Geochemistry Geophysics Geosystems G3

Article
Volume 11, Number 6
8 June 2010
Q06005, doi:10.1029/2009GC002758
ISSN: 1525-2027

Published by AGU and the Geochemical Society

Click Here for Full Article

A snapshot of climate variability at Tahiti at 9.5 ka using a fossil coral from IODP Expedition 310

Kristine L. DeLong
Department of Geography and Anthropology, Louisiana State University, 227 Howe-Russell Geoscience Complex, Baton Rouge, Louisiana 70803, USA (kdelong@lsu.edu)

Terrence M. Quinn
Institute for Geophysics, Jackson School of Geosciences, University of Texas at Austin, 4412 Spicewood Springs Road, Austin, Texas 78759, USA

Also at Department of Geological Sciences, Jackson School of Geosciences, University of Texas at Austin, 1 University Station C1100, Austin, Texas 78712, USA

Chuan-Chou Shen and Ke Lin
Department of Geosciences, National Taiwan University, No. 1, Sec. 4, Roosevelt Road, Taipei 10617, Taiwan

[1] The Integrated Ocean Drilling Program (IODP) Expedition 310 recovered drill cores from the drowned reefs around the island of Tahiti (17°40′S, 149°30′W), many of which contained samples of massive corals from the genus *Porites*. Herein we report on one well-preserved fossil coral sample: a 13.6 cm long *Porites* sp. dated by uranium series techniques at 9523 ± 33 years. Monthly $\delta^{18}O$ and Sr/Ca determinations reveal nine clear and robust annual cycles. Coral $\delta^{18}O$ and Sr/Ca determinations estimate a mean temperature of ~24.3°C (~3.2°C colder than modern) for Tahiti at 9.5 ka; however, this estimate is viewed with caution since potential sources of cold bias in coral geochemistry remain to be resolved. The interannual variability in coral $\delta^{18}O$ is similar between the 9.5 ka coral record and a modern record from nearby Moorea. The seasonal cycle in coral Sr/Ca is approximately the same or greater in the 9.5 ka coral record than in modern coral records from Tahiti. Paired analysis of coral $\delta^{18}O$ and Sr/Ca indicates cold/wet (warm/dry) interannual anomalies, opposite from those observed in the modern instrumental record.

Components: 8100 words, 5 figures, 1 table.

Keywords: Sr/Ca; oxygen isotopes; early Holocene; Integrated Ocean Drilling Program; Expedition 310; Tahiti sea level.

Index Terms: 4916 Paleoceanography: Corals (4220); 4922 Paleoceanography: El Nino (4522); 4954 Paleoceanography: Sea surface temperature.

Received 4 August 2009; **Revised** 2 April 2010; **Accepted** 19 April 2010; **Published** 8 June 2010.

DeLong, K. L., T. M. Quinn, C.-C. Shen, and K. Lin (2010), A snapshot of climate variability at Tahiti at 9.5 ka using a fossil coral from IODP Expedition 310, *Geochem. Geophys. Geosyst.*, *11*, Q06005, doi:10.1029/2009GC002758.

1. Introduction

[2] Subannually resolved geochemical time series developed from the aragonite skeletons of modern (live) corals from the tropical Pacific Ocean have been reported on extensively to investigate and reconstruct climate variability, especially on interannual (El Niño–Southern Oscillation; ENSO) to multidecadal time scales (e.g., Pacific Decadal Oscillation) [e.g., *Linsley et al.*, 2000; *Cobb et al.*, 2003]. Climate studies using fossil (dead but not remineralized) corals have the potential to provide insight into past behavior of the tropical ocean-climate system [e.g., *Gagan et al.*, 1998; *Felis et al.*, 2004; *Gagan et al.*, 2004; *Kilbourne et al.*, 2004; *Shen et al.*, 2005]. The most commonly developed records of proxy climate variability in coral skeletons are oxygen isotopes (δ^{18}O) and strontium-to-calcium ratios (Sr/Ca). Variations in the coral δ^{18}O record reflect the combined effects of thermal and hydrologic (δ^{18}O of seawater; δ^{18}Osw) changes in the surface ocean. These variations are a particularly good monitor of ENSO dynamics, because the response of the tropical oceans to ENSO forcing tends to be both thermal and hydrologic in the ENSO-sensitive regions of the oceans. Coral Sr/Ca variations record changes in seawater temperature and seawater Sr/Ca at the time of skeletal precipitation. Uncertainties of this thermal proxy include the possible variability of glacial-interglacial oceanic Sr budget [e.g., *Stoll and Schrag*, 1998], heterogeneity in Sr distribution of aragonite in the microdomain [e.g., *Allison et al.*, 2001], complicated physiological processes during the biomineralization [e.g., *Cohen et al.*, 2001], and growth-rated kinetic effects [e.g., *de Villiers et al.*, 1995].

[3] The challenge for coral paleoclimatologists working on fossil corals is to identify coral skeletons that have not undergone postdepositional alteration. Evidence of alteration in corals takes many forms, ranging from changes in mineralogy detectable by X-ray diffraction techniques, to changes in original chemistry detectable by radiogenic and stable isotope techniques, to changes in skeletal architecture and the addition of secondary aragonite detectable by petrographic techniques. A recent paper by *Hendy et al.* [2007] reviews the forms of diagenesis in corals.

[4] In this study, we describe geochemical variations from a fossil coral recovered by Integrated Ocean Drilling Program (IODP) Expedition 310. This coral sample exhibits sections with unaltered primary mineralogy, providing a window into early Holocene surface-ocean conditions for the central South Pacific. We describe the coral Sr/Ca variations, a proxy for sea surface temperature (SST), and coral δ^{18}O, a proxy for SST and hydrological variability (i.e., evaporation and precipitation), and compare these variations to modern conditions.

2. Background and Methods

2.1. Setting

[5] Tahiti (17°40′S, 149°30′W) is located in the central Pacific and is part of the Society Archipelago (French Polynesia; Figure 1). Tahiti may be best known in the climatological sense as the eastern end-member of the Southern Oscillation Index (SOI). A summary of in situ monthly variations in SST and sea surface salinity (SSS) at Papeete, Tahiti, measured from 1979 to 1990 reports that (1) SST reaches a maximum (minimum) in February, March, and April (July, August, and September) and (2) SSS reaches a maximum (minimum) in July, August, and September (January, February, and March) [*Boiseau et al.*, 1998]. The SSS maximum (minimum) lags precipitation wherein the dry (rainy) season at Tahiti extends from May to October (November to April) [*ORSTOM*, 1993]. The seasonal cycle in SST at Tahiti is ~2.5°C and the seasonal cycle in SSS is ~0.4 practical salinity units (psu). The interannual SSS variations are twice as large as seasonal variations and are coherent with changes in the SOI; furthermore, the interannual SST signal at Tahiti is smaller than the seasonal cycle and less coherent with SOI [*Gouriou and Delcroix*, 2002]. Interannual anomalies in SST and SSS warm and wet summers (cold and dry winters) and in phase with seasonal variations [*Gouriou and Delcroix*, 2002]. The local observations agree with those reported by *Conkright et al.* [2002], which document a mean SST of 27.5°C and a mean SSS of 36 psu.

2.2. Coral Samples

[6] IODP Expedition 310 drilled 37 holes into the reefs around Tahiti [*Camoin et al.*, 2007]. Core recovery was exceptional in this challenging shallow water drilling environment and included ~30 m of massive coral colonies, mostly of the genus *Porites*. One of the sections of *Porites,* 310-M0007B-11R-2, was recovered from site M0007 at 57 m below sea level offshore from Maara, Tahiti (Figure 1b). A 0.5 cm thick slab was extracted from the middle of the 13.6 cm long core cylinder and X-rayed to image the annual density banding (Figure 2b). Measure-

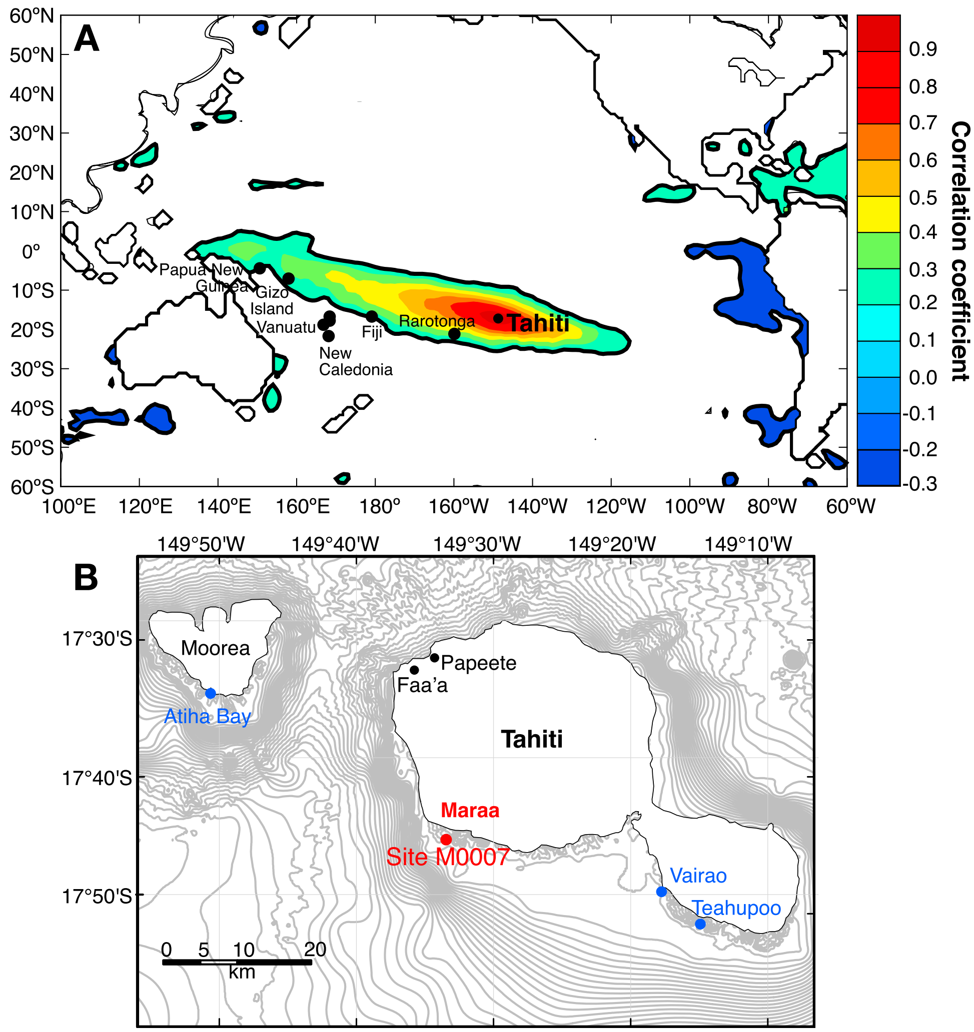

Figure 1. (a) Correlation map of monthly SST anomalies for Tahiti (17.5°S, 149.5°W) with SST anomalies from OISST [*Reynolds et al.*, 2002] for 1981–2008. Correlations >0.2 or <−0.2 are significant at the 95% confidence level (heavy black line); degrees of freedom adjusted for autocorrelation. (b) Location of IODP Expedition 310 site M0007 (17°45.9462′S, 149°33.0682′W) [*Camoin et al.*, 2007] and locations of other sites discussed.

ments of the density bands in the X-radiograph revealed a mean annual extension rate of 1.1 cm (Figure 2c). The coral slab was microsampled parallel to the extending corallites at 0.071 cm per sample for approximately monthly samples with a 0.14 cm diameter drill bit using a continuous routing program on a computer-aided triaxial micromill described in detail by *Quinn et al.* [1996]. The samples were split for isotopic and trace element analyses.

2.3. Analytical Determinations

[7] δ^{18}O and δ^{13}C analyses were performed by dissolution of powdered samples in phosphoric acid at 70°C in a Kiel III automated carbonate preparation device connected to a stable isotope ratio mass spectrometer, Thermo Finnigan Delta Plus XL. Isotopic values are reported in delta notation (δ) relative to the Vienna Pee Dee Belemnite (VPDB) isotopic standard. Average external precision is 0.05‰ for δ^{18}O and 0.04‰ for δ^{13}C (1σ, n = 24, NBS-19). External precision, based on years of determinations of NBS-19, is 0.08‰ for δ^{18}O and 0.05‰ for δ^{13}C (1σ).

[8] Sr/Ca determinations were made using a PerkinElmer 4300 Dual View Inductively Coupled Plasma Optical Emission Spectrometer (ICP-OES). The Sr/Ca of a gravimetrically prepared standard solution (IGS) was measured before and after each dissolved coral sample to correct for instrumental drift and noise [*Schrag*, 1999]. The average corrected precision of the IGS standard is 0.013 mmol/

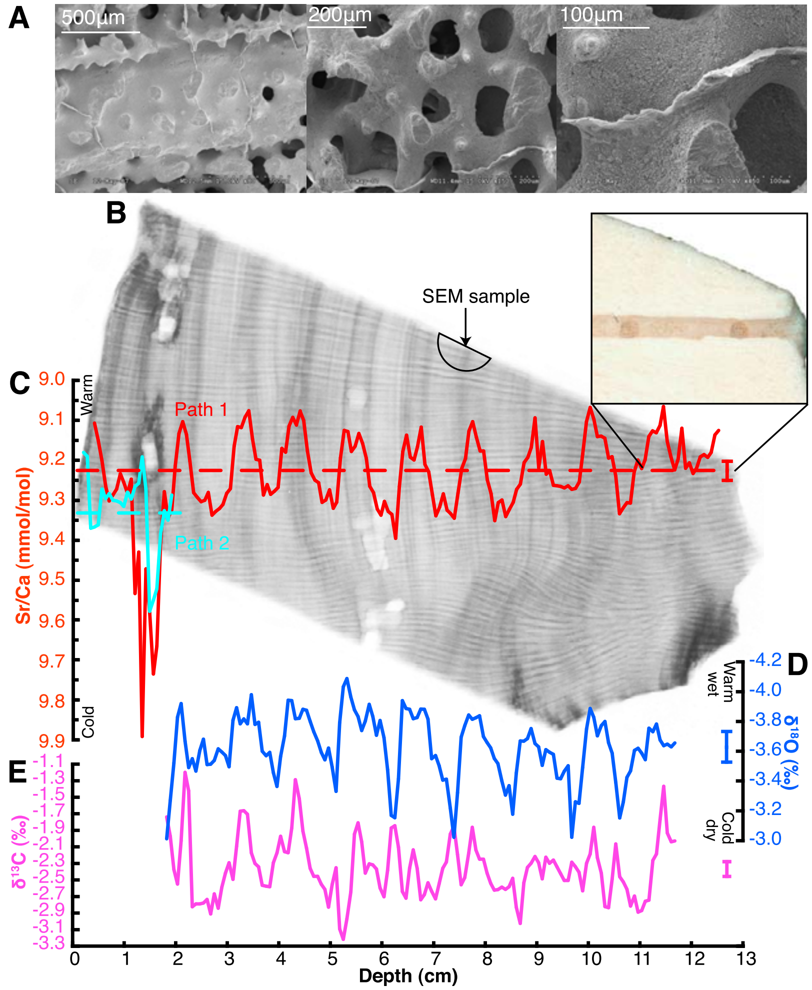

Figure 2. Fossil coral 310-M0007B-11R-2 shown with (a) SEM images and (b) X-ray positive with (c) monthly coral Sr/Ca determinations superimposed along the sampling path (red dashed line). A second path (cyan dashed line) was sampled parallel to the first path avoiding the bioerosion. The corallites are not optimally aligned to the sampling path from 11.5 cm to end (see inset of coral). Coral (d) δ^{18}O and (e) δ^{13}C variations for the same sampling path as Sr/Ca. Error bars (analytical precision, 2σ) are centered on their respective means.

mol for Sr/Ca (1σ, n = 45). The average corrected precision of a second standard (homogenized powder from a *Porites lutea* coral) is 0.012 mmol/mol for Sr/Ca (1σ, n = 68), consistent with the long-term ICP-OES analytical precision.

[9] The U-Th chemistry was performed in the High-Precision Spectrometry and Environment Change Laboratory (HISPEC) of the Department of Geosciences, National Taiwan University [*Shen et al.*, 2003, 2008]. The fossil coral ^{230}Th age was determined on a multicollector inductively coupled plasma mass spectrometer (MC-ICP-MS), Thermo Neptune [*Frohlich et al.*, 2009].

[10] The coral was examined for indications of diagenesis using a Hitachi S-3500N variable pressure Scanning Electron Microscope (SEM). A sample for the SEM was removed from the midsection of the coral next to the sampling path (Figure 2b).

3. Results

[11] The determined corrected ^{230}Th date is 9.523 ± 0.033 ka for 310-M0007B-11R-2 (Appendix A). The uranium concentration of 2.8 ppm is consistent with those for modern *Porites* and the corrected initial δ^{234}U of 144.3 ± 1.7 matches seawater values of 144–148 [e.g., *Shen et al.*, 2008]. The evidence strongly supports this fossil *Porites* is pristine, expect for the top 2 cm (see below). Analysis of the SEM images (Figure 2a) reveals that the primary aragonite is intact with only one occurrence of secondary precipitation that does not appear to be

aragonite. The sampling path includes a crust on the surface of the coral (0 cm) with low Sr/Ca values (2.5–8.9 mmol/mol; not shown). The first 2 cm are questionable as the sampling path penetrated a bioerosional feature below the slab surface that contains noncoral material with elevated Sr/Ca values (9.4–9.9 mmol/mol; Figure 2). Below 2 cm, δ^{18}O and Sr/Ca determinations exhibit nine clear annual cycles (Figure 2e). An additional 2 cm long path was sampled parallel to the original path avoiding the alteration. The results are improved, yet no clear annual cycles are present (Figure 2c). The last centimeter of the sampling path (11.5 cm to end) lacks the winter maximum in coral Sr/Ca that is present in the previous samples (Figure 2c). The absence of a winter Sr/Ca maximum may be the result of the corallites not being aligned parallel to the sampling path (Figure 2). Previous studies with *Porites lutea* in New Caledonia found similar results for this type of suboptimal sampling [*DeLong et al.*, 2007; *DeLong*, 2008]. The remaining analysis focuses on the geochemical determinations from 2 to 11.5 cm for which the means are -3.63 ± 0.21‰ (1σ) for δ^{18}O and 9.225 ± 0.078 (1σ) mmol/mol for Sr/Ca (Figure 2). The coral geochemistry was converted from the depth domain to the time domain by assigning maximum (minimum) coral Sr/Ca to minimum (maximum) SST climatology, then linearly interpolating to monthly intervals; the Sr/Ca chronology was applied to the isotopic records. Coral δ^{13}C is useful for the detection of alteration of the primary mineralogy; however, the interpretation of coral δ^{13}C as a climate proxy is ambiguous and will not be discussed further.

4. Discussion

4.1. Modern Coral Perspective From French Polynesia

[12] The reconstruction of climate variables from fossil coral records requires a modern calibration of geochemical proxies. Modern coral records are not available for our study site; therefore, we use *Porites* spp. records from nearby locations in Atiha Bay, Moorea [*Boiseau et al.*, 1998] and Teahupoo, Tahiti [*Cahyarini et al.*, 2009], 36 and 31 km from our site, respectively (Figure 1b). The Moorea record is a bimonthly resolved, 137 year long record of coral δ^{18}O that was interpreted as a SST record [*Boiseau et al.*, 1998]. Those authors found a lack of correlation between the Moorea coral δ^{18}O record and a precipitation record from Faa'a, Tahiti, 28 km from Moorea, thus concluded there was little influence from δ^{18}Osw on coral δ^{18}O. This result is unexpected as the Tahiti region experiences large interannual variability in SSS as the result of ENSO. For their δ^{18}O to SST calibration, *Boiseau et al.* [1998] used an 11 year SST record from Tahiti, which they extended to 1958 using an air temperature record from Faa'a and found coral δ^{18}O temperature dependencies varying from −0.12 to −0.23 ‰/°C. We reassessed the Moorea record using various gridded SST databases using standard calibration and verification methods [e.g., *Crowley et al.*, 1999]. The Optimum Interpolation Sea Surface Temperature Analysis (OISST) [*Reynolds et al.*, 2002], which includes satellite-derived SST for complete coverage, has the highest correlation ($r = -0.75$ for 1981–1990). The correlation decreases with the longer SST databases (e.g., $r = -0.59$ for 1900–1990 from Comprehensive Ocean-Atmosphere Data Set (COADS) [*Slutz et al.*, 1985]). The weighed least squares (WLS) [*York and Evensen*, 2004] regression slopes for the Moorea coral δ^{18}O with the OISST and COADS are −0.18 and −0.21 ‰/°C, respectively; these slopes are less than the slope for inorganic calcification (0.22 ‰/°C) and may reflect contributions from δ^{18}Osw.

$$\text{Coral } \delta^{18}\text{O}(\text{‰}) = -0.18(\pm 0.04;\ 1\sigma) \times \text{SST}(^\circ\text{C}) + 0.56(\pm 1.07;\ 1\sigma) \quad (1)$$

[13] The studies at Teahupoo [*Cahyarini*, 2006; *Cahyarini et al.*, 2009] report on monthly resolved coral Sr/Ca from two cores (TH1 vertical and TH1b horizontal; 72 and 21 years, respectively) extracted from the same colony and a third vertical core from Vairao (TH2; 92 years). Those authors report that the intracolony cores (TH1 and TH1b) exhibit significantly different means (0.06 mmol/mol) and the Vairao core has a higher mean than the other two cores (~0.15 mmol/mol). Furthermore, *Cahyarini* [2006] report that the intracolony cores have a low correlation outside the seasonal cycle, and the intercolony cores are not significantly correlated. These differences are troublesome for our fossil-based coral Sr/Ca reconstruction. Similar studies with *Porites lutea* in New Caledonia [*Stephans et al.*, 2004; *DeLong et al.*, 2007] did not yield differences of this magnitude. Examination of the X-radiographs for TH2 [*Cahyarini*, 2006] found anomalously high coral Sr/Ca values that coincide with sampling paths along the intersections of corallite bundles or "valleys" [see *Cahyarini et al.*, 2009, Figure 3]. Previous studies have demonstrated that sampling in valleys produces higher Sr/Ca values [*Alibert and McCulloch*, 1997; *Cohen*

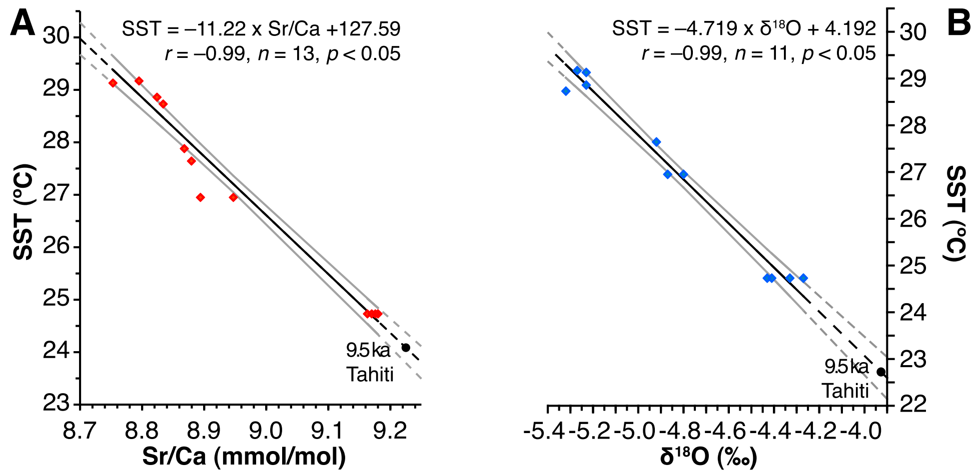

Figure 3. The southwest Pacific mean calibration for A.D. 1969–1992 between SST and coral (a) Sr/Ca and (b) δ^{18}O from *Porites* spp. studies including Amédée Island, New Caledonia (22°S, 166°E) [*Quinn et al.*, 1998; *Stephans et al.*, 2004; *DeLong et al.*, 2007]; Efate, Vanuatu (17°S, 168°E); Santo, Vanuatu (15.7°S, 167.2°E) [*Kilbourne et al.*, 2004]; Gizo Island (8°S, 157°E) [*Liu et al.*, 2005]; Rabaul, Papua New Guinea (4°S, 152°E) [*Quinn et al.*, 2006]; Papua New Guinea (4°S, 152°E); and Sabine Bank, Vanuatu (15.9°S, 166.14°E) [*Dunn et al.*, 2008] (see Figure 1 for locations). SST was extracted from HadISST [*Rayner et al.*, 2003] for the 1° grid box that contains each site. Regression performed using WLS [*York and Evensen*, 2004]; equations shown (black solid line) with error of the regression (gray lines, 2σ) and error of prediction (gray dashed lines, 2σ). The fossil coral δ^{18}O was corrected for ice volume (0.30‰; see section 4.3).

and Hart, 1997; *DeLong*, 2008]. Therefore, we do not use the TH2 record. *Cahyarini et al.* [2009] invoked "vital effects" such as low extension and calcification rates yielding high coral Sr/Ca values to explain their observed differences. Yet, these authors note that the horizontal core has slightly slower extension rates and lower coral Sr/Ca; therefore, growth-related vital effects do not explain the intracolony differences. These authors state that the top of the coral colony had reached the sea surface. The surface breach may have biased the geochemical record in TH1, reducing intracolony correlation. *Cahyarini et al.* [2009] report that the TH1b has a higher correlation with SST than the TH1. We selected the slope for the core TH1b (−0.05 mmol/mol/°C) [*Cahyarini et al.*, 2009] for the modern Tahiti calibration as that core has no reported bias and the sampling path on X-radiographs (in the work by *Cahyarini* [2006]) appears to be optimal.

4.2. Modern Coral Perspective From South Pacific

[14] We examined other coral studies in the South Pacific since the monthly coral Sr/Ca to SST calibration for Tahiti is questionable. The correlation map of SST anomalies for Tahiti (Figure 1) reveals significant correlation with Rarotonga and Fiji, the sites of two coral Sr/Ca reconstructions by *Linsley et al.* [2000, 2006]. We determined the coral Sr/Ca to SST calibration slope (−0.075 mmol/mol/°C for Rarotonga and −0.062 mmol/mol/°C for Fiji) similar to our reassessment of the Moorea record. The Rarotonga and Fiji slopes are greater than the slope determined by *Cahyarini et al.* [2009] for Tahiti; therefore, we evaluate our the variations in fossil coral record with a range of slopes. We are not able to use the Sr/Ca calibration equations from these studies to reconstruct absolute SST as different laboratory standards were used.

[15] To reconstruct absolute SST, we present a new southwest Pacific mean coral geochemistry to mean SST calibration. The regional calibration was determined using WLS regression [*York and Evensen*, 2004] for a collection of *Porites* spp. corals analyzed by T. M. Quinn and his students over the years, covering a wide latitudinal range (4°S–22°S) and range of mean temperatures (24–29°C; Figure 3). These studies used the same standards as this study.

$$\text{Coral Sr/Ca(mmol/mol)} = -0.089(\pm 0.038;\ 1\sigma) \times \text{SST}(^\circ\text{C}) + 11.371(\pm 1.023;\ 1\sigma) \quad (2)$$

$$\text{Coral } \delta^{18}\text{O}(‰) = -0.212(\pm 0.095; 1\sigma) \times \text{SST}(^\circ\text{C}) + 0.888(\pm 2.569;\ 1\sigma) \quad (3)$$

4.3. Fossil Coral Mean SST Reconstruction

[16] The reconstruction from the 9.5 ka coral yields a mean SST (24.1 ± 0.3°C, 2σ; equation (2) and Figure 3a), ~3.4°C cooler than modern. The Sr/Ca of seawater in the early Holocene was different from the modern seawater due to changes in the flux of Sr to the oceans in response to dissolution of aragonite-rich sediments on carbonate banks during subaerial exposure [*Stoll and Schrag*, 1998]. Those authors estimate a decrease of 0.5%–1.1% from modern Sr/Ca since the LGM. We do not have a modern Sr/Ca value for the correction, thus we determined the mean modern Sr/Ca (8.924 mmol/mol) from equation (2) and the mean SST for Tahiti. The Sr/Ca of seawater correction resulting in a correction of 0.11–0.23°C assuming sea level was ~27 m lower with respect to modern sea level [*Fleming et al.*, 1998]. The resulting SST estimates are 3.2–3.3°C cooler than modern. A coral Sr/Ca reconstruction from Espiritu Santo, Vanuatu (15.5°S, 167°E) found ~4°C colder than modern for corals dated 9688 years BP [*Beck et al.*, 1997]. SST cooling of this magnitude in the early Holocene is difficult to accept given that tropical Pacific SST reconstructions during the last glacial maximum (LGM) yield similar values of SST cooling [e.g., *Lea et al.*, 2000; *Visser et al.*, 2003; *Gagan et al.*, 2004]. A global climate model study driven by orbital forcing for 9 ka found slightly colder SST (~0.7°C) for the tropical Pacific [*Kutzbach and Gallimore*, 1988]. Furthermore, another study from IODP Expedition 310 found a 2.6–3.1°C cooling for Tahiti in the Younger Dryas [*Asami et al.*, 2009]. Several factors that could produce cold biases in coral Sr/Ca-based SST reconstructions are (1) coral sampling, (2) marine diagenesis, and (3) environmental setting.

[17] Recovering fossil corals in the original growth orientation during ocean drilling is difficult. It is possible that the vertical axis through the center of the fossil coral colony was not recovered. Studies with *Pavona clavus* in the Galapagos found significantly higher coral Sr/Ca and δ^{18}O values (colder by 1–1.6°C) for lower extension rates associated with horizontal sampling paths [*McConnaughey*, 1989; *de Villiers et al.*, 1994]. Studies with *Porites* spp. in the Red Sea and Japan found no difference between horizontal and vertical sampling paths (extension rate >0.7 cm/yr) [*Heiss et al.*, 1999; *Mitsuguchi et al.*, 2003]. A horizontal versus vertical test for a small *Porites* colony from Vanuatu did not find any difference in coral Sr/Ca values between paths (extension rate ~2 cm/yr) (K. L. DeLong, unpublished results, 2008). Conversely, the vertical core in the modern Tahiti study had colder values than the horizontal core [*Cahyarini et al.*, 2009]. The colder temperatures noted in the Galapagos studies are more likely related to slower growth rates in the horizontal samples. The growth rate for the early Holocene fossil coral is within the range observed for modern *Porites*; therefore, it is not obvious how growth-related influences are influencing the coral Sr/Ca. As discussed in section 4.1, microsampling *Porites* spp. along "valleys" can produce cold anomalies of 1–3°C [*Alibert and McCulloch*, 1997]; this study did not sample any valleys (Figure 2).

[18] Diagenesis includes dissolution of centers of calcification and secondary aragonite deposition within the coral skeleton, producing elevated Sr/Ca values (i.e., colder, up to 2°C for micron-scale sampling) [*Cohen and Hart*, 2004; *Allison et al.*, 2005]. The centers of calcification are a small percentage of the coral skeleton (~4% for modern corals); if dissolution had occurred, the impact on the overall coral Sr/Ca for millimeter-scale sampling would be small [*Allison et al.*, 2005]. *Hendy et al.* [2007] found a −1.7°C offset for early marine diagenesis in a coral with a 10% density increase that was visually apparent in X-radiographs with secondary aragonite and dissolution revealed by SEM. Similarly, *Quinn and Taylor* [2006] found dense regions in coral X-radiographs associated with diagenesis that coincide with shifts in δ^{18}O, δ^{13}C, Sr/Ca, and Mg/Ca (~10°C for Sr/Ca). The top 2 cm of the fossil coral clearly reveals indications of alteration, as evidenced by increase in Sr/Ca values from 9.4 to 9.9 mmol/mol (Figure 2) with a wide range of Mg/Ca values (4.0–185 mmol/mol; not shown). Similar to *Quinn and Taylor* [2006] and *Hendy et al.* [2007], this portion of the coral X-radiograph reveals a darker region associated with an increase in skeletal density. Below 2 cm, the coral radiograph does not reveal any dense regions along the sampling path. Examination of the SEM images found a single occurrence of secondary precipitation in a pore space; however, we did not find any indications of diagenesis by comparing our SEM images with those in previous studies [*Kilbourne et al.*, 2004; *Quinn and Taylor*, 2006; *Allison et al.*, 2007; *Hendy et al.*, 2007; *Asami et al.*, 2009]. It is possible that diagenesis may have altered the coral Sr/Ca signal; however, the alternation is not readily apparent. We conclude the magnitude of such alteration is less than that reported in studies that document alteration.

[19] Differences in environmental factors between modern and early Holocene settings might be

responsible for the colder SST estimated from the fossil coral. For example, *Porites* spp. occur in a wide range of water depths (0–30 m); therefore, the colder temperature reconstructed from the 9.5 ka coral may reflect a deeper water temperature versus SST [e.g., *Felis et al.*, 2004]. The warm (>26°C) surface water in the Tahiti region extends down to 100 m [*Rougerie*, 1994]; therefore, the possible temperature change for the depths that *Porites* spp. occur is limited to <1°C. Maara is located along the southwest coast of Tahiti where the seasonal trade winds from the southeast [*ORSTOM*, 1993] may produce localized upwelling. An upwelling signal of 1–3°C at 15 m was detected in current meters off Papeete [*Wolanski and Delesalle*, 1995]. A study of early to mid-Holocene corals off the southwest coast of New Caledonia found evidence for increased upwelling compared with modern corals [*Montaggioni et al.*, 2006]. In the early Holocene, sea level was ~27 m lower than today [*Fleming et al.*, 1998] and the reefs formed along the outer edge of the Tahitian platform. These early Holocene reefs may have been exposed to colder water from depth if strong southeast trade winds occurred. Climate modeling studies for 9 ka found increased cross equatorial flow into the northern hemisphere associated with the enhanced Asian Monsoon [*Kutzbach and Gallimore*, 1988; *Marzin and Braconnot*, 2009]. Forthcoming ecological reconstructions for Tahiti based on fossil coral samples may provide better constraints on the environment of the early Holocene reefs. In summary, there are three possible explanations or combinations of for the colder coral Sr/Ca temperature reconstruction for the early Holocene: water depth (<1°C), localized upwelling (1–3°C), and diagenesis (<1.7°C); these hypotheses can be tested but are beyond the scope of this study.

[20] The potential cold biases discussed for coral Sr/Ca may influence coral δ^{18}O to varying degrees, in which environmental setting influences both coral Sr/Ca and δ^{18}O. The variations in coral δ^{18}O respond to changes in ice volume, δ^{18}Osw, and SST. Assuming sea level was ~27 m lower with respect to modern sea level [*Fleming et al.*, 1998], the ice volume correction ranges from 0.22 to 0.30‰ assuming 0.083‰/m [*Schrag et al.*, 2002] and 0.011‰/m [*Fairbanks and Matthews*, 1978], respectively. The δ^{18}O corrected for ice volume yields temperatures (22.7–22.3°C) for mean calibration (equation (3)) whereas the equation for modern Tahiti (equation (1)) yields temperatures (24.7–24.3°C) ~ the same as the Sr/Ca estimate for 9.5 ka. The differences between the mean calibration for the southwest Pacific (equation (3)) and the modern Moorea (equation (1)) may be related to different hydrological regimes on opposing sides of the South Pacific Convergence Zone. The Moorea study used a cleaning method described by *Boiseau and Juillet-Leclerc* [1997] that shifted the mean isotopic values by −0.17‰ (~0.85°C) relative to untreated corals. The difference in means between the modern coral δ^{18}O (−4.42‰ for 1979–1990; SST = 27.4°C), corrected for cleaning, and the 9.5 ka coral, corrected for ice volume, is 0.4–0.32‰ (~1.6–2.2°C) for sea level corrections of 0.22 to 0.3‰, respectively.

4.4. Fossil Coral Interannual to Seasonal Variability

[21] We examine the interannual variability by comparing our 9.5 ka coral record with the modern Moorea coral δ^{18}O record [*Boiseau et al.*, 1998] in which we use in 9 year windows (Figure 4a) to facilitate the comparison. The short window removes the influence of the long-term trend present in the modern record (~0.35‰ decrease (0.5°C increase) over the 140 years). The 9.5 ka coral δ^{18}O standard deviation and the range (0.21 ± 0.03‰ (1σ); 1.07‰ range) are within the observed values for 9 year windows of the modern coral δ^{18}O (average standard deviation = 0.21 ± 0.03‰ (1σ); average range = 0.87 ± 0.15‰ (1σ); n = 129; Figure 4a).

[22] There are methodological differences between the modern and 9.5 ka coral studies to be considered outside their similarities namely common species and similar mean extension rates (1.1 cm/yr). The Moorea coral was discretely sampled with ~6 samples/yr [*Boiseau et al.*, 1998] whereas this study used continuous ~12 samples/yr. Bimonthly sampling resolution may reduce variability compared to monthly sampling, whereas discrete sampling may increase variability compared with the time-averaging continuous sampling method. *Boiseau et al.* [1998] conducted a sampling resolution test in the youngest portion of the record that revealed no improvement for monthly versus bimonthly sampling; however, shifts may occur in older portions of the core [cf. *Boiseau et al.*, 1998; *Bessat and Buigues*, 2001]. An examination of the Moorea record, in 9 year segments, did not reveal a systematic decrease in range or standard deviation with age that would indicate a shift in interannual variability due to sampling resolution. However, the effect of discrete versus continuous sampling remains to be quantified. The Moorea study used a cleaning method

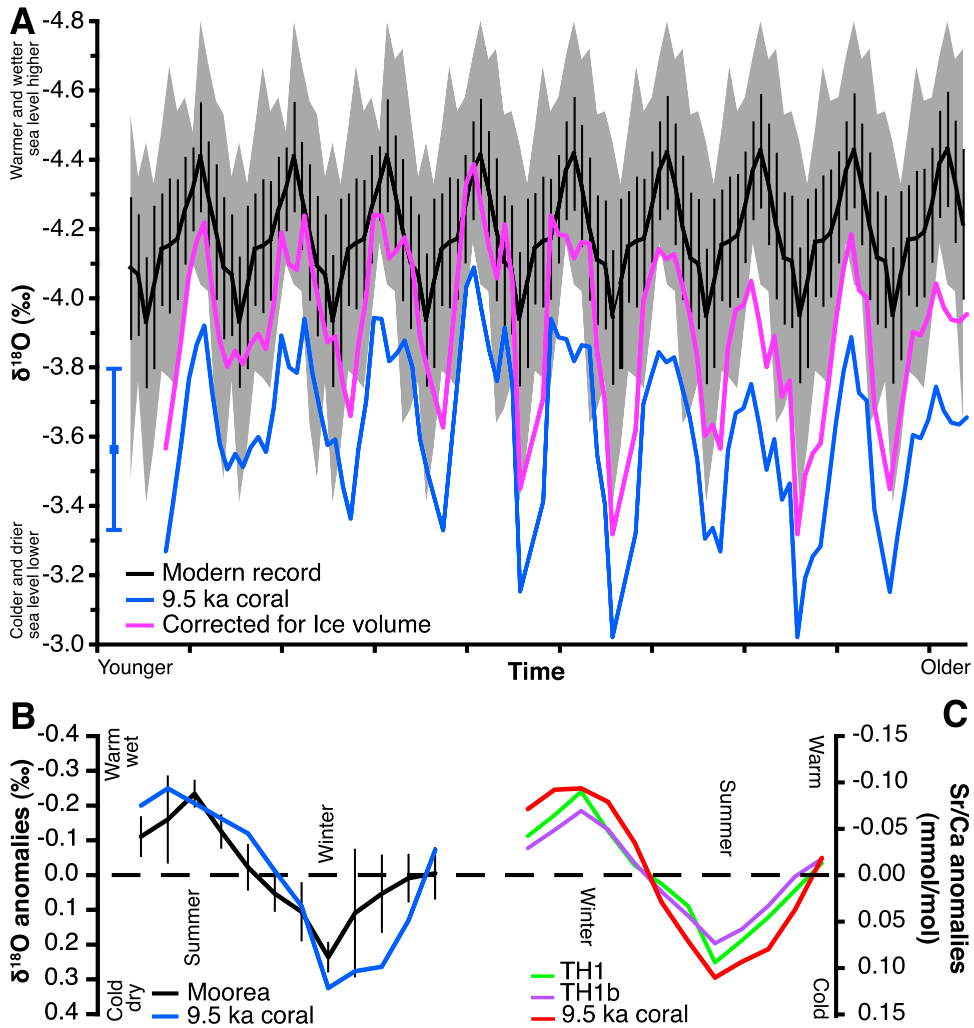

Figure 4. (a) The coral δ^{18}O variations for the 9 year long 9.5 ka coral and any 9 years from the modern Moorea coral [*Boiseau et al.*, 1998] shown as the average of the 9 year windows (−4.18 ± 0.11‰, 1σ; −3.99 minimum; −4.49 maximum; heavy black line) with an average standard deviation (0.21 ± 0.03‰,1σ; error bars) and average range (0.87 ± 0.15‰, 1σ; gray shaded area). The fossil coral δ^{18}O average is −3.63 ± 0.21‰ (1σ; blue error bars) with a range of 1.07‰. The fossil coral δ^{18}O was corrected for ice volume (0.30‰; see section 4.3). (b) The monthly means for the coral δ^{18}O records in Figure 4a. The summer minimums are offset as different months were used in each study for assigning chronologies. (c) The monthly means for coral Sr/Ca from the 9.5 ka coral and the modern Tahiti corals (TH1 and TH1b, averaged from 1974 to 1995 (S. Cahyarini, personal communication, 2009)).

described by *Boiseau and Juillet-Leclerc* [1997] that increased the interannual variability (~0.05‰) [see *Boiseau and Juillet-Leclerc*, 1997, Figure 7]; therefore, the interannual coral δ^{18}O variability may be greater in the early Holocene coral than in the modern.

[23] An examination of the seasonal variability in the Moorea coral δ^{18}O reveals slightly larger seasonality in the 9.5 ka coral (Figure 4b) indicating an increase in seasonal temperature variability in the early Holocene. To discern SST variability from δ^{18}Osw, we compared monthly mean coral Sr/Ca between the 9.5 ka coral and two modern intracolony cores (TH1 vertical, TH1b horizontal) from Teahupoo, Tahiti (Figure 4c) (S. Cahyarini, personal communication, 2009). The vertical and horizontal cores from the same colony provided a measure of variability regardless of growth direction. The horizontal core (TH1b) has reduced annual amplitude compared with the vertical core, and the 9.5 ka coral has greater seasonal amplitude than both modern cores (Figure 4c). A coral Sr/Ca reconstruction for 9688 BP from Espiritu Santo, Vanuatu [*Beck et al.*, 1997] and 14.2 and 12.5 ka from Tahiti [*Asami et al.*, 2009], found similar results for the seasonal cycle compared with the

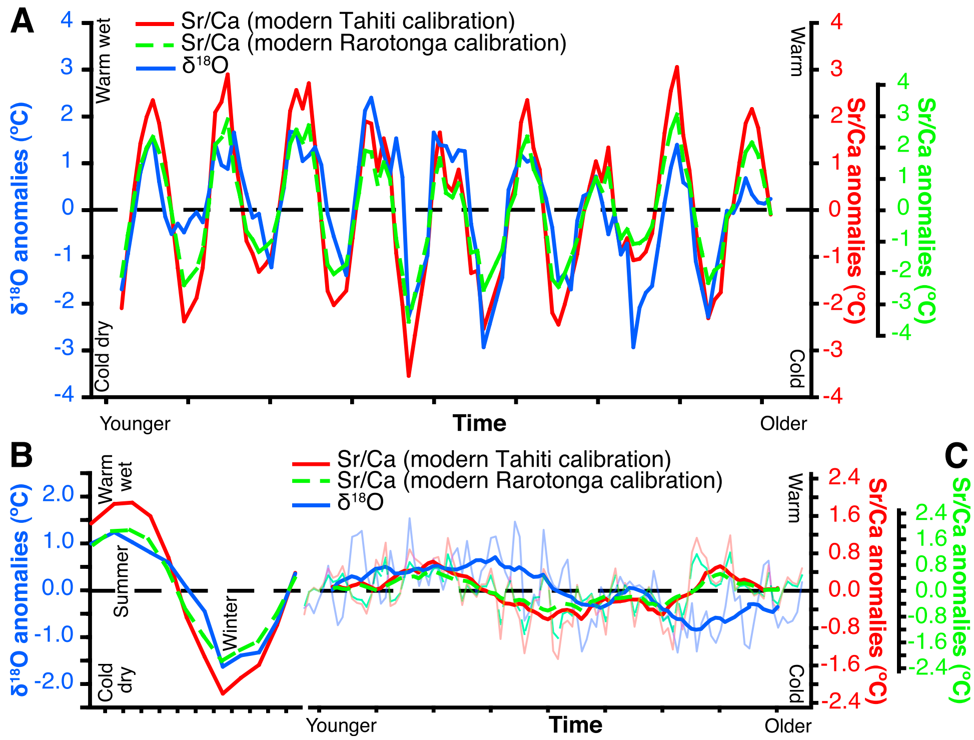

Figure 5. The coral Sr/Ca and δ^{18}O variations from the 9.5 ka coral with respective means removed shown as (a) monthly variations, (b) monthly mean, and (c) monthly anomalies. Monthly anomalies determined by removing monthly mean for each respective month. Monthly anomalies were smoothed with a 12 month moving average as some of the variability in monthly anomalies is the result of inaccuracies in the intra-annual chronology. Axes are scaled to the same temperature range relative to δ^{18}O (−0.2‰/°C; 0.05 mmol/mol/°C (red) for Tahiti [*Cahyarini et al.*, 2009] and 0.075 mmol/mol/°C (green) for Rarotonga).

modern. These results indicate the seasonal SST variability in the early Holocene was 0.4–1.2°C greater than modern (assuming 0.05°C/mmol/mol [*Cahyarini et al.*, 2009]); the ~1.2°C increase in seasonality is unreasonably large. The regional mean slope for coral Sr/Ca (equation (2)) provides a more reasonable increase (0.2–0.7°C). We view this result with caution since the modern Tahiti data were provided as monthly means and the methodological differences and the environmental uncertainties of the fossil coral, noted previously, may alter this result. Model studies for 9 ka driven by the Earth's orbital parameters indicate the seasonal cycle in the tropical southern hemisphere would be slightly less than the modern in the early Holocene [*Kutzbach and Gallimore*, 1988].

[24] The δ^{18}Osw can be determined using paired analysis of coral δ^{18}O and Sr/Ca [e.g., *Gagan et al.*, 1998; *Shen et al.*, 2005]. For this study, the lack of a modern calibration equation for coral Sr/Ca hinders the direct determination of absolute δ^{18}Osw change. We considered the range of Sr/Ca and δ^{18}O slopes noted earlier and ice volume corrections and found the hydrology for Tahiti in the early Holocene is close to modern conditions or drier. We compared the interannual and seasonal temperature variability between monthly coral Sr/Ca and δ^{18}O in the fossil coral for a range of coral Sr/Ca to SST slopes, assuming 0.20‰/°C for coral δ^{18}O (Figure 5). The slope for Rarotonga (0.075 mmol/mol/°C) produces similar seasonal temperature variability for coral Sr/Ca and δ^{18}O (~2.7°C; Figure 5b), whereas the slope for Tahiti (0.05 mmol/mol/°C) [*Cahyarini et al.*, 2009] produces larger seasonal variability (~4.0°C), a 60% increase from the modern SST record. Paired analysis of coral Sr/Ca and δ^{18}O using the range of slopes indicates the summers (winters) were drier (wetter) or close to modern, opposite of the seasonality observed in the modern instrumental record. Examination of the monthly anomalies (Figure 5c) reveals the interannual anomalies for this 9 year window are colder (warmer) and wetter (drier), opposite of those observed in modern records and modern ENSO events, and the anomalies persist for 1 year. The uncertainty in the modern calibration for Tahiti and our short 9 year window precludes our defining ENSO variability in the early Holocene, yet these results may indicate a shift in the hydrological

Table A1. U-Th Dating

Sample	Weight (g)	^{238}U (ppb)	^{232}Th (ppt)	δ^{234}U Measured[a]	[^{230}Th/^{238}U] Activity[b]	[^{230}Th/^{232}Th][c] (ppm)	Age Uncorrected	Age Corrected[b,d]	$\delta^{234}U_{initial}$ Corrected[e]
7B.11.R2	0.2581	2791.0 ± 2.3	1935.9 ± 4.0	140.5 ± 1.7	0.09555 ± 0.00024	2275 ± 7	9539 ± 29	9523 ± 33	144.3 ± 1.7

[a] $\delta^{234}U = ([^{234}U/^{238}U]_{activity} - 1) \times 1000$.
[b] $[^{230}Th/^{238}U]_{activity} = 1 - e^{-\lambda_{230}T} + (\delta^{234}U_{measured}/1000)\,[\lambda_{230}/(\lambda_{230} - \lambda_{234})](1 - e^{-(\lambda_{230}-\lambda_{234})T})$, where T is the age. Decay constants are 9.1577×10^{-6} yr^{-1} for ^{230}Th, 2.8263×10^{-6} yr^{-1} for ^{234}U [*Cheng et al.*, 2000], and 1.55125×10^{-10} yr^{-1} for ^{238}U [*Jaffey et al.*, 1971].
[c] The degree of detrital ^{230}Th contamination is indicated by the [^{230}Th/^{232}Th] atomic ratio instead of the activity ratio.
[d] Age corrections were calculated using an estimated atomic ^{230}Th/^{232}Th ratio of 4 ± 4 ppm.
[e] $\delta^{234}U_{initial}$ corrected was calculated based on ^{230}Th age (T), i.e., $\delta^{234}U_{initial} = \delta^{234}U_{measured} \times e^{\lambda_{234} * T}$, and T is corrected age.

cycle for Tahiti. These results are similar to those found in a 9 ka global modeling study [*Kutzbach and Gallimore*, 1988] that found decrease precipitation over the southern hemisphere oceans and increased precipitation over land associated with enhanced northern summer monsoon [also see *Marzin and Braconnot*, 2009] and little change in the seasonal variability of SST and precipitation.

5. Conclusions

[25] In lieu of a modern in situ calibration for Tahiti, we applied a calibration determined for mean coral Sr/Ca from *Porites* colonies from the southwest Pacific that estimated the early Holocene as ~3.2°C cooler than modern SST and greater than that predicted by modeling studies. We recognized a large uncertainty in coral Sr/Ca is related to or the combined effects of differences observed between corals from the same region, various water depths, local environmental conditions, and possible undetected diagenetic alteration. The interannual variability in the 9.5 ka coral δ^{18}O is similar to the interannual variability observed in a modern *Porites* study from nearby Moorea with slightly greater seasonality. A comparison of coral Sr/Ca reveals the seasonal cycle in the 9.5 ka coral is the same or greater than that observed in the modern Tahiti records. Other coral studies for the early Holocene and deglacial found little reduction in seasonal cycle whereas a model study for the early Holocene found little or no reduction in seasonality for the tropics. A comparison of monthly fossil coral δ^{18}O and Sr/Ca reveals that the seasonal hydrological cycle is approximately modern or drier with interannual anomalies that is colder (warmer) and wetter (drier); different from those observed in the modern instrumental record. Further studies are needed to investigate the tropical ocean climate dynamics in the early Holocene.

Appendix A

[26] Chemistry [*Shen et al.*, 2003] was performed on 18 November 2008 for U-Th isotopic determination on MC-ICP-MS [*Frohlich et al.*, 2009] (Table A1). Analytical errors are 2σ of the mean.

Acknowledgments

[27] This material is based upon work supported by the National Science Foundation under a Cooperative Agreement with Joint Oceanographic Institutions for a U.S. Science Support Program associated with the IODP. This research used

samples and/or data provided by the IODP. We thank Tony Greco for assistance with SEM images, Tim Ourbak for assistance on the project, Ethan Goddard for analytical assistance, Brad Linsley for coral data from Fiji and Rarotonga, and Yuda Cahyarini for providing coral geochemical summaries. Funding for U-Th isotopic determinations in the HISPEC was provided by Taiwan ROC NSC grants (95-2752-M002-012-011-PAE and 97-2745-M-002-009 to C.C.S.).

References

Alibert, C., and M. T. McCulloch (1997), Strontium/calcium ratios in modern *Porites* corals from the Great Barrier Reef as a proxy for sea surface temperature: Calibration of the thermometer and monitoring of ENSO, *Paleoceanography*, *12*, 345–363, doi:10.1029/97PA00318.

Allison, N., A. A. Finch, S. R. Sutton, and M. Newville (2001), Strontium heterogeneity and speciation in coral aragonite: Implications for the strontium paleothermometer, *Geochim. Cosmochim. Acta*, *65*, 2669–2676, doi:10.1016/S0016-7037(01)00628-7.

Allison, N., A. A. Finch, A. W. Tudhope, M. Newville, S. R. Sutton, and R. M. Ellam (2005), Reconstruction of deglacial sea surface temperatures in the tropical Pacific from selective analysis of a fossil coral, *Geophys. Res. Lett.*, *32*, L17609, doi:10.1029/2005GL023183.

Allison, N., A. A. Finch, J. M. Webster, and D. A. Clague (2007), Palaeoenvironmental records from fossil corals: The effects of submarine diagenesis on temperature and climate estimates, *Geochim. Cosmochim. Acta*, *71*, 4693–4703, doi:10.1016/j.gca.2007.07.026.

Asami, R., T. Felis, P. Dechamps, K. Hanawa, Y. Iryu, E. Bard, N. Durand, and M. Murayama (2009), Evidence for tropical South Pacific climate change during the Younger Dryas and the Bølling-Allerød from geochemical records of fossil Tahiti corals, *Earth Planet. Sci. Lett.*, *288*, 96–107, doi:10.1016/j.epsl.2009.09.011.

Beck, J. W., J. Récy, F. W. Taylor, R. L. Edwards, and G. Cabioch (1997), Abrupt changes in early Holocene tropical sea surface temperature derived from coral records, *Nature*, *385*, 705–707, doi:10.1038/385705a0.

Bessat, F., and D. Buigues (2001), Two centuries of variation in coral growth in a massive *Porites* colony from Moorea (French Polynesia): A response of ocean-atmosphere variability from south central Pacific, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *175*, 381–392, doi:10.1016/S0031-0182(01)00381-9.

Boiseau, M., and A. Juillet-Leclerc (1997), H_2O_2 treatment of recent coral aragonite: Oxygen and carbon isotopic implications, *Chem. Geol.*, *143*, 171–180, doi:10.1016/S0009-2541(97)00112-5.

Boiseau, M., A. Juillet-Leclerc, P. Yiou, B. Salvat, P. Isdale, and M. Guillaume (1998), Atmospheric and oceanic evidences of El Niño–Southern Oscillation events in the south central Pacific Ocean from coral stable isotopic records over the last 137 years, *Paleoceanography*, *13*, 671–685, doi:10.1029/98PA02502.

Cahyarini, S. Y. (2006), Paired δ^{18}O and Sr/Ca records of *Porites* corals from Tahiti (French Polynesia) and Timor (Indonesia), Ph.D. thesis, 179 pp., der Christian-Albrechts-Univ. zu Kiel, Kiel, Germany.

Cahyarini, S. Y., M. Pfeiffer, and W. C. Dullo (2009), Improving SST reconstructions from coral Sr/Ca records: Multiple corals from Tahiti (French Polynesia), *Int. J. Earth Sci.*, *98*, 31–40, doi:10.1007/s00531-008-0323-2.

Camoin, G. F., Y. Iryu, D. B. McInroy, and the Expedition 310 Scientists (2007), *Tahiti Sea Level*, *Proc. Integrated Ocean Drill. Program*, 310.

Cheng, H., R. L. Edwards, J. Hoff, C. D. Gallup, D. A. Richards, and Y. Asmerom (2000), The half-lives of uranium-234 and thorium-230, *Chem. Geol.*, *169*, 17–33, doi:10.1016/S0009-2541(99)00157-6.

Cobb, K. M., C. D. Charles, H. Cheng, and R. L. Edwards (2003), El Niño/Southern Oscillation and tropical Pacific climate during the last millennium, *Nature*, *424*, 271–276, doi:10.1038/nature01779.

Cohen, A. L., and S. R. Hart (1997), The effect of colony topography on climate signals in coral skeleton, *Geochim. Cosmochim. Acta*, *61*, 3905, doi:10.1016/S0016-7037(97)00200-7.

Cohen, A. L., and S. R. Hart (2004), Deglacial sea surface temperatures of the western tropical Pacific: A new look at old coral, *Paleoceanography*, *19*, PA4031, doi:10.1029/2004PA001084.

Cohen, A. L., G. D. Layne, S. R. Hart, and P. S. Lobel (2001), Kinetic control of skeletal Sr/Ca in a symbiotic coral: Implications for the paleotemperature proxy, *Paleoceanography*, *16*, 20–26, doi:10.1029/1999PA000478.

Conkright, M. E., R. A. Locarnini, H. E. Garcia, T. D. O'Brien, T. P. Boyer, C. Stephens, and J. I. Antonov (2002), World Ocean Atlas 2001: Objective analyses, data statistics, and figures—CD-ROM documentation, *Internal Rep., 17*, 17 pp., Natl. Oceanogr. Data Cent., Silver Spring, Md.

Crowley, T. J., T. M. Quinn, and W. T. Hyde (1999), Validation of coral temperature calibrations, *Paleoceanography*, *14*, 605–615, doi:10.1029/1999PA900032.

DeLong, K. L. (2008), Paleoclimatic reconstruction and evaluation of sub-centennial climate variability in the late Holocene using records from massive corals (New Caledonia), tree-rings (New Mexico) and speleothems (China), Ph.D. thesis, 160 pp., Univ. of South Fla., Tampa.

DeLong, K. L., T. M. Quinn, and F. W. Taylor (2007), Reconstructing twentieth-century sea surface temperature variability in the southwest Pacific: A replication study using multiple coral Sr/Ca records from New Caledonia, *Paleoceanography*, *22*, PA4212, doi:10.1029/2007PA001444.

de Villiers, S., G. T. Shen, and B. K. Nelson (1994), The Sr/Ca-temperature relationship in coralline aragonite: Influence of variability in (Sr/Ca) seawater and skeletal growth parameters, *Geochim. Cosmochim. Acta*, *58*, 197, doi:10.1016/0016-7037(94)90457-X.

de Villiers, S., B. K. Nelson, and A. R. Chivas (1995), Biological controls on coral Sr/Ca and δ^{18}O reconstructions of sea surface temperatures, *Science*, *269*, 1247–1249, doi:10.1126/science.269.5228.1247.

Dunn, E.M., T.M. Quinn, F.W. Taylor, J.A. Austin, G. Cabioch, B. Pelletier, C. Maes, and V. Ballu (2008), Coral-based climate reconstructions from a massive *Porites* coral from a mid-Pacific shallowly submerged atoll: Sabine Bank (Vanuatu), *Eos Trans. AGU*, *89*(53), Fall Meet. Suppl., Abstract PP11C-1398.

Fairbanks, R. G., and R. K. Matthews (1978), Marine oxygen isotope record in Pleistocene coral, Barbados, West Indies, *Quat. Res.*, *10*, 181–196, doi:10.1016/0033-5894(78)90100-X.

Felis, T., G. Lohmann, H. Kuhnert, S. J. Lorenz, D. Scholz, J. Pätzold, S. A. Al-Rousan, and S. M. Al-Moghrabi

(2004), Increased seasonality in Middle East temperatures during the last interglacial period, *Nature*, *429*, 164, doi:10.1038/nature02546.

Fleming, K., P. Johnston, D. Zwartz, Y. Yokoyama, K. Lambeck, and J. Chappell (1998), Refining the eustatic sea-level curve since the Last Glacial Maximum using far- and intermediate-field sites, *Earth Planet. Sci. Lett.*, *163*, 327–342, doi:10.1016/S0012-821X(98)00198-8.

Frohlich, C., M. J. Hornbach, F. W. Taylor, C.-C. Shen, A. Moala, A. E. Morton, and J. Kruger (2009), Huge erratic boulders in Tonga deposited by a prehistoric tsunami, *Geology*, *37*, 131–134, doi:10.1130/G25277A.1.

Gagan, M. K., L. K. Ayliffe, D. Hopley, J. A. Cali, G. E. Mortimer, J. Chappell, M. T. McCulloch, and M. J. Head (1998), Temperature and surface-ocean water balance of the mid-Holocene tropical western Pacific, *Science*, *279*, 1014–1018, doi:10.1126/science.279.5353.1014.

Gagan, M. K., E. J. Hendy, S. G. Haberle, and W. S. Hantoro (2004), Post-glacial evolution of the Indo-Pacific Warm Pool and El Niño–Southern Oscillation, *Quat. Int.*, *118–119*, 127–143, doi:10.1016/S1040-6182(03)00134-4.

Gouriou, Y., and T. Delcroix (2002), Seasonal and ENSO variations of sea surface salinity and temperature in the South Pacific Convergence Zone during 1976–2000, *J. Geophys. Res.*, *107*(C12), 3185, doi:10.1029/2001JC000830.

Heiss, G. A., W.-C. Dullo, M. M. Joachimski, J. J. G. Reijmer, and H. Schuhmacher (1999), Increased seasonality in the Gulf of Aqaba, Red Sea, recorded in the oxygen isotope record of a *Porites lutea* coral, *Mar. Biodiversity*, *30*, 17–26, doi:10.1007/BF03042826.

Hendy, E., M. K. Gagan, J. M. Lough, M. McCulloch, and P. B. deMenocal (2007), Impact of skeletal dissolution and secondary aragonite on trace element and isotopic climate proxies in *Porites* corals, *Paleoceanography*, *22*, PA4101, doi:10.1029/2007PA001462.

Jaffey, A. H., K. F. Flynn, L. E. Glendenin, W. C. Bentley, and A. M. Essling (1971), Precision measurement of half-lives and specific activities of ^{235}U and ^{238}U, *Phys. Rev. C*, *4*, 1889–1906, doi:10.1103/PhysRevC.4.1889.

Kilbourne, K. H., T. M. Quinn, and F. W. Taylor (2004), A fossil coral perspective on western tropical Pacific climate ~350 ka, *Paleoceanography*, *19*, PA1019, doi:10.1029/2003PA000944.

Kutzbach, J. E., and R. G. Gallimore (1988), Sensitivity of a coupled atmosphere/mixed layer ocean model to changes in orbital forcing at 9000 years B.P, *J. Geophys. Res.*, *93*, 803–821, doi:10.1029/JD093iD01p00803.

Lea, D. W., D. K. Pak, and H. J. Spero (2000), Climate impact of late Quaternary equatorial Pacific sea surface temperature variations, *Science*, *289*, 1719–1724, doi:10.1126/science.289.5485.1719.

Linsley, B. K., G. M. Wellington, and D. P. Schrag (2000), Decadal sea surface temperature variability in the subtropical South Pacific from 1726 to 1997 AD, *Science*, *290*, 1145–1148, doi:10.1126/science.290.5494.1145.

Linsley, B. K., A. Kaplan, Y. Gouriou, J. Salinger, P. B. deMenocal, G. M. Wellington, and S. S. Howe (2006), Tracking the extent of the South Pacific convergence zone since the early 1600s, *Geochem. Geophys. Geosyst.*, *7*, Q05003, doi:10.1029/2005GC001115.

Liu, J., T. J. Crowley, and T. M. Quinn (2005), Breakdown of Sr/Ca paleothermometer in a coral record from New Georgia, West Pacific warm pool, *Eos Trans. AGU*, *86*(52), Fall Meet. Suppl., Abstract PP44A-07.

Marzin, C., and P. Braconnot (2009), Variations of Indian and African monsoons induced by insolation changes at 6 and 9.5 kyr BP, *Clim. Dyn.*, *33*, 215–231, doi:10.1007/s00382-009-0538-3.

McConnaughey, T. (1989), ^{13}C and ^{18}O isotopic disequilibrium in biological carbonates: I. Patterns, *Geochim. Cosmochim. Acta*, *53*, 151–162, doi:10.1016/0016-7037(89)90282-2.

Mitsuguchi, T., E. Matsumoto, and T. Uchida (2003), Mg/Ca and Sr/Ca ratios of *Porites* coral skeleton: Evaluation of the effect of skeletal growth rate, *Coral Reefs*, *22*, 381–388, doi:10.1007/s00338-003-0326-1.

Montaggioni, L. F., F. Le Cornec, T. Corrège, and G. Cabioch (2006), Coral barium/calcium record of mid-Holocene upwelling activity in New Caledonia, South-West Pacific, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *237*, 436–455, doi:10.1016/j.palaeo.2005.12.018.

ORSTOM (1993), *Atlas de la polynésie Francaise*, Paris.

Quinn, T. M., and F. W. Taylor (2006), SST artifacts in coral proxy records produced by early marine diagenesis in a modern coral from Rabaul, Papua New Guinea, *Geophys. Res. Lett.*, *33*, L04601, doi:10.1029/2005GL024972.

Quinn, T. M., F. W. Taylor, T. J. Crowley, and S. M. Link (1996), Evaluation of sampling resolution in coral stable isotope records: A case study using records from New Caledonia and Tarawa, *Paleoceanography*, *11*, 529–542, doi:10.1029/96PA01859.

Quinn, T. M., T. J. Crowley, F. W. Taylor, C. Hénin, P. Joannot, and Y. Join (1998), A multicentury stable isotope record from a New Caledonia coral: Interannual and decadal sea surface temperature variability in the southwest Pacific since 1657 A.D., *Paleoceanography*, *13*, 412–426, doi:10.1029/98PA00401.

Quinn, T. M., F. W. Taylor, and T. J. Crowley (2006), Coral-based climate variability in the Western Pacific Warm Pool since 1867, *J. Geophys. Res.*, *111*, C11006, doi:10.1029/2005JC003243.

Rayner, N. A., D. E. Parker, E. B. Horton, C. K. Folland, L. V. Alexander, D. P. Rowell, E. C. Kent, and A. Kaplan (2003), Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century, *J. Geophys. Res.*, *108*(D14), 4407, doi:10.1029/2002JD002670.

Reynolds, R. W., N. A. Rayner, T. M. Smith, D. C. Stokes, and W. Wang (2002), An improved in situ and satellite SST analysis for climate, *J. Clim.*, *15*, 1609–1625, doi:10.1175/1520-0442(2002)015<1609:AIISAS>2.0.CO;2.

Rougerie, F. (1994), The Polynesian south ocean: Features and circulation, *Mar. Pollut. Bull.*, *29*, 14–25, doi:10.1016/0025-326X(94)90421-9.

Schrag, D. P. (1999), Rapid analysis of high-precision Sr/Ca ratios in corals and other marine carbonates, *Paleoceanography*, *14*, 97–102, doi:10.1029/1998PA900025.

Schrag, D. P., J. F. Adkins, K. McIntyre, J. L. Alexander, D. A. Hodell, C. D. Charles, and J. F. McManus (2002), The oxygen isotopic composition of seawater during the Last Glacial Maximum, *Quat. Sci. Rev.*, *21*, 331–342, doi:10.1016/S0277-3791(01)00110-X.

Shen, C.-C., H. Cheng, R. L. Edwards, S. B. Moran, H. N. Edmonds, J. A. Hoff, and R. B. Thomas (2003), Measurement of attogram quantities of ^{231}Pa in dissolved and particulate fractions of seawater by isotope dilution thermal ionization mass spectroscopy, *Anal. Chem.*, *75*, 1075–1079, doi:10.1021/ac026247r.

Shen, C.-C., T. Lee, K.-K. Liu, H.-H. Hsu, R. L. Edwards, C.-H. Wang, M.-Y. Lee, Y.-G. Chen, H.-J. Lee, and H.-T. Sun (2005), An evaluation of quantitative reconstruction of past precipitation records using coral skeletal Sr/Ca and $\delta^{18}O$ data, *Earth Planet. Sci. Lett.*, *237*, 370, doi:10.1016/j.epsl.2005.06.042.

Shen, C.-C., et al. (2008), Variation of initial $^{230}Th/^{232}Th$ and limits of high precision U–Th dating of shallow-water corals, *Geochim. Cosmochim. Acta*, *72*, 4201–4223, doi:10.1016/j.gca.2008.06.011.

Slutz, R. J., S. J. Lubker, J. D. Hiscox, S. D. Woodruff, R. L. Jenne, P. M. Steurer, and J. D. Elms (1985), Comprehensive Ocean-Atmosphere Data Set, release 1, *Rep. NTIS PB86-105723*, Clim. Res. Program, Environ. Res. Lab., NOAA, Boulder, Colo.

Stephans, C. L., T. M. Quinn, F. W. Taylor, and T. Corrège (2004), Assessing the reproducibility of coral-based climate records, *Geophys. Res. Lett.*, *31*, L18210, doi:10.1029/2004GL020343.

Stoll, H. M., and D. P. Schrag (1998), Effects of Quaternary sea level cycles on strontium in seawater, *Geochim. Cosmochim. Acta*, *62*, 1107–1118, doi:10.1016/S0016-7037(98)00042-8.

Visser, K., R. Thunell, and L. Stott (2003), Magnitude and timing of temperature change in the Indo-Pacific warm pool during deglaciation, *Nature*, *421*, 152, doi:10.1038/nature01297.

Wolanski, E., and B. Delesalle (1995), Upwelling by internal waves, Tahiti, French Polynesia, *Cont. Shelf Res.*, *15*, 357–368, doi:10.1016/0278-4343(93)E0004-R.

York, D., and N. M. Evensen (2004), Unified equations for the slope, intercept and standard errors of the best straight line, *Am. J. Phys.*, *72*, 367–375, doi:10.1119/1.1632486.