly submerged below the range of sea-level fluctuations shortly after deposition (Fig. 8). Hence, sample BG831 has never been subaerially exposed nor has it been subjected to the ravages of freshwater diagenesis. Any post-depositional alteration of sample BG831 is restricted to those processes attendant with residence in marine waters and pore fluids.

Sample BG831 was examined for physical evidence of diagenetic alteration. X-ray diffraction analyses revealed only the presence of aragonite; no calcite peaks were identified, as expected from an originally aragonite sample that has never been subaerially exposed. Scanning electron microscopy (SEM) revealed no secondary cements (aragonite or calcite) in sample BG831. SEM images of a modern Porites coral sample are strikingly similar to those of the fossil coral sample BG831 (Fig. 9). Examination of a petrographic thin section of sample BG831 documents similar preservation between modern and fossil coral, and neither secondary cements nor recrystallization are discernable (Fig. 10).

The low initial $\delta^{234}$U value of sample BG81 with respect to a modern coral is consistent with diagenetic alteration of the uranium system. However, uranium is an easily perturbed system, and open system behavior of uranium does not in and of itself imply that all other isotopic systems are perturbed (i.e., uranium isotopes can be open and strontium can be closed; Bar-Matthews et al., 1993). Indeed, the Sr/Ca, oxygen isotope, and carbon isotope data themselves provide good evidence that the coral has not undergone significant diagenesis. Diagenetic alteration of coral aragonite causes a sharp decrease in Sr/Ca values (towards 1-2mmol/mol from 8-9mmol/mol), obliterating the relatively small (~0.1 mmol/mol) seasonal variations (Enmar et al., 2000; McGregor and
Gagan, in press). The annual temperature cycles visible in the Sr/Ca data are one of the first things to deteriorate in a diagenetically altered coral (Enmar et al., 2000; McGregor and Gagan, in press); therefore the retention of annual cyclicity in Sr/Ca ratios in the fossil coral is consistent with preservation of a primary geochemical signal. Additionally, meteoric diagenesis in both open and closed systems causes large (several per mil) perturbations in the oxygen and carbon isotopes that would decimate the relatively small amplitude annual cycles found in this study (Fig. 11; Quinn et al., 1991). Similarly, inorganic aragonite or high-magnesium calcite precipitation from marine waters would have increased the carbon isotope values by several per mil (James and Choquette, 1983), but this type of isotopic behavior is not seen in the carbon isotope data (Fig. 11).

4.3. Geochemical Data

The stable isotope ratios and Sr/Ca ratios used in this study were obtained for the entire 12 cm length of BG831 and the first 26 cm of MCB. The raw data for both samples are shown in Fig. 11; they include oxygen isotopes, carbon isotopes and Sr/Ca ratios. The data are plotted in both the depth domain and the time domain to illustrate that the data are not fundamentally changed by the conversion to time. All subsequent analyses utilize the time series. The data from BG831 and MCB both show clear annual cycles in all of the measured geochemical variables. BG831 grew at an average rate of 9.8 mm/year and MCB grew at 20.0 mm/year. All significance tests and confidence intervals were calculated with same the number of degrees of freedom as years in the records. This assumes that there is one independent data point per year in these serially correlated records.

BG831 has a mean oxygen isotope ratio of $-4.25 \pm 0.07$ ‰ VPDB (2σ, Vienna Pee Dee Belemnite), whereas the mean MCB value is $-4.99 \pm 0.13$ ‰ VPDB (2σ, Fig. 12). The two means are different at greater than the 99.9% confidence level. A check of the age model is possible by comparing difference between the modern and fossil coral mean $\delta^{18}O$ value. If the entire 0.74‰ offset were only due to salinity changes and not ice volume affects, then the salinity of the area would have to be 37.7 (a 2.7 psu offset using the $\delta^{18}O$-SSS relationship of Fairbanks, 1997). This is exceptionally high for an open
ocean setting, therefore is taken as evidence that BG831 grew when glacial ice was more extensive than today, i.e. during a glacial period. The annual δ¹⁸O range averages 0.27 ± 0.09‰ in BG831, compared to a value of 0.45 ± 0.23 ‰ in MCB (Fig. 13).

Sr/Ca ratios are similar in both modern and fossil corals (Fig. 11b). The mean Sr/Ca ratios for BG831 and MCB are 8.712 ± 0.017 mmol/mol and 8.691 ± 0.015 mmol/mol respectively (1σ). These mean values are the same within a single standard deviation. With the variance of these records, there would have to be 135 degrees of freedom in the data before the two means were significantly different at the 95% confidence level, many more degrees of freedom than are available from a ~12-year quasi-sinusoidal curve. The mean annual cycle in the two Sr/Ca records is also very similar (Fig. 13b). The 95% confidence limits overlap and differences between the curves are mostly within analytical error. The average annual Sr/Ca range is 0.13 ± 0.04 mmol/mol and 0.12 ± 0.03 mmol/mol (2σ) for BG831 and MCB, respectively.
5. DISCUSSION

5.1. Modern Proxies and Instrumental Data

5.1.1. Calibration of SST and Coral Sr/Ca

Climate proxies must be calibrated and validated against the instrumental record of climate change before they can be applied with confidence to paleoclimate questions (e.g., Crowley et al., 1999). There is no in situ, instrumental SST record at Vanuatu, therefore a global SST product, which integrates shipboard and satellite SST data and produces SST estimates at 1° latitude by 1° longitude grid points, must be used. The GISST2.36 database (Parker et al., 1995) and the NCEP reanalysis database (Reynolds and Smith, 1994) have been used in this study.

The relationship between coral Sr/Ca and SST at Vanuatu was calibrated using a reduced major axis (RMA) linear regression (Davis, 1986) on data from 1980-1990. The most recent few years of the core were not used to minimize any contamination from coral tissue at the top of the core (Fig. 15). RMA regression assumes equivalent error in both the dependent and independent variable, in contrast to ordinary least squares regression that assumes error only in the dependent variable (e.g., Davis, 1986).

Calibration of MCB coral Sr/Ca measurements against the GISST and NCEP SST records produces the following equations (95% confidence on the slope and intercept):

\[
\text{Sr/Ca} = 9.9626(\pm 0.0066) - 0.0462(\pm 0.0046) \cdot T, r = -0.83 \quad \text{(GISST)}
\]

\[
\text{Sr/Ca} = 9.8980(\pm 0.0059) - 0.0446(\pm 0.0044) \cdot T, r = -0.84 \quad \text{(NCEP)}.
\]

The calibration slopes are not statistically different, and the minor difference in slopes can result in temperature errors < 0.5°C at 25°C. The difference in intercept values can be explained by the fact that the temperature data sets are nearly constantly offset from each other.
The slope of the Porites spp. Sr/Ca-SST relationship in equations 1 and 2 are less than the 0.062±0.014 (2σ) slope value that Gagan et al. (2000) recently reported as the average slope value of 9 coral Sr/Ca-SST calibrations. Much debate has focused on the origin of the observed variability in the Porites Sr/Ca-SST relationships and many factors have been cited as possible causative agents including: differences in field and laboratory sampling procedures, inter-laboratory differences in Sr/Ca spike for TIMS analyses, inter-laboratory differences in standard determination, seawater Sr/Ca variability, calcification rate, symbiont activity, and lack of standardization of instrumental SSTs among calibration sites (de Villiers et al., 1995; Gagan et al., 2000; Cohen et al., 2002; Marshall and McCulloch 2002; Quinn and Sampson, 2002).

A key for coral paleoclimate reconstructions is that there is a robust, quantifiable relationship between in situ SST variations and those recorded in the gridded SST database. If this is true, then the corals recording local SST changes will correlate well with regional SST, as is the case in this study (Figs. 14, 15). Regional SST variations are ultimately the goal in paleoclimate reconstruction, thus this method is justified.

5.1.2. Relationship between Coral δ¹⁸O, SST and SSS

The δ¹⁸O composition of coralline aragonite is a function of seawater temperature and the δ¹⁸O of seawater at the time of skeletal precipitation (e.g., Goreau, 1977; Weber and Woodhead, 1972; McConnaughey, 1989). In oceanic regions characterized by strong atmospheric convection, like the tropical western Pacific, seawater δ¹⁸O values are highly correlated with salinity (Fairbanks et al., 1997). Coral δ¹⁸O variations at Vanuatu also reflect the combined influences of SST and SSS-driven changes in seawater δ¹⁸O (Quinn et al., 1993; this study). Separate regressions for both δ¹⁸O-SST and δ¹⁸O-SSS are justified because the monthly resolved instrumental records of SST and SSS that are used in the regression analyses are not significantly correlated (r=-0.0086). This is not to say that SSS and SST at Vanuatu have no relationship, they are quite correlated on interannual time scales, and therefore future work on this core should include a simultaneously solved regression for both SST and SSS.
The relationship between coral $\delta^{18}O$ and SST at Vanuatu is defined using RMA regression, producing the following equations (95% confidence on the slope and intercept):

$$\delta^{18}O = 1.461(\pm 0.065) - 0.232(\pm 0.033) \cdot T, \ r = -0.63 \text{ (GISST)} \quad (3)$$

$$\delta^{18}O = 1.137(\pm 0.056) - 0.224(\pm 0.030) \cdot T, \ r = -0.67 \text{ (NCEP).} \quad (4)$$

The correlation coefficients are significant at the 99.9% confidence level ($p<0.0005$), but they are not as high as the Sr/Ca-SST correlations. The monthly $\delta^{18}O$ signal is strongly forced by temperature, but the correlation diminishes from the Sr/Ca-SST correlation because SSS also affects coral $\delta^{18}O$ values.

Local records of SSS variations in remote areas of the tropical oceans are scarce and so are gridded SSS databases. A gridded SSS database for parts of the western Pacific (including Vanuatu) has recently been developed using ship-of-opportunity salinity measurements (Gouriou and Delcroix, in press). SSS data extracted from the Vanuatu grid point of the Gouriou and Delcroix (in press) database can be compared with coral $\delta^{18}O$ data to define a coral $\delta^{18}O$-SSS relationship using RMA regression. The following equation is produced (95% confidence on the slope and intercept):

$$\delta^{18}O = 34.96(\pm 0.29) - 0.86(\pm 0.12) \cdot S, \ r = 0.65. \quad (5)$$

The correlation coefficient is significantly different from zero at the 99.9% confidence level ($p<0.0005$), and is similar in magnitude to the $\delta^{18}O$-SST correlation coefficient.

A comparison of the interannual $\delta^{18}O$-SST relationship with the $\delta^{18}O$-SSS relationship reveals that coral $\delta^{18}O$ is nearly equally forced by SST and SSS at both monthly and interannual time scales. All of the following correlation coefficients are significantly different from zero with the confidence level indicated in parentheses. The correlation coefficients on the monthly data are -0.67 (>98%, NCEP), -0.63 (>98%, GISST) and 0.65 (>98%, SSS). Similarly, the correlation coefficients for monthly anomaly data are –0.34 (>75%, GISST) and 0.41 (>65%, SSS), and the correlation coefficients for smoothed (5-month boxcar filter) monthly anomalies are -0.56 (>95%, GISST) and 0.65 (>90%, SSS). Covariance between SST and SSS does not cause the similar correlations between the two environmental variables and $\delta^{18}O$. Monthly SSS and
SST (GISST) data have a correlation coefficient that is not significantly different from zero (r = –0.0086).

5.1.3. Skeletal $\delta^{18}$O and Sr/Ca Variations and ENSO at Vanuatu

The response of the climate proxies (coral $\delta^{18}$O and Sr/Ca) to seasonal SST and SSS forcing has been defined, but the relationship of these climate proxies to interannual forcing, especially ENSO forcing, must also be demonstrated in the modern record before interannual relationships in the fossil record can be considered. Interannual climate variability in the tropical Pacific Ocean is dominated by ENSO variations. Warm (cold) events are manifest in this region as cool (warm) and dry (wet) anomalies (Delcroix, 1998). Corals respond to lower (higher) SST and higher (lower) SSS values attendant with a warm (cool) phase by precipitating skeletal material that has higher (lower) $\delta^{18}$O and Sr/Ca values.

The $\delta^{18}$O and Sr/Ca anomaly plots from MCB are very similar to the SSSA and SSTA plots from Vanuatu (Fig. 16, A and B). Coral $\delta^{18}$O and Sr/Ca both increase during the 1983, 1987, and 1991 El Niño events, and they decrease in the 1989 La Niña event. Additionally, modern coral geochemistry provides details about the characteristics of each event that we know to be true from the instrumental record. The 1983 warm phase event was characterized in the region by a large salinity anomaly and a small temperature anomaly. Salinity in the Western Pacific lagged behind the eastern Pacific warming (represented by gray bars in Fig. 15, A and B) during this event. The timing and relative magnitude of the temperature versus salinity anomalies in the proxy data are perfectly in line with the instrumental record (Fig. 16, A and B). The 1987 ENSO event manifest in this region as large temperature anomalies and much smaller salinity anomalies. Again, the proxy record shows this detail (Fig. 16, B). By using the definition of ENSO events described in section 2.2, one can easily identify modern ENSO events in the geochemical proxy records, and even further describe details about the temperature and salinity anomalies associated with these events.
5.2. Pleistocene Interannual Variability

Recent non-stationary behavior of the ENSO system in the late 20th century (Wang and An, 2001, and references therein) has focused attention on the possible affects of changing boundary conditions (e.g., CO₂, insolation, and temperature) on interannual variability. This has lead to a series of studies investigating changes in ENSO during the Pleistocene and early Holocene (e.g., Clement et al., 1999; Hughen et al., 1999; McCulloch et al., 1999; Rittenour et al., 2000; Tudhope et al., 2001; Kukla et al., 2002). Some initial findings from sediment cores and fossil mollusks from archaeological sites indicated that interannual ENSO variability was a phenomenon of the middle to late Holocene and did not exist in the Late Pleistocene (Sandweiss et al., 1997; Rodbell et al., 1999). However, data from fossil corals appear to indicate that ENSO variations were similar to recent (19th, 20th century) variations during interglacial periods and that ENSO was weaker but still active during glacial periods (Hughen et al., 1999; Tudhope et al., 2001). Tudhope et al. (2001) show that ENSO has existed back to at least 130±2ka.

The mechanisms driving ENSO variability on geologic time scales are still enigmatic (Clement et al., 1999). Simplified coupled ocean-atmosphere general circulation models have been used to investigate some of the possible processes involved. Climate modeling indicates that the precessional cycle of insolation can interact with zonal asymmetry in Pacific SST and cause an increased frequency of warm (cool) events when there is an anomalous cooling (warming) in the late summer/fall (Clement et al., 1999). Another similar model indicates that an increase in warm events during early glaciation may induce high latitude-ice buildup (Kukla, et al., 2002). The paleo data that exist are insufficient to adequately verify model results, and more paleo-data are needed.

The Sr/Ca and δ¹⁸O time series in the fossil coral from Vanuatu contain two geochemical excursions that are consistent with our definition of ENSO events as recorded in the geochemistry of a modern coral (see section 2.2; Figs. 16, 17). The first “paleo-ENSO” event is primarily manifest as a temperature anomaly that lasts for essentially 16 months. The positive Sr/Ca anomaly reaches the 1.5-σ threshold criterion in June of year 2, and continues through September of year 3, with a 4-month period in between where the anomaly remains positive, but drops below the threshold criterion of 1.5 σ. Meanwhile, the δ¹⁸O anomaly does not increase until September of year 2, and
remains positive only until March of year 3. The maximum SSTA is -1.7°C using the -
0.0462 mmol/mol°C⁻¹ relationship (Equation 1).

The second “paleo-ENSO” event lasts 17 months and strictly speaking should be
split into 2 events by our definition. Both proxies have anomalies during year 6 and
again in year 7. The event is primarily a δ¹⁸O anomaly, indicating that salinity deviations
dominated this event. δ¹⁸O reaches the 1.5 σ threshold criterion in July of year 6, drops
below the threshold criterion in December of the same year, and then bounces back again
in June through November of the following year. The Sr/Ca anomalies are minimal over
this time, but they do show minor positive anomalies associated with the larger δ¹⁸O
positive anomalies.

These two climate events—interpreted as “paleo-ENSO” events—appear to
agree with the findings of Tudhope et al. (2001) who reported decreased interannual
variability in coral δ¹⁸O during glacial periods at Papua New Guinea. The absolute
amplitudes of the δ¹⁸O and Sr/Ca anomalies in BG831 are smaller than δ¹⁸O and Sr/Ca
anomalies observed in the modern coral. However, as a percentage of the average
seasonal range, the anomalies are very similar to modern events. BG831 is too short of a
record to divulge significant information about the frequency of events and therefore
cannot be used to rigorously test the theory that ENSO is controlled by the ~20kyr
precessional cycle (Clement et al., 1999).

It is interesting to note that northern hemisphere summer insolation (65°N, July)
exhibits minimal variations at this time and ranges between ~430 and ~470Wm⁻² (Laskar
1990). Additionally, seasonality in equatorial insolation (March minus September) is
also at a minimum and is similar to that found between 80ka and the last glacial
maximum. Seasonal variations in insolation drive the mechanism that may control
ENSO variability on long time scales (described by Clement et al., 1999) and one would
expect lower frequency of El Niño events at this time.

5.3. Modeling the Pleistocene Mean Climate State

Modern climate reconstructions essentially utilize the relationships between
Sr/Ca, δ¹⁸O, salinity, and temperature to solve two equations for two unknowns. Sr/Ca is
used to calculate temperature, which is then input into the δ¹⁸O-temperature equation,
theoretically yielding $\delta^{18}O_{sw}$, a variable directly related to salinity. However, over geologic time scales these relationships are not so straightforward.

Unlike modern climate reconstructions, changes in seawater chemistry must be accounted for when interpreting Pleistocene coral records. Over glacial/interglacial time scales, the oxygen isotopic composition of seawater changes with ice volume due to the effects of preferential evaporation and sequestration of $^{16}$O in glacial ice. This effect is well documented and the global average is 1.0±0.1 %o for 120 m (sea-level equivalent) of ice build-up (Schrag et al., 2002). The Sr/Ca ratio of the ocean is also thought to change as a function of ice volume because a low sea level exposes Sr-rich shelf carbonates to meteoric dissolution and recrystallization (Stoll and Schrag, 1998). In contrast to $\delta^{18}O_{sw}$, seawater Sr/Ca variations are not well known. Model results of Stoll and Schrag (1998) indicate that seawater Sr/Ca may have increased by about 3% during glacial periods. Some data agree with this model output, but more data are needed to verify the predictions of the Stoll and Schrag (1998) model (Stoll et al., 1999).

A simple mathematical, mass-balance model has been developed to investigate the effects of changing ocean chemistry on fossil coral Sr/Ca and $\delta^{18}O$. The following two equations were simultaneously solved under variable seawater Sr/Ca condition sand variable ice volume inputs:

$$\Delta Sr = \frac{\partial Sr}{\partial T} \cdot \Delta T + \frac{\partial Sr}{\partial I} \cdot \Delta I$$  \hspace{1cm} (6)

$$\Delta O = \frac{\partial O}{\partial T} \cdot \Delta T + \frac{\partial O}{\partial I} \cdot \Delta I + \frac{\partial O}{\partial S} \cdot \Delta S$$  \hspace{1cm} (7)

Where $\Delta Sr$ is the difference between the modern Sr/Ca and fossil Sr/Ca ratios (fossil minus modern), and $\Delta O$ is similarly the difference between modern and fossil $\delta^{18}O$. The variables $\Delta T$, $\Delta I$, and $\Delta S$ are changes in temperature, ice volume, and salinity respectively, from modern to ancient times. The partial derivative terms are empirically determined and the values used are listed in Table 1.

Two constraints on the equations are essential for limiting the domain and range of the equations. First, we know that the maximum change in ice volume over glacial/interglacial cycles is equivalent to about 120m of sea level and that we are currently at a sea-level highstand (ice-volume low). Sea level can be further constrained
from the age of the sample. Results of stratigraphic forward modeling suggest that BG831 grew while sea level was between 98m and 50m below present sea level (Fig. 8). Thus, the $\Delta I$ term is constrained to between 50m and 98m. Secondly, seawater Sr/Ca could not theoretically change more than 5% in any glacial/interglacial cycle and is more likely to be around 1-3% (Stoll et al., 1999). The mathematical model was run several times, each time the $\partial \text{Sr}/\partial I$ term was held at different values.

The full range of possible model results of SST and SSS are illustrated graphically in Fig. 18. Straight lines in Fig. 18 denote different values for the $\partial \text{Sr}/\partial I$ term in the model, the boundaries of which are limited by maximal and minimal ice volumes. All of the lines converge on a single point that represents seawater $\delta^{18}$O and Sr/Ca values the same as today (i.e. no change in ice volume). Increasing the amount of Sr/Ca change on glacial-interglacial cycles causes an increase in the temperature uncertainty, whereas decreasing the amount of Sr/Ca change on glacial-interglacial cycles causes an increase in the uncertainty of salinity.

The model output (Fig. 18) must be interpreted in light of other information to determine a mostly likely “paleoclimate solution” (gray shading in Fig. 18) from the full range of “model” possibilities. Thermodynamic processes limit mean annual SST to no more than about 30˚C (Ramanathan and Collins, 1991; Cubukcu and Krishnamurti, 2002). Since the modern mean annual SST at Vanuatu is 27.69˚C, the maximum SST increase possible is judged to be 2.31˚C. Similarly, modern average salinity is 35 and it is difficult to imagine open-ocean salinity greater than 37.5 when the highest average annual salinity in the open Pacific is currently about 36.5 (Levitus and Boyer, 1994). A third limit on the data comes with the seawater Sr/Ca term. While it is possible that Sr/Ca does not change on glacial/interglacial time scales, convincing data and modeling have been put forth by Stoll et al. (1999) that document at least some glacial/interglacial change in Sr/Ca. Therefore, we will assume at least 1% change on glacial/interglacial time scales. With these limits on the model, data from fossil coral BG831 indicates that the surface ocean was 1-2˚C warmer and 1-2 psu saltier than today during glacial stage 10.

Of all the three limits, seawater Sr/Ca is the least well known and contributes the most to the uncertainty in the SST and SSS determinations. As an example, if sea level
were known to be 120m below present, then there would still be an uncertainty of 7 °C and almost 4 psu simply from not knowing if Sr/Ca was stable or varied up to 3%. Constraining SST to a thermodynamic upper limit does limit the uncertainty in this case, but if the same study were carried out in a region with lower mean SST, then the thermodynamic limit would not constrain the possible SST and SSS outcomes. This demonstrates that accurate estimates of past values of seawater Sr/Ca remain the largest obstacle to the accurate reconstruction of tropical SST in the past using pristine fossil corals.

Removing the limits on seawater Sr/Ca, and maintaining the 30 °C limit, results in an SST range of 0 to 2.3 °C and SSS range of –0.3 to 2.5. Thus one can say with good certainty that sea surface conditions around 350ka before present were not likely to have been colder or fresher than present-day conditions. The result that SSS and SST values similar or increased relative to today while this coral grew holds true even if one ignores the results of the age model and assumes that this coral grew during an interglacial time.

Propagating the error in Equations 6 and 7 according to Beers (1957; equation 37), shows that the SST and SSS values are robust within the above stated assumptions (30°C max, 50-98m of ice, 1-2% Sr/Ca change, SSS max 37.5). The exact magnitude of the standard error shown in Table 2 is dependent on the ice volume (ΔI) and temperature (ΔT) derived from the model. Apart from the basic assumptions, the largest sources of error are the lengths of the coral geochemical records, specifically the variance in the mean value of each record. The second largest source of error is the ∂Sr/∂T relationship. Essentially, the more sensitive the paleo-thermometer, the greater the error associated with it.

Equation (7) assumes that δ¹⁸O_sw is only a function of global ice volume and local salinity (due to mixing with meteoric waters), but there are other possibilities. Another way one could change the relationship of δ¹⁸O_sw and SSS is to change the isotopic composition of meteoric waters by either changing the source location for local precipitation or by changing the temperature of that source location (Charles et al., 2001). A model of the isotopic composition of precipitation during glacial times (forced with both CLIMAP SST and with cooler tropical SST fields) indicates that the δ¹⁸O of precipitation around Vanuatu would increase slightly during glacial times. Similarly, the
δ¹⁸O of precipitation in the eastern tropical Pacific would also generally increase, and that might increase the isotopic composition of the South Equatorial Current, which bathes Vanuatu today. Tropical enrichment of precipitation is thought to be driven by rainfall variability because lower surface temperatures correspond to less intense rainfall (Charles et al., 2000). Local changes in the intensity of rainfall alone, without changing source location of precipitation, may cause an increase in the isotopic composition of that rainfall, thereby amplifying the isotopic signature recorded in the coral. The coral would record an isotopic enrichment due to increased salinity and increased isotopic composition of rainfall.

Lastly, the question of how representative is a 13-year record of SST relative to multidecadal to century long records of SST needs to be addressed. This issue is addressed in two ways. First, thirteen year moving averages of monthly Sr/Ca and δ¹⁸O data from the modern Vanuatu coral (1928-1992) were used to demonstrate that these mean values were never outside of the 95% confidence level for mean values calculated using the full records (8.66 ± 0.03 mmol/mol and –4.85 ± 0.12 for Sr/Ca and δ¹⁸O respectively. Second, thirteen year moving averages of monthly SST data for the 20th century were used to demonstrate that this proxy-based conclusion is also robust when instrumental data are used. The mean value over the same time period was 27.57±0.57°C and similarly to the results from the proxy record, the minimum and maximum 13-year mean were well within the error on the total mean value (27.27°C and 27.79°C for minimum and maximum respectively). Thus a thirteen year long record is representative of multidecadal to century long records of SST and the mean value is not likely to be strongly affected by higher frequency oscillations.

5.1. Possible Effects of Diagenesis on the Climate Interpretation

The lack of concordance between the ²³⁰Th and ²³¹Pa ages and the disturbed initial δ²³⁴U value (Taylor et al., 1994) is consistent with diagenetic perturbation of the U-series elements. The fossil coral sample has been evaluated for diagenesis using petrography, mineralogy and geochemistry. Petrographic thin sections and SEM images have been examined and no evidence of secondary carbonates has been observed (Fig. 7, 9). X-ray
diffraction results indicate 100% aragonite mineralogy. The sample has never been subaerially exposed nor exposed to the ravages of freshwater diagenesis.

The Sr/Ca record can provide additional evidence that the sample is not diagenetically altered. Three types of diagenetic material are common in corals: secondary aragonite, secondary calcite, and aragonite that has recrystallized into calcite. Secondary calcite is a meteoric diagenetic cement. It commonly fills the pores of subaerially exposed corals with calcite spar. Sr/Ca ratios in secondary calcite cements are much lower than skeletal aragonite (2.1 mmol/mol compared with ~8.8 mmol/mol in coral skeletons; McGregor and Gagan, in press). This is because the partition coefficient of Sr into calcite is much lower than for aragonite (Kinsman and Holland, 1969).

McGregor and Gagan (in press) calculated that the apparent change in temperature per percent calcite in a coral skeleton is 1.15± 0.03°C. Thus small amounts of secondary calcite cause large positive temperature errors. Patchy distribution of small amounts of calcite can cause high frequency variability that is visible at high sampling resolutions (near weekly; McGregor and Gagan, in press). BG831 was not likely to have ever been subaerially exposed, does not appear to have calcite, and does not display large high frequency variability when sampled at high resolution. BG831 shows no sign of being affected by secondary calcite.

Skeletal recrystallization from aragonite to calcite is another diagenetic pattern that does not appear in BG831. The partition coefficient of Sr in aragonite to calcite transitions has been reported to be 0.05 (Katz et al., 1972). Most of the Sr in the aragonite is lost in the transformation to calcite. If even 2 % of a coral (Sr/Ca=8.9 mmol/mol) turns into calcite (Sr/Ca=0.445 mmol/mol, using 0.05 from Katz et al., 1972), then the bulk coral Sr/Ca ratio would be 8.731. There is a 0.169 mmol/mol difference between the two numbers. That translates into a 3.7°C error using the Sr/Ca-SST calibration from this study, and 2.7°C using the Sr/Ca-SST calibration of Beck et al. (1997). Even a small amount of recrystallization causes large temperature anomalies. It is theoretically possible that a small amount (<1%) of recrystallization occurred uniformly in the skeleton of BG831, and it would be possible to miss it with X-ray diffraction. One would not realize the temperatures calculated from the coral were about 1°C too high. McGregor and Gagan’s (in press) argument for the Sr/Ca pattern of patchy
Secondary calcite can be applied to negate the above scenario. Recrystallization can also be patchy and the coral was sampled at greater than monthly resolution. A particular sample that was pristine would show quite a different Sr/Ca result from a sample that had calcite recrystallization. The record would have a sample-to-sample variance almost as large as the seasonal cycle even if the patches were only 1% calcite. The Sr/Ca data do not contain large deviations between near-by samples and recrystallization is not judged to be a problem in BG831.

Secondary aragonite is an inorganically precipitated marine cement. BG831 has been exposed to marine conditions since its origin so we must entertain the possibility of secondary aragonite contamination in BG831. The distribution coefficient for Sr in inorganic aragonite is higher \( K_{\text{Sr}} \approx 1.14 \) at 25°C; Kinsman and Holland, 1969; \( K_{\text{Sr}} = 1.21\pm0.03 \) (Enmar et al., 2000) than the distribution coefficient for Sr in coral aragonite (e.g., for Porites near Japan \( K^A_{\text{Sr}} \approx 1.056\pm0.003 \); Livingston and Thompson, 1971; Shen et al., 1996; Marshall and McCulloch 2002). This implies that inorganic aragonite will have a higher Sr/Ca ratio than coral aragonite precipitated from the same solution.

Secondary aragonite has been shown to affect SST reconstructions from corals resulting in calculated temperatures that are lower than measured temperatures (Enmar et al., 2000; Müller et al., 2001). A quick calculation using values from Enmar et al. (2000) can confirm that minor amounts (\( \sim 4\% \)) of inorganically precipitated aragonite can have a significant affect on climate reconstruction. The bulk coral Sr/Ca ratio is equal to a weighted average of the coral Sr/Ca and the secondary aragonite Sr/Ca. Using 0.0229 (10.4751 mmol/mol) for secondary aragonite and 0.0203 (9.2858 mmol/mol) for primary aragonite, the Sr/Ca ratio of the bulk, contaminated coral would be 0.2040 (9.3333 mmol/mol). The effect translates to 0.8°C using the Sr/Ca-SST relationship of Beck et al. (1997) and 1.0°C using the Sr/Ca-SST relationship from this study. In reality, small percentages (1-4%) of secondary aragonite are not likely to be missed if the coral samples are examined for their presence because secondary aragonite is discernable using standard techniques. Although it is theoretically possible to underestimate SST by 0.4 - 0.5°C given the undiagnosed presence of 2% secondary aragonite, it is not likely that the calculated temperatures from fossil coral BG831 are too low since they are close to the upper bound of physically possible SSTs.
5.2. Climatic Implications of Fossil Coral Data and Model Results

5.2.1. Regional Implications

The seasonal cycles of $\delta^{18}O$ in the fossil and modern corals may be used to investigate the source of seasonal variability at Vanuatu (e.g., reduced atmospheric convection or a lateral shift in the center of convection). As reported in section 4.3, the seasonal range in $\delta^{18}O$ is smaller in the fossil record than in the modern. Reduced amplitude in this case cannot result from under-sampling of the slower-growing coral because the sampling resolution was increased proportionally for the slower-growing fossil coral. Furthermore, the continued success of studies showing primary Sr/Ca dependence on temperature (e.g., Houck et al., 1977; Smith et al, 1979; Beck et al. 1992, 1997; Hughen et al., 1999; Gagan et al., 2000; Al-Rousan et al., 2002; Marshall and McCulloch 2002; Quinn and Sampson, 2002) has diminished initial concerns over the growth-rate dependence of Sr/Ca in coral skeletons (Weber, 1973; de Villiers et al., 1995).

The smaller amplitude annual cycle of $\delta^{18}O$ in BG831 relative to the modern must be due to reduced seasonal salinity variations in ancient times because ice volume affects are not possible on seasonal time scales, and the difference cannot be attributed to temperature because the seasonal Sr/Ca changes (i.e., SST changes) are similar in amplitude between modern and fossil records. Seasonal-scale salinity changes are due to the migration of the SPCZ and the SSS front associated with it in the Vanuatu region today. The fossil coral data are consistent with an interpretation that the SPCZ was either weakened or displaced from its present location during the time that the fossil coral lived. The model results support this interpretation because a weakened or displaced convergence zone would cause higher mean salinity values.

Warmer than modern SST’s indicated by the coral proxy record support the hypothesis that the mean salinity offset results from lateral movement of the center of atmospheric convergence rather than weaker convection over the region. Warmer tropical SSTs are conducive to increasing convergence, but the local record indicates smaller seasonal hydrologic changes and a generally higher E-P ratio (i.e., reduced
convergence). Thus the coral did not likely experience the brunt of rainfall associated with the center of convergence, and the convergence zone was not likely to have been directly over Vanuatu during the years of coral growth.

Could the fossil coral represent a single decadal-scale shift in the SPCZ as has occurred this century due to shifts in the Inter-decadal Pacific Oscillation? Modern shifts in rainfall at Vanuatu associated with movements of the IPO are not clearly recorded as shifts in the mean value of $\delta^{18}$O in the full modern coral record (1928-1992). It is therefore unlikely that the fossil coral represents a similarly scaled event, but instead represents a larger spatial displacement of organized convection.

5.2.2. Global Implications

At first glance, warmer and saltier (drier) seawater conditions seem contradictory in the tropics where increases in SST are often accompanied by increases in evaporation and precipitation, but there are modern exceptions. In the WPWP today, the region around 179.5$^\circ$E, 3.5$^\circ$S, is 1.2$^\circ$C warmer and 0.5 saltier than Vanuatu (Levitus and Boyer, 1994). That area is in the center of the warm pool and lies between the ITCZ and SPCZ.

The paleoclimate implications of the fossil coral data from Vanuatu can be evaluated in light of the results of other paleoclimate studies in the region. Foraminiferal Mg/Ca ratios indicate about 3$^\circ$C cooling in the western tropical Pacific (tropical South China Sea — Stott et al., 2002; Ontong Java Plateau — Lea et al., 2000) during the last glacial maximum (Stage 2) and other glacial stages over the past 450 kyr. The age estimate of BG831 is centered close to the minimum sea level of MIS 10, so it is reasonable to hypothesize that the coral chemistry should indicate ~ 3$^\circ$C cooling as well. One glance at Fig. 18 shows that 3$^\circ$C cooling is not even in the range of possibilities for BG831 given the coral data and model input variables. To obtain 3$^\circ$C cooling, one must either melt more than the total modern volume of glacial ice into the sea, or have seawater Sr/Ca decreasing with increasing ice volume. Neither scenario seems possible. How can this be so?

One scenario to be considered is that the fossil coral has been altered by diagenesis, but the discussion in section 5.4 show this to be unlikely. If diagenesis cannot be used to “explain away” the climatic implications of the fossil coral data, then a
climate solution consistent with the fossil coral data needs to be determined. According to the foraminiferal Mg/Ca record of Lea et al. (2000), the Western Pacific (specifically the Ontong Java Plateau; OJP (159°22'E, 0°19’N) warmed early during MIS 10 so that by ~340 ka, surface temperatures were close to their present value. The age estimate of BG831 ranges between 337 and 371, overlapping the early warming event so one cannot discount that BG831 grew during that early warming to explain why seawater during a glacial period was similar or warmer than present day conditions.

Another simple explanation is that meridional thermal gradients in the Pacific are compressed and shifted southwards during glacial times because of sea ice in the northern Hemisphere. Vanuatu is on the edge of the WPWP, and a southward shift in the warm pool could maintain or actually increase local temperatures, even during modest cooling of the warm pool. Although the western equatorial Pacific seems to have cooled by a couple of degrees Celsius during the MIS 10 glacial maximum as is evidenced by planktonic foraminiferal Mg/Ca data (Lea et al., 2000), there is also evidence that it warmed back up to almost 28 °C around 350 ka. Vanuatu would have experienced similar to warmer than present SST if both Vanuatu and the OJP were bathed by the warmest equatorial waters at the time (i.e. both in the warm pool). A southward shift in the northern edge of the WPWP is documented by other data (Trent-Staid and Prell, 2002), but a southward shift in the southern border of the WPWP has not yet been documented by other studies.

Another scenario that has been suggested is that salinity in the western Pacific decreases during glacial periods. Lea et al. (2000) suggested this when they found a smaller range of seawater δ¹⁸O in the western equatorial Pacific (WEP) than the eastern equatorial Pacific (EEP) over glacial/interglacial cycles. Since seawater δ¹⁸O in the EEP changed by the same amount as the global average, salinity changes in the WEP were invoked. Model results (Fig. 18) illustrate that salinity was not likely to be less than today in the waters where BG831 grew. To have even a slight salinity decrease, one must assume negligible changes to seawater Sr/Ca over glacial/interglacial cycles. While this is possible, it is not probable, since at least some increase in seawater Sr/Ca is expected during a glacial period (Stoll and Schrag 1998).
Additionally, Lea et al. (2000) used the average glacial/interglacial values to draw their conclusions, but their data for MIS 10 appears anomalous. During glacial stage 10, seawater $\delta^{18}O$ in the EEP is considerably greater than in the WEP, whereas during glacial stages 2, 4, 6 and 8, seawater $\delta^{18}O$ values are almost equal in the WEP and EEP. Thus, the zonal isotopic gradient in the tropical Pacific during glacial stage 10 completely reverses from the modern gradient. Recent work has demonstrated that the isotopic composition of seawater during glacial times is heterogeneous (Adkins et al., 2002; Schrag et al., 2002) and clearly more work is required to further constrain seawater $\delta^{18}O$ during glacial stage 10.

New planktonic foraminiferal $\delta^{18}O$ and Mg/Ca ratio data from the WPWP indicates a moderate salinity increase during the LGM along the eastern edge of the Indonesian archipelago (Stott et al., 2002). These new data are consistent with model results and data from fossil coral BG831, which indicate an increase in salinity during glacial times. At first, the data from both of these studies appear to directly conflict with Lea et al. (2000) who indicated decreasing salinity in the WEP during glacial periods, but Stott et al. (2002) point out a simple solution. The Ontong Java Plateau is situated to receive large increases in rainfall during warm phase events as the locus of atmospheric convection shifts eastward from Indonesia. If atmospheric convection shifted locations in a similar manner during cold periods in the northern Hemisphere, then one would expect both Vanuatu and Indonesia to be more saline, and other regions (such as the OJP) to become less saline.

Two related coupled atmosphere ocean general circulation models (AOGCM’s) for the Last Glacial Maximum (LGM) support the theory of a northeastward shift in the center of convergence. Kitoh et al. (2001) initially used their AOGCM to investigate the influence of freshwater on thermohaline circulation in the north Atlantic, but they also found positive precipitation anomalies in the central Pacific and negative precipitation anomalies in the WEP. Further investigation revealed an eastward shift in Walker circulation associated with a weaker Walker cell and a westerly trade wind anomaly during the LGM (Kitoh and Murakami, 2002).

Fully coupled models have only recently been used to simulate LGM climate phenomena (Bush and Philander, 1998; 1999; Hewitt et al., 2001; Kitoh et al., 2001;
Kitoh and Murakami, 2002) and not all of them agree. Recent modeling studies (Bush and Philander, 1998; 1999; Hewitt et al., 2001) illustrate an alternative hypothesis, which is consistent with the data at Vanuatu and Indonesia (Stott et al., 2002). Bush and Philander (1998, 1999) concluded that the center of atmospheric convection moves west into the Indian Ocean and the tropical Pacific is characterized by a deeper thermocline, increased trade wind activity, increased salinity, and decreased convection between about 120˚E and 180˚E during glacial periods. The model predicts a simultaneous decrease in trade wind strength and decrease in atmospheric pressure in the Indian Ocean between 60˚E and 120˚E (Bush and Philander, 1998; 1999). These physical phenomena are consistent with a variety of paleo-data including: eolian deposits, planktonic foraminifera, pore waters, pollen records, and ice cores (Bush and Philander, 1998; 1999; and references therein). However they are inconsistent with a freshening of the WPWP (Lea et al., 2000).

SST data from foraminiferal assemblage work similar to the CLIMAP project (CLIMAP project members 1976, 1981) also support a center of convergence in the Indian Ocean rather than over the Indo-Pacific region (Trend-Staid and Prell, 2002). Faunal evidence suggest decreased average temperatures around Indonesia, a smaller WPWP, warmer temperatures in the equatorial Indian Ocean, and an increased zonal thermal gradient across the Pacific Ocean (Trend-Staid and Prell, 2002). All of these features are consistent with increased trade winds in the WEP and increased convection in the equatorial Indian Ocean. Admittedly, the coupled model results in SST’s that are cooler than the foraminiferal data indicate, but the general patterns are similar.

In summary, the data from this study are consistent with two different models of glacial periods that both predict a displaced center of convergence during glacial periods. Additional “paleo” data are needed to constrain the east-west thermal and salinity gradients across both the Pacific and Indian Oceans during glacial periods, and to help resolve the discrepancies in current model predictions.
6. CONCLUSIONS

Analysis of two coral geochemical records, one modern, one fossil has shed light on the following questions posed at the beginning of this work: 1) did ENSO exist during other glacial periods, and 2) what was the background state of the surface ocean in terms of temperature and salinity, during those periods?

To answer these questions, this study confirmed that modern *Porites* spp. corals from Vanuatu record significant surface ocean variability in the $\delta^{18}$O and Sr/Ca of their skeletons. Sr/Ca ratios in skeletal aragonite strongly correlate with SST from blended-source, 1°x1° gridded data sets. Vanuatu coral oxygen isotope ratios are nearly equally affected by SST and SSS in the calibration interval of 1980-1990. Hydrologic and thermal details of specific ENSO events are in fact resolvable in the modern coral geochemical record.

Fossil corals can preserve climatologically significant information in their skeletal aragonite over long (10^5 yrs) time scales. Changes in seawater chemistry can have large affects on the climatological interpretation of fossil coral records and must be accounted for even if the geochemical signature of environmental variability is preserved in the fossil coral skeletons. A simple mathematical model using empirically derived relationships joining $\delta^{18}$O and Sr/Ca with seawater Sr/Ca variations, seawater $\delta^{18}$O variations, SST, and SSS, facilitates the interpretation of coral geochemical records.

The following conclusions have been reached regarding interannual variations and mean climate state in the western tropical Pacific around 350ka before present:

1) Coral proxy data and mathematical modeling of Pleistocene mean SST and SSS results in temperature estimates up to ~2°C warmer and salinity up to ~2 psu saltier than present-day conditions, if seawater Sr/Ca were 1-2% higher in MIS10 than they are today. Our fossil coral data and modeling results preclude colder SST and lower SSS at Vanuatu during MIS10. Accurate estimates of past values
of seawater Sr/Ca remain the largest obstacle to the accurate reconstruction of tropical SST in the past using pristine fossil corals.

2) During this same time window, seasonal SST ranges were very similar to modern seasonal SST ranges, while seasonal hydrologic variations were reduced in amplitude compared to modern. The reduced seasonal SSS variations and increased SSTs near Vanuatu are interpreted as evidence that the SPCZ was displaced from its present location during at least part of MIS 10.

3) ENSO or ENSO-like interannual variations are not unique to interglacial time periods, nor just the past 130 kyrs, but characterize the tropical Pacific at least back to MIS 10, between 347-371 ka.
REFERENCES


Figure 1. Map of the southwest Pacific basin showing location of Espiritu Santo, Republic of Vanuatu. Inset map shows detail of the locations where modern (MCB, Malo Channel) and fossil (BG831; Bougainville Guyot, ODP Site 831) corals were recovered.
Figure 2. Maps of mean sea surface salinity (A) and mean sea surface temperature (B) in the southwest Pacific basin showing the approximate location of the South Pacific Convergence Zone (SPCZ), Intertropical Convergence Zone (ITCZ) and Western Pacific Warm Pool (WPWP). The black circles lie over the study site, and the data are from evitus and Boyer (1994).
Figure 3. Monthly climatology of SST (black line, X symbols) and SSS (green line, plus symbols) at Vanuatu for the period 1980-1992. Salinity data are from Gouriou and Delcroix (in press). SST data have been extracted from the appropriate 1° x 1° grid box of GISST2.36 database (Parker et al., 1995). Note that minima (maxima) in SST coincide with maxima (minima) in SSS.
Figure 4. Climatological average monthly rainfall (purple line; closed circles) for Luganville, Vanuatu (-15.52, -167.22). Data are from the NOAA Climate Prediction Center (http://www.cpc.ncep.noaa.gov/pacdir/NDATA.html). SSS data (green line; crosses) are from Fig. 2. Rainfall at Vanuatu is inversely proportional to SSS.
Figure 5. Monthly Niño 3.4 anomaly (NOAA Climate Prediction Center, http://www.cpc.ncep.noaa.gov/data/indices/), monthly SSTA (black line, X symbols; GISST 2.36; Parker et al., 1995), and SSS (green line, plus symbols; Gouriou and Delcroix, in press) at Espiritu Santo Island, Vanuatu. Shaded vertical bars denote ENSO warm phase years as defined by Trenberth (1997). Temperatures are cooler, and/or salinity increases in Vanuatu during ENSO warm events.
Figure 6. Five month running average of monthly SSSA and SSTA at Vanuatu. Data source is as listed in Fig. 4. Anomalies are calculated as deviations from the average annual cycle between 1980-1992 (Fig. 2). Shaded regions are ENSO warm phase years (Trenberth, 1997), and the labels on the lower graphs indicate the local response to ENSO. Black horizontal lines are 1.5 standard deviations of the anomaly data and represent the threshold over which anomalies at Vanuatu are functionally defined as ENSO events.
Figure 7: X-radiographs of Malo Channel B and BG831 coral slabs used in this study. Annual banding is clearly visible throughout the cores. The light colored strip along each core is the area routed out in sampling.
Figure 8. Stratigraphic forward model constrains the age of sample BG831 based on zero age depth of sample BG831, subsidence rate, and sealevel history. Sealevel for the past 450 ka from SPECMAP stack of global oxygen isotope variations (Imbrie et al., 1984) and from atmospheric oxygen isotope variations in ice core records (Shackleton et al., 2000) are plotted on the same axes with subsidence rates of Bougainville Guyot (Taylor et al., 1994). The intercepts between the subsidence rate lines (dashed green and red) and the SPECMAP sealevel curve (solid blue) place the age of coral sample BG831 between 337 ka and 371 ka (denoted by the grey vertical bar). Using the revised sealevel curve from Shackleton (2000) extends the possible age range to a somewhat older limit. It is unlikely that the coral grew in that extended time range because the oldest age for the coral would overlap the U-series-date for a coral from ~75m lower in the core.
Figure 9. Scanning electron microscope images of modern coral MCB (A, B) and fossil coral BG831 (C, D) illustrate the unaltered skeletal and crystalline structure of BG831. Scale bars are 50µm in A and C, and 200µm in B and D.
Figure 10: Photomicrographs under cross-polarized light (A) and transmitted light (B) of BG831 illustrating an example of good preservation in a fossil coral. Aragonite crystal growth centers are visible (arrows) within the skeletal matrix and the edges of the pores show no secondary crystal growth. Scale bars are approximately 100µm.
Figure 11. Comparison between depth series and re-sampled, monthly resolved time series for Sr/Ca (blue), δ¹⁸O (red) and δ¹³C (black). Data from modern coral MCB are plotted in panels A and B; whereas data from fossil coral BG831 are plotted in panels C and D. Converting depth series to a time series does not fundamentally alter the information in the data, though re-sampling to monthly intervals does smooth the data.
Continued on next page
Figure 12. (A) Monthly $\delta^{18}$O data for fossil coral BG381 (open red circles and red line) and modern coral MCB (red filled circles and red line). (B) Monthly Sr/Ca ratios for fossil coral BG831 (open blue squares) and modern coral MCB (filled blue squares). Mean (solid black horizontal lines) and two standard deviations (coarsely dotted line for BG831 and finely dotted line for MCB) of the data, including monthly variations and analytical error, are noted. The fossil and modern coral have significantly different mean $\delta^{18}$O values, but equivalent mean Sr/Ca values.
Figure 13. Climatological mean annual cycle of coral δ\(^{18}\)O (A) and Sr/Ca (B) in fossil coral BG831 (open symbols) and modern coral MCB (closed symbols). Error bars are 95% confidence of the monthly means. Black vertical bar in upper left indicates two standard deviations of the analytical error. The seasonal ranges of Sr/Ca in the MCB and BG831 are not significantly different from each other, implying that the seasonal temperature ranges in Vanuatu were similar to today. Conversely, the seasonal δ\(^{18}\)O range in the fossil coral is much smaller than the modern seasonal range, implying weaker seasonal hydrologic variations during the time of fossil coral growth.
Figure 14. Monthly coral Sr/Ca record from modern coral MCB (blue solid line; solid squares) plotted versus monthly SST record at Vanuatu (GISST2.36; Parker et al., 1995). The increase in coral Sr/Ca ratios in the 1990’s is consistent with Sr enrichment at the top of the core from coral tissue contamination of the samples (Alison, 1996). The two records are well-matched as evidence by their high correlation ($r=-0.83; p<0.0005; 1980-1990$).
Figure 15. Reduced major axis (RMA) regression results for the comparison of coral Sr/Ca (MCB) and SST for the period 1980-1990. Solid line is the slope of the Sr/Ca-SST relationship using the SST data from GISST2.36 (Parker et al., 1995) and the dashed is the slope of the Sr/Ca-SST relationship using the SST data from NCEP (Reynolds and Smith, 1994).
Figure 16. Instrumental SST and SSSA records compared with coral proxy records of climate variability at Vanuatu. (A) Monthly SST (black line, X symbol; Parker et al., 1995) and SSSA (green line, plus symbol; Gouriou and Delcroix, in press). (B) Modern coral monthly Sr/Ca and δ¹⁸O anomaly. Grey bars in A and B represent ENSO warm phase events as defined by Trenberth (1997). (C) Pleistocene coral Sr/Ca and δ¹⁸O anomalies from sample BG831. Grey bars in C denote Pleistocene ENSO warm phase events, applying the criteria defined for the modern relationship between coral proxy records and instrumental records.
Figure 17. Sr/Ca and $\delta^{18}$O anomalies from fossil coral BG831 illustrating fossil ENSO warm phase events (gray bars). The horizontal lines represent 1.5 standard deviations of the anomaly data (i.e., the criteria defined through the analysis of the modern relationship between coral proxy records and instrumental records).
Figure 18. Four model solution sets of Pleistocene SST changes relative to modern (ΔT, °C) and SSS changes relative to modern (ΔS) representing seawater Sr/Ca changes relative to modern of 0%, 1%, 2%, and 3% as labeled. Plots are limited to those SST and SSS values that correspond with ice volume changes no greater than 120 m. The blue dots represent the maximum and minimum age/ice-volume boundaries from the forward stratigraphic modeling (see Fig. 8). The red open crosses denote the central date and ice volume obtained by forward stratigraphic modeling (see Fig. 8). A thin black horizontal line denotes a mean annual SST of 30°C, a value assumed to be the upper bound of possible mean annual SST in this region. The gray shaded area illustrates the “most-likely” solutions of SST and SSS given model inputs for changes in ice volume/sea level and physical SST limits (see text for complete discussion).
Table 1. List of values and sources used to solve equations 6 and 7 for temperature, ice volume and salinity.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Values Used</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\frac{\partial O}{\partial T}$</td>
<td>-0.18 ‰/ °C</td>
<td>Gagan et al. (1998)</td>
</tr>
<tr>
<td>$\frac{\partial O}{\partial I}$</td>
<td>0.00833%‰/m</td>
<td>Schrag et al. (2002)</td>
</tr>
<tr>
<td>$\frac{\partial O}{\partial S}$</td>
<td>0.273%‰/psu</td>
<td>Fairbanks et al. (1997)</td>
</tr>
<tr>
<td>$\frac{\partial Sr}{\partial T}$</td>
<td>-0.0462 mmol/mol °C$^{-1}$</td>
<td>This study</td>
</tr>
<tr>
<td>$\frac{\partial Sr}{\partial I}$</td>
<td>0-3% per 120m Ice</td>
<td>Stoll and Schrag (1998)</td>
</tr>
</tbody>
</table>

1 $\partial O/\partial T$; change in coral $\delta^{18}O$ (‰) per 1°C change in SST.
2 $\partial O/\partial I$; change in coral $\delta^{18}O$ (‰) per 1m change in ice volume.
3 $\partial O/\partial S$; change in coral $\delta^{18}O$ (‰) per 1psu change in salinity.
4 $\partial Sr/\partial T$; change in coral Sr/Ca (mmol/mol) per 1°C change in SST.
5 $\partial Sr/\partial I$; change in coral Sr/Ca (mmol/mol) per 1m change in ice volume.
Table 2. Summary of errors (1σ) associated with model estimates (Fig. 17; equations 6 and 7) of changes in Pleistocene SST (ΔT, °C) and SSS (ΔS) relative to modern conditions at Vanuatu. Other variables in the calculations are changes in seawater Sr/Ca ratios and ice volume/sealevel (ΔI, meters below present sealevel). Each row in the table represents a single vertex of the shaded polygon in Figure 17, which highlights ΔT and ΔS values judged to be most likely (see text for discussion).

<table>
<thead>
<tr>
<th>%Sr/Ca change mmol/mol°C⁻¹</th>
<th>ΔI meter</th>
<th>ΔT °C</th>
<th>Error in ΔT(1σ) °C</th>
<th>ΔS psu</th>
<th>Error in ΔS (1σ) psu</th>
</tr>
</thead>
<tbody>
<tr>
<td>1%</td>
<td>98</td>
<td>1.9</td>
<td>0.4</td>
<td>1.0</td>
<td>0.4</td>
</tr>
<tr>
<td>1%</td>
<td>50</td>
<td>1.0</td>
<td>0.4</td>
<td>1.8</td>
<td>0.5</td>
</tr>
<tr>
<td>1.2%</td>
<td>98</td>
<td>2.3</td>
<td>0.4</td>
<td>1.2</td>
<td>0.5</td>
</tr>
<tr>
<td>2%</td>
<td>58.9</td>
<td>2.3</td>
<td>0.4</td>
<td>2.4</td>
<td>0.6</td>
</tr>
<tr>
<td>2%</td>
<td>50</td>
<td>2.0</td>
<td>0.4</td>
<td>2.5</td>
<td>0.6</td>
</tr>
</tbody>
</table>