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Tracking the extent of the South Pacific Convergence Zone since the early 1600s

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[1] The South Pacific Convergence Zone (SPCZ) is the largest and most persistent spur of the Intertropical Convergence Zone. At the southeastern edge of the SPCZ near 170°W and 15°–20°S a surface ocean salinity frontal zone exists that separates fresher Western Pacific Warm Pool water from saltier and cooler waters in the east. This salinity front is known to shift east and west with the phase of the El Niño Southern Oscillation. We have generated subannually resolved and replicated coral oxygen isotopic time series from Fiji (17°S, 179°E) and Rarotonga (21.5°S, 160°W) that have recorded interannual displacements of the salinity front over the last 380 years and also indicate that at lower frequencies the decadal mean position of the salinity front, and eastern extent of the SPCZ, has shifted east-west through 10° to 20° of longitude three times during this interval. The most recent and largest shift began in the mid 1800s as the salinity front progressively moved eastward and salinity decreased at both sites. Our results suggest that sea surface salinity at these sites is now at the lowest levels recorded and is evidence for an unprecedented expansion of the SPCZ since the mid 1800s. The expansion of the SPCZ implies a gradual change in the South Pacific to more La Niña-like long-term mean conditions. This observation is consistent with the ocean thermostat mechanism for the Pacific coupled ocean-atmosphere system, whereby exogenous heating of the atmosphere would result in greater warming in the western Pacific and a greater east-west surface temperature gradient.

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1. Introduction

[2] In the South Pacific sector of the Western Pacific Warm Pool (WPWP), a 200–400 km broad zone of low-level wind convergence, high cloudiness, and enhanced precipitation is known as the South Pacific Convergence Zone (SPCZ) [*Trenberth*, 1976; *Kiladis et al.*, 1989; *Vincent*, 1994] (Figure 1). The SPCZ stretches southeast from the Intertropical Convergence Zone (ITCZ) near the equatorial Solomon Islands to Fiji, Samoa, Tonga and further southeast. With the advent of satellite monitoring in the 1970s, the SPCZ is now known to be the largest and most persistent spur of the ITCZ. On seasonal timescales the SPCZ is most active in the southern hemisphere summer, but it is present year-round. On interannual timescales the axis of maximum rainfall associated with the SPCZ is known to shift northeast during El Niño events and southwest during La Niña events [*Salinger et al.*, 2001; *Folland et al.*, 2002; *Gouriou and Delcroix*, 2002]. Limited instrumental data also indicate that on interdecadal timescales the SPCZ shifts mean position with the polarity of the Interdecadal Pacific Oscillation (IPO) [*Salinger et al.*, 2001; *Folland et al.*, 2002].

[3] Enhanced convection and rainfall in the oceanic sectors of the ITCZ is one predicted outcome of increased exogenous heating of the tropical atmosphere [*Held and Soden*, 2000; *Chou and Neelin*, 2004; *Kumar et al.*, 2004]. Another predicted outcome is an increase in the zonal sea surface temperature (SST) gradient across the equatorial Pacific [*Cane et al.*, 1997]. In regions with a deep thermocline like the WPWP, the mixed layer will adjust with an increased temperature in response to increased atmospheric heating. However, in the eastern Pacific, where the thermocline is shallow due to upwelling, the vertical advection of cool water tends to offset the effects of surface heating. The result would be a greater east-west SST gradient and more La Niña-like mean conditions and strengthened trade winds that in turn strengthens the upwelling in the eastern Pacific, further increasing the east-west temperature gradient [*Clement et al.*, 1996; *Cane et al.*, 1997]. Reconstructing past variability in the extent and position of the SPCZ may provide evidence about pre-anthropogenic "baseline" convergence zone variability and the response to twentieth century atmospheric heating in order to test these predictions.

[4] One means of examining the past extent and variability of the SPCZ is to reconstruct ocean properties that are directly related to the SPCZ today. In the surface ocean of the subtropical South Pacific a salinity front exists near the eastern edge of the WPWP [*Gouriou and Delcroix*, 2002] (Figure 1). This salinity front is currently located near 175°W and 15°–20°S [*Gouriou and Delcroix*, 2002], the result of the juxtaposition of relatively high salinity waters formed in the subtropical central gyre and relatively low salinity surface waters of the SPCZ region.

[5] Analysis of instrumental SST, sea surface salinity (SSS), and precipitation records in the SPCZ - SSS front region beginning in 1976 indicate that each contains an interannual signal that correlates with the Southern Oscillation Index (SOI) and El Niño Southern Oscillation (ENSO). In this region the amplitudes of the interannual signals in SST and precipitation are an order of magnitude less than the amplitude of the seasonal cycle, whereas for SSS, the interannual signal of 1 to 1.5 p.s.u. is double the amplitude of the seasonal signal [*Gouriou and Delcroix*, 2002]. These facts reflect the northeastward (southwestward) shift of the SPCZ during El Niño (La Niña) events. In the SPCZ region, SSS is higher and SST and precipitation are lower during El Niño events. The opposite occurs during La Niña events. More importantly, interannual displacements of the salinity front along a northwest-southeast oriented axis are in part due to zonal advection by oceanic currents and in part due to changes in the evaporation-precipitation (E-P) budget [*Gouriou and Delcroix*, 2002]. During an El Niño event the SPCZ rainfall maximum axis moves generally northeast making the E-P budget more positive. At the same time the South Equatorial Current (SEC)

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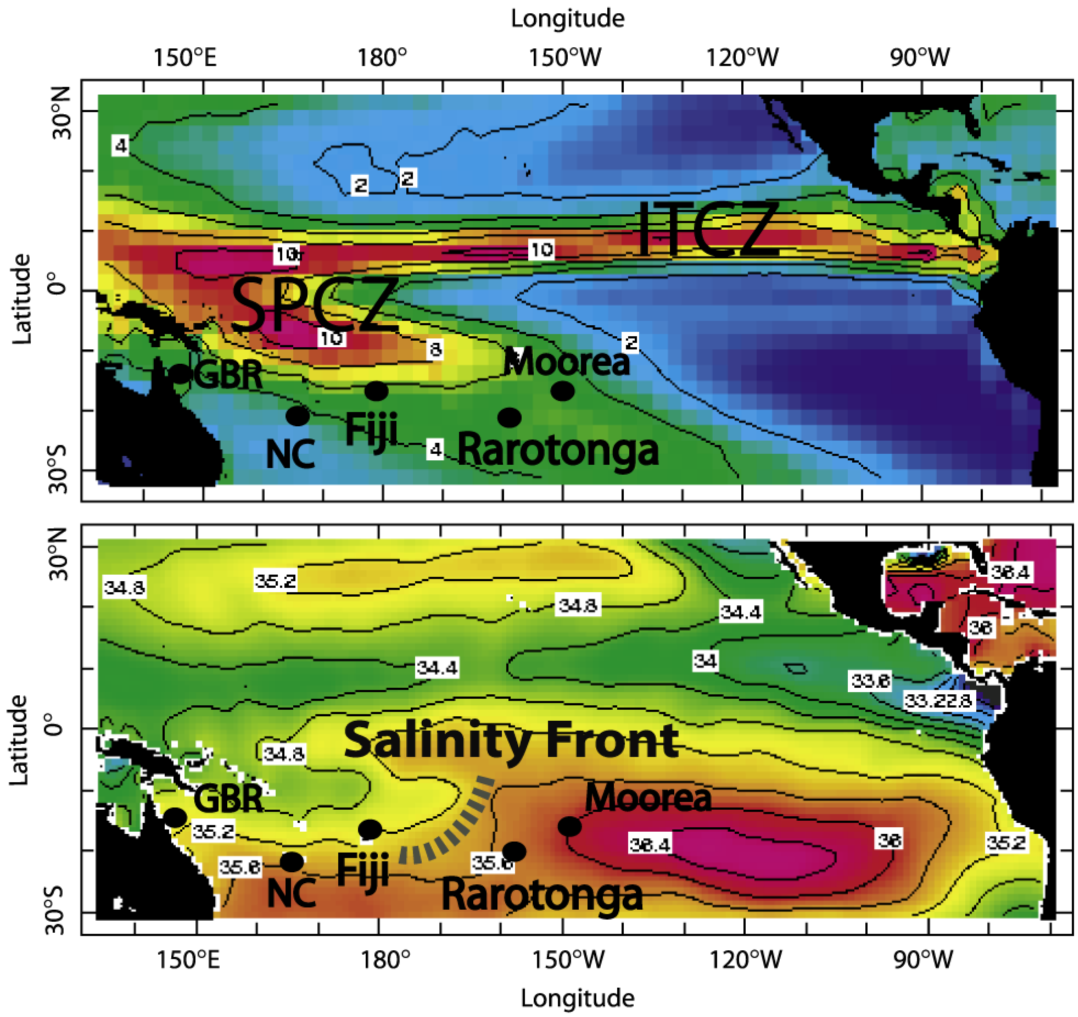

Figure 1. (top) Annual average precipitation (1979–1992) [*Xie and Arkin*, 1997] and (bottom) annual average salinity (1979–1992) [*Conkright et al.*, 2002] in the tropical Pacific Ocean. Precipitation is in units of mm per day. Salinity is in practical salinity units (p.s.u.) and is contoured in 0.4 p.s.u. intervals. Orange-red colors indicate (top) greater precipitation and (bottom) higher salinity. The locations of the study sites in Fiji and Rarotonga are indicated. The locations of coral records from New Caledonia (NC) [*Quinn et al.*, 1998], Moorea [*Boiseau et al.*, 1999], and the Great Barrier Reef (GBR) [*Hendy et al.*, 2002] are also indicated. The sea surface salinity frontal region, the South Pacific Convergence Zone (SPCZ), and Intertropical Convergence Zone (ITCZ) are labeled.

strengthens, advecting higher salinity waters from the east. The opposite happens during La Niña events. Due to the lack of SSS data prior to 1976, little is known about past changes in the position of this salinity front.

[6] Here we discuss oxygen isotopic ($\delta^{18}O$) time series generated from cores recovered from massive corals in the SPCZ region. We have generated two *Porites lutea* coral $\delta^{18}O$ records from Fiji (17°S, 179°E: spanning 1619–2001 C.E., replicated from 1780 C.E.; C.E. is "Common Era," which is equivalent to the term A.D.) and three from Rarotonga (21.5°S, 160°W: spanning 1726–2000 C.E., replicated from 1874 and 1906 C.E.), both situated in the salinity front region (Table 1). Our longest coral $\delta^{18}O$ time series is a new subannual record from Savusavu Bay in Fiji extending from 1619 to 2001 C.E. We will show that interannual, decadal, and secular trend components of variability in these Fiji and Rarotonga $\delta^{18}O$ series appear to be predominantly due to changes in surface salinity and movements of the salinity front since at least 1870. Comparison of the coral $\delta^{18}O$ series from Fiji and Rarotonga with previously published coral $\delta^{18}O$ records from Moorea [*Boiseau et al.*, 1999], New Caledonia [*Quinn et al.*, 1998], and the Great Barrier Reef [*Hendy et al.*, 2002] provide additional evidence of regional salinity variability in this region prior to 1870 and of movements in the decadal-mean

Table 1. Fiji and Rarotonga Porites Cores Analyzed

Core ID	Species	Length, years	$\delta^{18}O$	Sr/Ca
		Fiji, Savusavu Bay		
Core 1F	*Porites lutea*	1997–1780	millimeter-scale	millimeter-scale
Core AB	*Porites lutea*	2001–1619	millimeter-scale	annual
		Rarotonga		
Core 2R	*Porites lutea*	1997–1726	millimeter-scale	millimeter-scale
Core 3R	*Porites lutea*	2000–1874	millimeter-scale	annual
Core 99	*Porites lutea*	1999–1906	millimeter-scale	annual

position of the SSS front and SPCZ prior to anthropogenic forcing.

2. Methods

[7] Descriptions of our Rarotonga coral cores (designated 2R, 3R, and 99) and shorter Fiji core (designated 1F) are given by *Ren et al.* [2003] and *Linsley et al.* [2004]. In December 2001 we retrieved a 3.89 m long continuous coral core from a colony of *Porites lutea* growing at 10 m depth in Savusavu Bay, Vanua Levu, Fiji (16°49′S, 179°14′E) (within 200 m of core 1F collected in 1997). This core has been designated Fiji AB. Time series of skeletal $\delta^{18}O$ and Sr/Ca from this core are being presented here for the first time. The usable length of the core was 3.63 m and spans from 2001 to 1619 C.E.

[8] Coral slabs 6 mm thick were cut from the cores and cleaned in deionized water in an elongate ultrasonic bath. Dry slabs were sampled with a low-speed micro-drill with a 1-mm-round diamond drill bit along the maximum growth axis in tracks parallel to corallite traces as identified in X-ray positives. In all cases a 1-mm-deep by 2-mm-wide groove was excavated at 1 mm increments. We measured oxygen isotopes ($\delta^{18}O$), reported as per mil (‰) deviations relative to Vienna Peedee belemnite (VPDB), using a Micromass Optima gas-source triple collector isotope ratio mass spectrometer at the University at Albany, State University of New York equipped with an individual acid reaction vessel system. For all cores from Fiji and Rarotonga we measured $\delta^{18}O$ every 1 mm in the most recent 30 years of growth (12–13 samples per year), and below that on every other 1 mm increments (6–7 samples per year). Samples of international standard NBS-19 were interspersed among the coral samples in analytical runs. The average $\delta^{18}O$ and standard deviation of 433 samples of NBS-19 analyzed with Fiji core AB over a 6 month period was -2.203 ± 0.028‰, comparable to the averages and standard deviations of NBS-19 analyzed previously with the other Fiji and Rarotonga cores. The average difference between duplicate analyses of 936 Fiji and Rarotonga coral samples was better than 0.04‰ for $\delta^{18}O$.

[9] We have previously measured Sr/Ca at 1 mm resolution (12–13 samples per year) on one core from both Rarotonga (2R; 1997–1726) and Fiji (1F; 1997–1780) [*Linsley et al.*, 2000, 2004]. These analyses were done on an inductively coupled plasma optical emission spectrophotometer (ICP-OES) at Harvard University. The average relative standard deviation of replicate samples analyzed was 0.15%. As part of this new work we measured Sr/Ca ratios on annually averaged samples from Fiji core AB and Rarotonga cores 3R and 99 (see Table 1) using an ICP-OES at the Lamont Doherty Earth Observatory (LDEO) of Columbia University. To make the annually averaged samples, equal aliquots from 1 mm samples within one year (summer to summer defined by annual $\delta^{18}O$ cycles) were mixed and homogenized before analysis. The average relative standard deviation of replicate samples analyzed was 0.20% ($n = 101$). Analysis of replicate samples analyzed at both the Harvard and LDEO laboratories indicates an analytical offset of 0.14 mmol/mol. A 0.07mmol/mol correction has been applied to Sr/Ca data from both labs to account for this analytical offset.

[10] The chronologies for all of the coral results presented here are based on the reconstructed annual cycle in millimeter-scale $\delta^{18}O$ and/or Sr/Ca down core analyses [*Linsley et al.*, 2000; *Ren et al.*, 2003; *Linsley et al.*, 2004]. Briefly, because annual minimum and maximum millimeter-scale $\delta^{18}O$ and Sr/Ca were phased-locked at zero lag in the down core series from both sites we constructed the chronologies by setting the annual minima in near-monthly Sr/Ca and/or $\delta^{18}O$ to February (on average the warmest month) and maxima in Sr/Ca and/or $\delta^{18}O$ to August/September (on average the coolest months). For Fiji core AB and Rarotonga cores 99 and 3R, we only used millimeter-scale $\delta^{18}O$ to construct the chronologies. We cross-checked all chronologies against

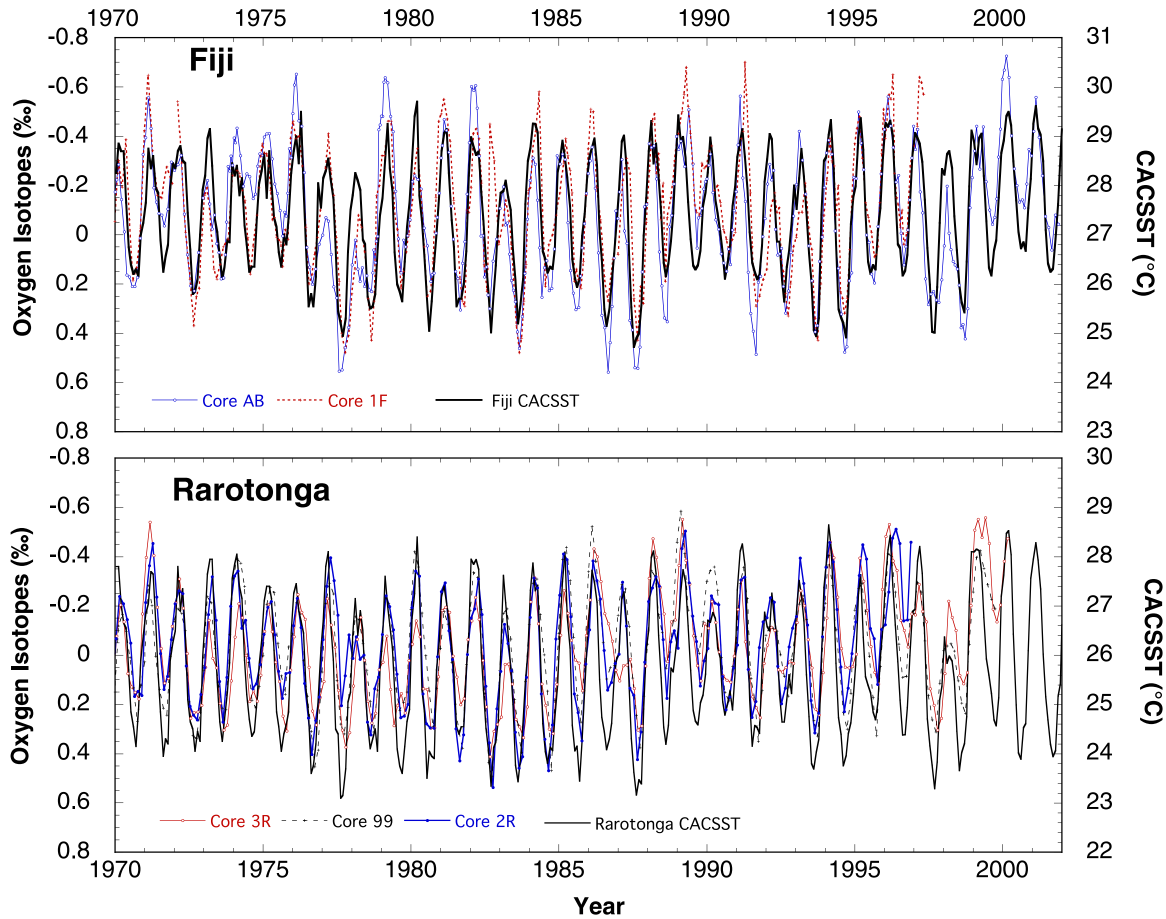

Figure 2. Fiji and Rarotonga millimeter-scale coral oxygen isotope ($\delta^{18}O$) data compared to monthly sea surface temperature (SST) from the NCEP OI SST database [*Reynolds and Smith*, 1994]. All $\delta^{18}O$ data have been centered by removing the twentieth century mean. (top) Centered $\delta^{18}O$ data from two *Porites* cores collected at 10 m water depth in Savusavu Bay, Fiji, compared to NCEP OI SST for the 2° × 2° latitude-longitude grid that includes Fiji. The two colonies were approximately 200 m from each other. (bottom) Centered $\delta^{18}O$ data from three *Porites* cores collected from Rarotonga compared to NCEP OI SST for the 2° × 2° latitude-longitude grid that includes Rarotonga. Core 2R was collected in 1997. Two other Rarotonga coral cores were collected in 1999 and 2000, one within 200 m of the first and at 18 m water depth and the other from a colony located on the north side of the island in 10 m of water near a small river outlet.

density bands observed in X-rays of slabs from the coral cores and also against times of known El Niño events based on the Niño 3.4 index that extends back to 1856 [*Kaplan et al.*, 1998].

3. Results

3.1. Oxygen Isotopes

[11] Due to offsets in twentieth century mean $\delta^{18}O$ in these coral skeletons of 0.37‰ (std. dev. = 0.11) at Fiji, and −0.10‰ (std. dev. = 0.11) and 0.08‰ (std. dev. = 0.10) at Rarotonga, we have removed the twentieth century mean $\delta^{18}O$ value from each coral time series (Figures 2, 3, and 4). Offsets in mean $\delta^{18}O$ value in the same *Porites* species analyzed at the same location have been previously observed [*Tudhope et al.*, 1995; *Linsley et al.*, 1999, 2004; *Cobb et al.*, 2003] and have been attributed to differences in disequilibrium "vital" effect offsets between coral colonies. It has been shown that once the mean $\delta^{18}O$ over a common time period has been subtracted (centered around 0.0‰) the $\delta^{18}O$ results are then directly comparable and highly correlated.

[12] At both Fiji and Rarotonga instrumental 2° × 2° latitude-longitude gridded SST [*Reynolds and Smith*, 1994] and 2° × 10° latitude-longitude gridded SSS [*Gouriou and Delcroix*, 2002] records reveal that each site has a pronounced seasonal SST cycle of 4° to 5°C and a weak seasonal SSS cycle. On interannual timescales the opposite takes place: a weak SST signal of 1° to 2°C and a larger 1 to 1.5 p.s.u. SSS signal (see Figures 2 and 3).

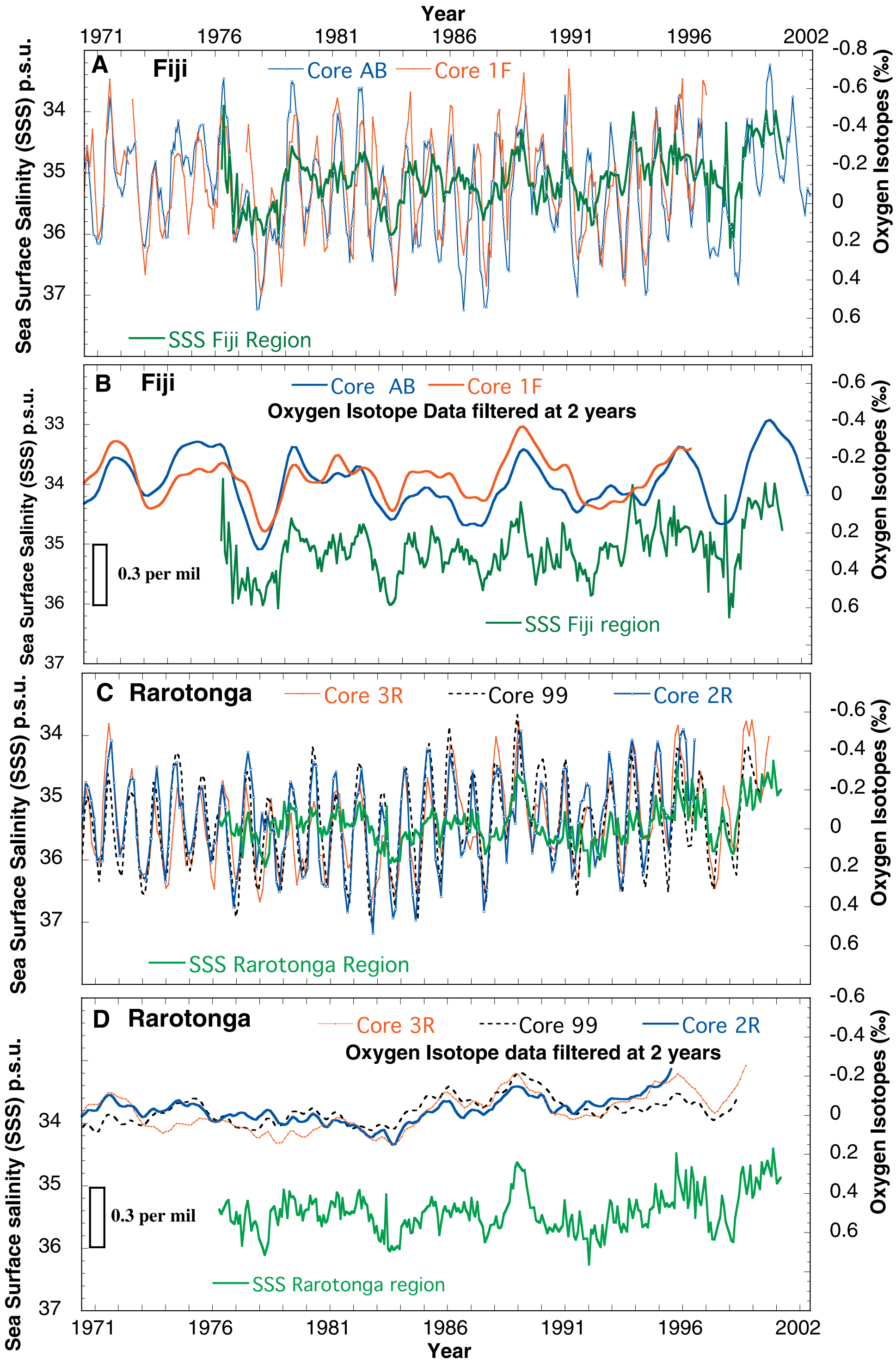

Figure 3. Comparison of Fiji and Rarotonga coral oxygen isotopic ($\delta^{18}O$) data to monthly sea surface salinity (SSS) from *Gouriou and Delcroix* [2002]. All $\delta^{18}O$ data have been centered by removing the twentieth century mean. (a) Subseasonal scale $\delta^{18}O$ from two cores collected in Savusavu Bay, Vanua Levu, Fiji, compared to monthly SSS for the 2° latitude × 10° longitude grid including Fiji (170°W–180°W; 16°S–18°S). (b) Fiji coral $\delta^{18}O$ band-pass-filtered at 2 years compared to the same Fiji monthly SSS. (c) Subseasonal scale $\delta^{18}O$ from three cores collected from Rarotonga compared to monthly SSS for the 2° latitude × 10° longitude grid including Rarotonga (160°W–170°W; 20°S–22°S). (d) Rarotonga coral $\delta^{18}O$ band-pass-filtered at 2 years compared to the same monthly SSS for Rarotonga. Bar in Figures 3b and 3d indicates the scaling of 0.3‰ per 1 p.s.u.

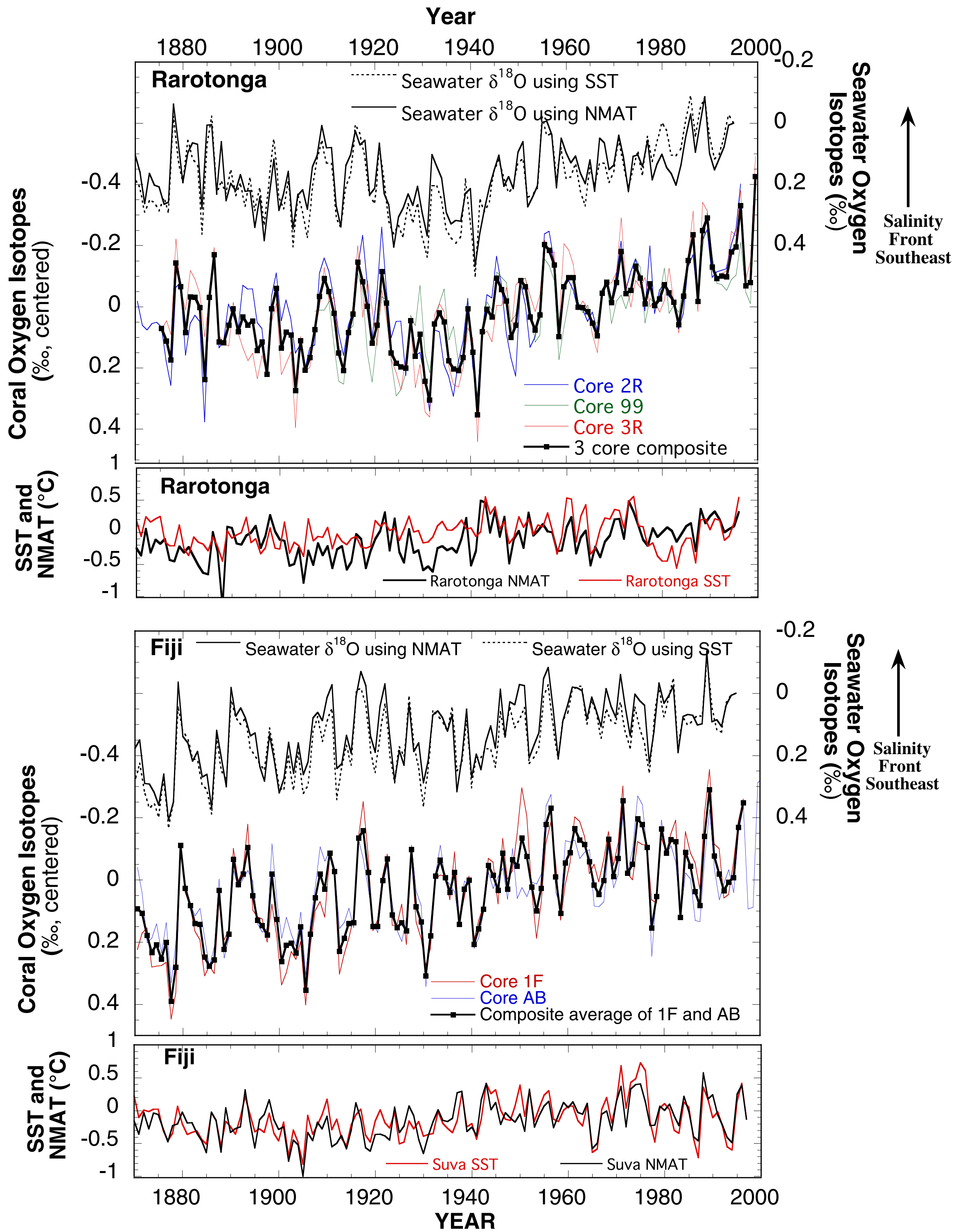

Figure 4. Rarotonga and Fiji annual average coral $\delta^{18}O$ results and calculated $\delta^{18}O_{seawater}$ using both nighttime marine air temperature anomaly (NMAT, from Hadley Centre Night Marine Air Temperature data set: HadMAT1) and SST anomaly (SSTa; 1° latitude by 1° longitude; from the Hadley Centre Sea Ice and Sea Surface Temperature database: HadISST1) for both the Rarotonga and Fiji regions and as shown in Figure 6. $\delta^{18}O_{seawater}$ was calculated using the method of *Ren et al.* [2003] and by assuming that coral $\delta^{18}O$ variability is only a function of SST or NMAT and $\delta^{18}O_{seawater}$ variability.

Table 2. Correlations Between Annual Averaged Coral $\delta^{18}O$ and Sr/Ca Series at Fiji and Rarotonga

	Overlap, years	r^2 for $\delta^{18}O$	r^2 for Sr/Ca
	Rarotonga		
Cores 2R and 3R	1874–1997	0.54	0.006
Cores 2R and 99	1906–1997	0.50	0.18
Cores 3R and 99	1906–1999	0.60	0.03
	Fiji		
Cores AB and 1F	1780–1997	0.60	0.22

[13] Since it is known that coral skeletal $\delta^{18}O$ is primarily related to water temperature and $\delta^{18}O$ of seawater (linearly related to salinity), we expected to see influences of both parameters in these coral $\delta^{18}O$ series. When analyzed at a resolution of ~10 samples per year we observe a coral $\delta^{18}O$ signal strongly modulated by the 4–5°C annual SST cycle (Figure 2), with interannual variability predominantly driven by SSS (Figure 3). The seasonal $\delta^{18}O$ range of 0.8–1.0‰ is close to that expected for the 4°–5°C annual SST cycle [*Epstein et al.*, 1953; *Dunbar et al.*, 1994; *Wellington et al.*, 1996]. Interannual coral $\delta^{18}O$ variability appears to be largely the result of the 1–1.5 p.s.u. irregular interannual SSS cycle and the advection of the SSS front in response to ENSO (Figures 3b and 3d). Correlations between annual averaged $\delta^{18}O$ and SSS over the interval from 1976 to 2000 have r^2 values of 0.5 for core 1F, 0.6 for core AB and 0.4 for core 2R. Other coral $\delta^{18}O$ results from Fiji support this interpretation [*Le Bec et al.*, 2000]. We conclude that the annual cycle in coral $\delta^{18}O$ at these sites is strongly modulated by SST variability whereas interannual $\delta^{18}O$ variability is strongly modulated by SSS variability. Thus seasonal and interannual variability in these coral $\delta^{18}O$ series is in agreement with observations based on analysis of the instrumental record [*Gouriou and Delcroix*, 2002].

[14] The multicoral $\delta^{18}O$ results also demonstrate the reproducibility of the interannual, decadal, and secular trend components of $\delta^{18}O$ variability at both sites within the measurement error and chronological uncertainties (see Figures 2, 3, and 4 and Table 2). We argue here that the reproducibility of these lower frequency modes of variability in coral $\delta^{18}O$ time series generated at each site from colonies of the same species that are proximal to each other but of different ages, supports a locally consistent, environmental origin for these modes of $\delta^{18}O$ variability.

[15] To further evaluate SST versus $\delta^{18}O_{seawater}$ influences on coral $\delta^{18}O$ at both sites we calculated annual averaged $\delta^{18}O_{seawater}$ from 1870 to 1997 using average annual coral $\delta^{18}O$ and both NMAT (from Hadley Centre Night Marine Air Temperature data set:HadMAT1) and SST anomaly (SSTa; 1° latitude by 1° longitude; from the Hadley Centre Sea Ice and Sea Surface Temperature database: HaDISST1) from each site (Figure 4) [*Rayner et al.*, 2003]. We utilized the technique of *Ren et al.* [2003] and assumed that the coral $\delta^{18}O$ variability is only a function of SST or NMAT and $\delta^{18}O_{seawater}$. We also assumed that the relationship between coral $\delta^{18}O$ and SST was −0.21‰/°C [*Epstein et al.*, 1953; *Dunbar et al.*, 1994; *Wellington et al.*, 1996]. We note that this relationship is generally consistent between different types of biogenic carbonates and between annual-averaged coral calibrations and seasonal-scale calibrations. *Epstein et al.* [1953] calculated a $\delta^{18}O_{carbonate}$–SST relationship of −0.22‰/°C based on analysis of bulk samples from abalone, oysters, other mollusks and calcareous worms grown at known temperatures and corrected for $\delta^{18}O_{seawater}$. *Dunbar et al.* [1994] found a slope of between −0.18 and −0.23‰/°C for annual-average coral $\delta^{18}O$ and annual SST. Seasonal-scale calibrations in corals have found $\delta^{18}O_{carbonate}$–SST relationships generally between ~−0.17 and −0.23‰/°C [e.g., *Wellington et al.*, 1996]. To calculate past changes in $\delta^{18}O_{seawater}$ we have assumed a modern $\delta^{18}O_{seawater}$ of 0.0‰ based on limited (n = 6) regional $\delta^{18}O_{seawater}$ data in this region of the Southwest Pacific (G. A. Schmidt et al., Global seawater oxygen-18 database, 1999; available at http://data.giss.nasa.gov/o18data/) and calculated relative changes back to 1870 (Figure 4). The results show the limited temperature effect on coral $\delta^{18}O$ at both sites on interannual and lower frequencies back to 1870. They also indicate that over the last 130 years annual coral $\delta^{18}O$ variability is predominantly forced by $\delta^{18}O_{seawater}$ variability at each site with r^2 correlations of 0.82 and 0.80 between annual coral $\delta^{18}O$ and $\delta^{18}O_{seawater}$ at Fiji and Rarotonga, respectively (i.e., 82% and 80% of annual averaged coral $\delta^{18}O$ explained by $\delta^{18}O_{seawater}$ variability) (Figure 4).

[16] Correlation of annual average SSS and reconstructed $\delta^{18}O_{seawater}$ over the period from 1976 to 1996 yields the following relationships:

$$\delta^{18}O_{seawater} \text{ from Rarotonga } \delta^{18}O \text{ and NMAT} : 0.23‰/1 \text{ p.s.u. } (r^2 = 0.35)$$

$$\delta^{18}O_{seawater} \text{ from Rarotonga } \delta^{18}O \text{ and SST} : 0.20‰/1 \text{ p.s.u. } (r^2 = 0.32)$$

$$\delta^{18}O_{seawater} \text{ from Fiji } \delta^{18}O \text{ and NMAT} : 0.23‰/1 \text{ p.s.u. } (r^2 = 0.64)$$

$$\delta^{18}O_{seawater} \text{ from Fiji } \delta^{18}O \text{ and SST} : 0.23‰/1 \text{ p.s.u. } (r^2 = 0.57)$$

These observed relationships between SSS and calculated $\delta^{18}O_{seawater}$ are in-line with the 0.27‰/

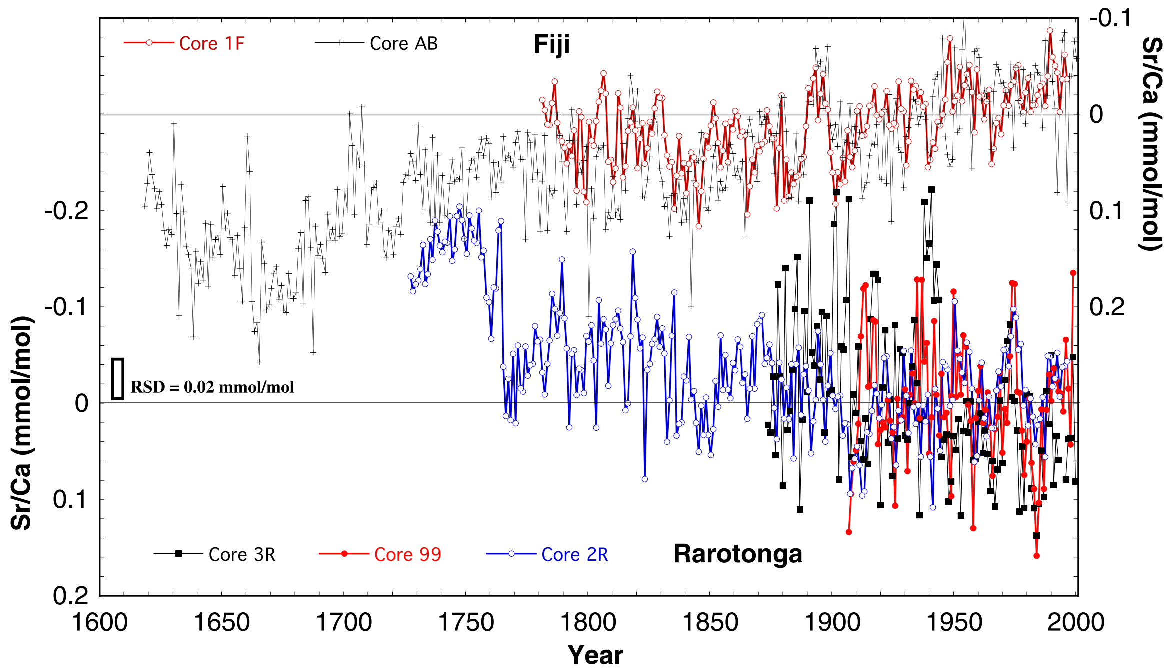

Figure 5. Annually averaged and centered Sr/Ca series from three Rarotonga coral cores and two Fiji cores. Data was centered by removing twentieth century mean. Note that the annually averaged Sr/Ca series from Fiji core 1F and Rarotonga core 2R were generated by binning 1 mm Sr/Ca data, whereas the other core results are from analyses of homogenized splits of samples from 1 year of skeletal growth (see text).

1 p.s.u. derived from instrumental data in the equatorial Pacific [*Fairbanks et al.*, 1997], given the errors associated with calculating $\delta^{18}O_{seawater}$ [*Ren et al.*, 2003]. Using either calibration the interannual $\delta^{18}O_{seawater}$ variability in Figure 4 represents between 1 and 1.5 p.s.u SSS changes. An interannual amplitude of this magnitude is in agreement with instrumental results [*Gouriou and Delcroix*, 2002]. Thus at both seasonal and lower frequencies our Fiji and Rarotonga coral $\delta^{18}O$ series appear to have recorded actual regional oceanographic conditions.

3.2. Strontium/Calcium

[17] Figure 5 compares annually averaged and centered Sr/Ca from all five cores (2 from Fiji and 3 from Rarotonga). As with our $\delta^{18}O$ results, we also found Sr/Ca offsets between cores at each site that cannot be environmental in origin. The Fiji Sr/Ca results from each core have an average offset of 0.14 mmol/mol (±0.05). At Rarotonga the Sr/Ca offset ranges from 0.08 to 0.1 mmol/mol (±0.11). To account for the offset, in Figure 5 we have centered the Sr/Ca results by removing the twentieth century average from each Sr/Ca series.

[18] We have previously reported that coral Sr/Ca at Fiji and Rarotonga analyzed at 1mm increments is highly correlated with SST on monthly timescales and appears to be modulated on decadal timescales by changes in SST [*Linsley et al.*, 2000, 2004]. However, our new annually averaged Sr/Ca results from two additional cores at Rarotonga and one additional core at Fiji cast some uncertainty on the relationship between decadal and secular changes in coral Sr/Ca and SST at these sites (see Table 2). The highest correlation occurs between the 2 Fiji Sr/Ca series which correlate with an r^2 = 0.22. Several interdecadal Sr/Ca changes and the Sr/Ca secular trend replicate well in the two Fiji Sr/Ca series, but other interdecadal changes do not replicate (Figure 5). At Rarotonga the Sr/Ca records from the three coral colonies have even lower correlation coefficients resulting from the obvious lack of agreement before 1950. The low correlations between these Fiji and Rarotonga Sr/Ca series could in part be due to the low signal-to-noise ratio given the measurement error (RSD = 0.02 mmol/mol for centered Sr/Ca or 0.15% RSD) and the small amplitude of the SST variability. However the low correlations also suggest that unknown complicating factors may have influenced coral Sr/Ca in some sections of core. Thin-section analysis identified some secondary aragonite in the bottom 12 cm of the longest Rarotonga core and we have not displayed data

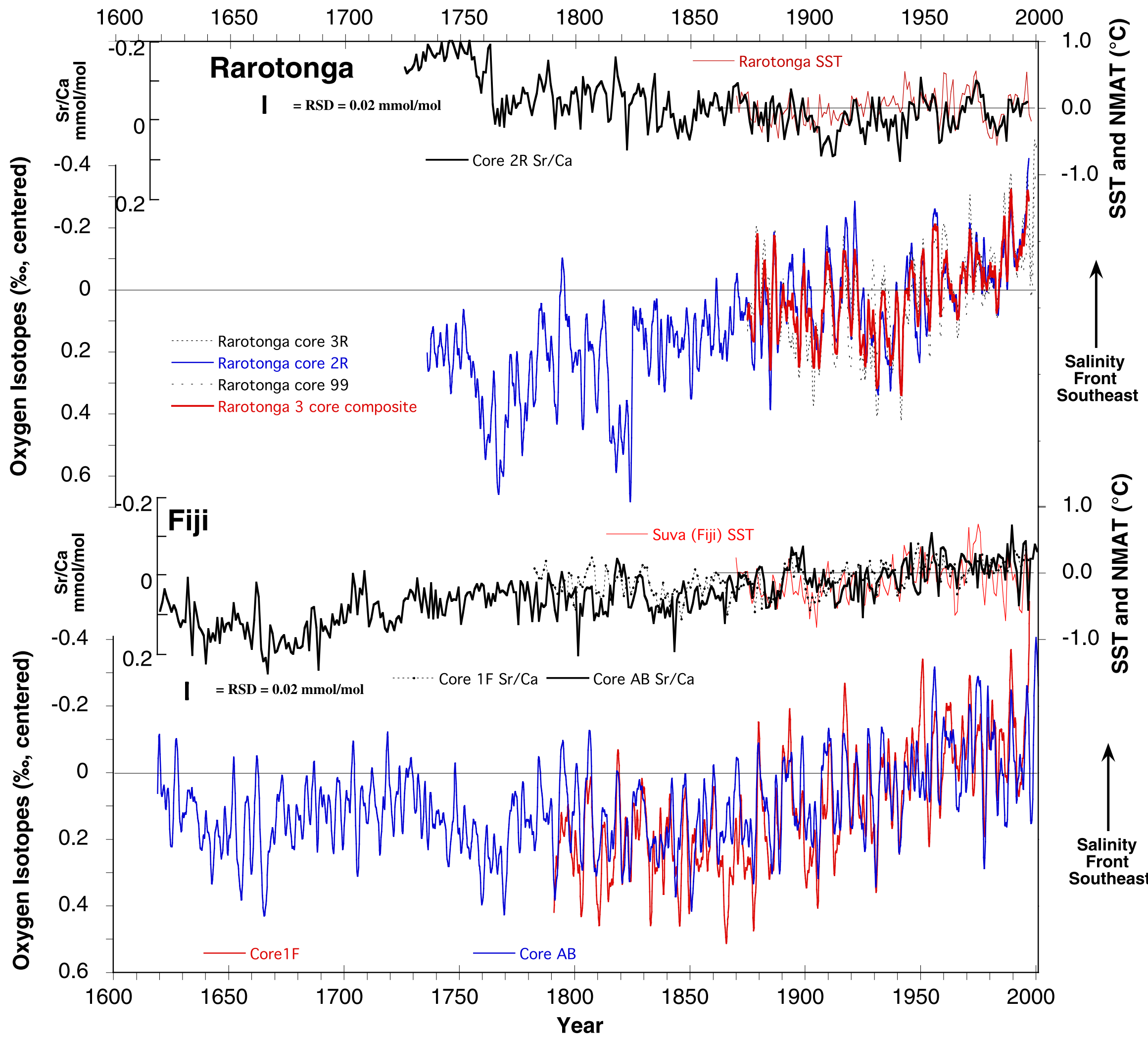

Figure 6. Comparison of coral $\delta^{18}O$ and Sr/Ca from Fiji and Rarotonga coral cores to annually averaged instrumental sea surface temperature anomaly (SST) from both regions. All $\delta^{18}O$ data have been band-pass-filtered at 2 years to remove the annual cycle and centered by removing the twentieth century average. The temperature data and coral $\delta^{18}O$ data have been scaled assuming 0.21‰ $\delta^{18}O$ = 1°C [*Epstein et al.*, 1953; *Wellington et al.*, 1996]. The Sr/Ca data are annual averages and have been scaled where 0.2 mmol/mol = 1°C, as discussed in text. (top) Rarotonga cores: 2R (1997–1726), 3R (2001–1874), 99 (1999–1906). (bottom) Fiji cores 1F (1997–1780) and AB (2001–1619). Note the overall reproducibility of the $\delta^{18}O$ results from different-aged coral colonies with the exception of several intervals (e.g., 1949–1950 in Fiji). RSD is error on Sr/Ca analyses.

from this interval. However, thin section analysis of several other intervals that were checked in this core and in both Fiji cores were found to contain no significant accumulations of secondary aragonite.

[19] Another puzzling aspect of our Sr/Ca results is that the slope of the relationship between monthly SST and near-monthly Sr/Ca over the interval from 1972 to 1997 results in unrealistically large SST secular trends when applied to the entire Sr/Ca series from both Fiji and Rarotonga. Over the interval since 1870 the Sr/Ca results from both sites would display variability more than twice as large as variability in SST and NMAT based on the 0.055 mmol/mol per °C calibration derived from the last 30 years of data. For another example, the Fiji Sr/Ca results for the 20 year period centered on 1660 would indicate a 4°–5°C cooling relative to the late twentieth century using this calibration [*Linsley et al.*, 2004]. In this section of core the Fiji $\delta^{18}O$ results show only a 0.3 to 0.4‰ increase relative to the late twentieth century (Figure 6). This discrepancy between Sr/Ca and $\delta^{18}O$ would indicate an unrealistically large ~2 p.s.u decrease in mean surface salinity relative to the twentieth century if Sr/Ca is only a function of SST and the

current calibration is correct. To get the secular trend and interdecadal Sr/Ca amplitude at Fiji to fit the amplitude of instrumental SST and NMAT variability from 1870 to 1997 requires a scaling of ~0.2 mmol/mol per 1°C (see Figure 6). This scaling also results in a more realistic amplitude of estimated SST changes before 1870.

[20] At Rarotonga, if the Sr/Ca-SST relationship of 0.063 mmol/mol per °C (based on the 1972–1997 calibration) [*Linsley et al.*, 2000] is applied to the Sr/Ca data prior to 1760 this would indicate that mean SST was an unrealistic ~3°C higher before 1760 (near 30°C) (Figures 5 and 6). Scaling of Rarotonga Sr/Ca by the 0.2 mmol/mol per °C results in a better fit to the SST and NMAT data for Rarotonga but assuming Sr/Ca variability is all due to temperature change still indicates that it was 1°C warmer prior to 1760 (see Figure 6). It is noteworthy that this anomalous Sr/Ca interval does not replicate in our new Fiji Sr/Ca results (Figure 5). Although the sites are separated by 20° of longitude, a temperature change of this magnitude would have affected a large section of the South Pacific Ocean and we would have expected it to be recorded by cores at both sites with even higher SST at Fiji based on the current latitudinal and longitudinal SST gradients in relationship to the WPWP. The fact that this anomalous interval (1726–1760 C.E.) in the Rarotonga Sr/Ca record does not replicate at Fiji is further evidence that other factors are influencing coral Sr/Ca at Rarotonga and possibly both sites. However, we believe that the secular trend in the Fiji Sr/Ca results is more realistic and in agreement with the climatology of the region.

[21] The apparent amplification of inferred SST changes from coral Sr/Ca prior to the calibration period was previously discussed with respect to interdecadal variability in Sr/Ca from both Fiji and Rarotonga by *Linsley et al.* [2004] and may suggest that at these sites the Sr/Ca-SST relationship has not been constant over the last 380 years. A similar amplification also appears in the results of a digenetically altered coral from Ningaloo reef in western Australia reported by *Mueller et al.* [2001]. More work is needed with coral Sr/Ca at these sites to understand the significance of the Sr/Ca trends and develop techniques to sort out meaningful paleo-environmental information.

4. Discussion

[22] Comparison of our coral $\delta^{18}O$ results with nighttime marine air temperature anomaly (NMAT, from Hadley Centre Night Marine Air Temperature data set:HadMAT1) and SST anomaly (SSTa; 2° latitude by 2° longitude; from the Hadley Centre Sea Ice and Sea Surface Temperature database: HaDISST1), clearly demonstrates that the multidecadal and secular trend components of $\delta^{18}O$ variability at both sites are significantly larger than what can be explained by temperature change alone (Figures 4 and 6). We have previously demonstrated that at least since the late 1800s the multidecadal signal in Fiji and Rarotonga $\delta^{18}O$ is related to the phase of the IPO [*Linsley et al.*, 2004]. During the most recent negative phase of the IPO (late 1940s to mid 1970s) the SPCZ shifted to the southwest, and SST and precipitation increased in this region of the South Pacific [*Salinger and Mullan*, 1999; *Salinger et al.*, 2001]. The network of coral $\delta^{18}O$ records at Fiji, Rarotonga, New Caledonia [*Quinn et al.*, 1998] and Moorea [*Boiseau et al.*, 1999] all show an expected decrease during this time (Figure 7). With the shift to the positive phase of the IPO in the mid 1970s, the SPCZ shifted to the northeast, resulting in lower SST, reduced precipitation, presumably higher SSS, and increased coral $\delta^{18}O$. Another abrupt shift in Rarotonga $\delta^{18}O$ observed in the mid 1940s at the end of an IPO positive phase, is not well defined in Fiji $\delta^{18}O$, but is clearly recorded in the New Caledonia [*Quinn et al.*, 1998], Moorea [*Boiseau et al.*, 1999] and Great Barrier Reef [*Hendy et al.*, 2002] coral $\delta^{18}O$ records.

[23] Prior to 1900 the regional coral $\delta^{18}O$ results demonstrate that a large multidecadal $\delta^{18}O$ change in the mid to late 1700s is observed at both Fiji and Rarotonga. The 0.4‰ increase in Rarotonga $\delta^{18}O$ from 1814 to 1823, although not observed in Fiji, is present in the New Caledonia $\delta^{18}O$ (see Figures 6 and 7). At Fiji and Rarotonga the magnitude of and $\delta^{18}O$ secular trend toward more negative $\delta^{18}O$ is −0.3 and −0.4‰, respectively. This secular-scale mode is the largest component of variance in each $\delta^{18}O$ series. Using a conservative SSS-$\delta^{18}O_{seawater}$ slope of 0.3‰ per 1 p.s.u., if the entire secular trend was due to salinity change [*Fairbanks et al.*, 1997; *Morimoto et al.*, 2002; *Kilbourne et al.*, 2004], this secular trend would equate to a 1 to 1.2 p.s.u. decrease since the mid 1800s. Taking into account the ~0.25° to 0.5°C of warming based on the SST and NMAT data since 1870 reduces the inferred magnitude of the SSS change to approximately 0.7–0.8 p.s.u. On the basis of compiled SSS data for the region [*Gouriou and Delcroix*, 2002; *Levitus et al.*, 1994], a change of SSS of this magnitude would require an eastward shift in the

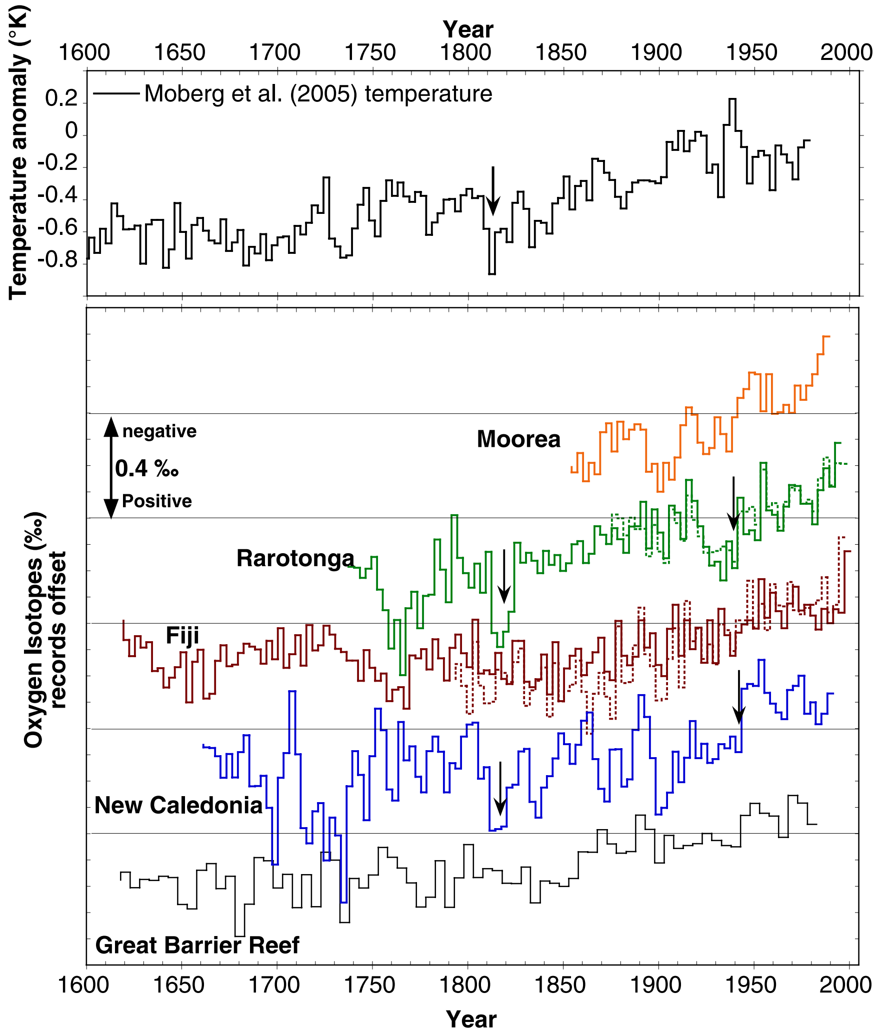

Figure 7. (top) Three-year binned average northern hemisphere temperature reconstruction [*Moberg et al.*, 2005]. (bottom) Comparison of our Fiji and Rarotonga δ^{18}O results to coral δ^{18}O series from New Caledonia [*Quinn et al.*, 1998], Moorea [*Boiseau et al.*, 1999], and the Great Barrier Reef [*Hendy et al.*, 2002]. For Rarotonga, Fiji, New Caledonia, and Moorea, coral δ^{18}O has been binned into 3-year averages. The Great Barrier Reef δ^{18}O results are the original 5-year averages. Data have been offset. Arrows indicate two intervals where inferred SSS decreases are seen at Rarotonga and New Caledonia but not at Fiji. The earlier interval may also correspond to the time following the Tambora volcanic eruption in 1815 A.D. when global temperatures declined.

decadal-mean position of the S. Pacific salinity front by ~20 degrees of longitude since the mid 1800s. A shift of this magnitude is 50–75% of the 1.0 to 1.5 p.s.u. interannual ENSO-related SSS changes of the last 25 years [*Gouriou and Delcroix*, 2002], indicating that these inferences are not unrealistic.

[24] The above interpretations are based solely on the instrumental SST data and coral δ^{18}O results from each site. We note below that adding the Sr/Ca results into this discussion leads to the same interpretation. During intervals of pronounced low-frequency coral δ^{18}O change at Fiji (e.g., late 1700s to early 1900s), only minimal variability is observed in Fiji coral Sr/Ca (Figure 6). If the secular trend component in our Fiji annually averaged Sr/Ca results is interpreted to reflect SST variability (with the above noted uncertainties), the results indicate a long-term warming trend beginning in the mid 1600s with smaller amplitude and non-synchronous decadal-scale variability when compared to Fiji coral δ^{18}O. Regardless of the uncertainty in the stability of the slope of the Sr/Ca-SST relationship over time, for the focus of this discussion the most important observation is that the timing of the secular trend and decadal-scale changes in Fiji Sr/Ca and δ^{18}O generally do not agree. In addition, the amplitude of the coral δ^{18}O secular trend and multidecadal changes at Fiji are 2

to 2.5 times larger than those observed in Fiji Sr/Ca as scaled in Figure 6. Thus although there remains some uncertainty we suggest that the secular trend in the Fiji Sr/Ca results further supports our conclusion that at these sites the $\delta^{18}O$ secular-scale and interdecadal-scale variability are both predominantly attributable to SSS changes over the last 380 years. If the Rarotonga Sr/Ca secular trend is also due to temperature (which we doubt) then this would only strengthen our observation that the Rarotonga $\delta^{18}O$ trend since the mid 1700s is also predominantly attributable to SSS changes.

[25] On the basis of all the available evidence discussed above we interpret the regional $\delta^{18}O$ results to indicate that the synchronous decadal $\delta^{18}O$ changes in the mid to late 1700s and long-term trends observed in Fiji, Rarotonga and Moorea coral $\delta^{18}O$ series are the result of SSS variations and E-W movements of the SSS front currently located near 175°W. The other decadal-scale $\delta^{18}O$ changes recorded at Rarotonga (21.5°S) and/or New Caledonia (22°S), but not at Fiji (17°S), are interpreted to reflect more N-S movements of the meridional SSS gradient that runs E-W along the southern boundary of the SPCZ near 20°S. On the eastern edge of the SPCZ the coral $\delta^{18}O$ results indicate that the decadal mean position of the salinity front has shifted through 10° to 20° of longitude three times since 1619 C.E. The earlier two episodes of eastward propagation lasted from ~1650 to ~1720, and from ~1770 until ~1800. The most recent and largest shift began in the mid 1800s as the salinity front progressively moved eastward until the 1950s, where it was relatively stable for 30 years, before continuing its eastward shift in the mid 1980s. The last shift since 1980 has already been noted in instrumental records [*Trenberth and Hoar*, 1997; *Salinger et al.*, 2001]. The Fiji and Rarotonga $\delta^{18}O$ results indicate that the salinity front and SPCZ currently occupy their most easterly positions in the last 380 years.

[26] Coral $\delta^{18}O$ and trace metal results from the Great Barrier Reef suggest that SST was higher in the Australian coastal region during the early 1800s [*Hendy et al.*, 2002]. These corals also record a similar decrease in $\delta^{18}O$ beginning after 1870 (Figure 6) [*Hendy et al.*, 2002]. Their proposed scenario to explain this observed coral $\delta^{18}O$ decrease draws on other reports of weakened trade winds after the mid 1800s end of the Little Ice Age [*Thompson et al.*, 1986] that led to a reduction in the SEC and concomitant changes in the western boundary current off NE Australia.

[27] An alternative but compatible scenario is that the decadal-mean position of the SPCZ and its bounding SSS front have progressively expanded both eastward and southward since the mid 1800s, albeit at different rates at different locations and with interruptions from decadal-scale variations in local SSS probably due to ocean current dynamics. The unprecedented eastward shift in the SPCZ that began in the mid 1800s and appears to be continuing has shifted the SPCZ out of the longitudinal range of SPCZ variability prior to the mid 1800s based on our Fiji and Rarotonga coral records. On the SPCZ's southern edge, both the Great Barrier Reef and New Caledonia coral $\delta^{18}O$ results indicate that the trend toward lower SSS has also attained unprecedented levels in the twentieth century given regional temperature trends [*Salinger et al.*, 1995; *Folland et al.*, 2003].

[28] On the basis of the current dynamics of the SSS front and SPCZ [*Gouriou and Delcroix*, 2002], the observed long-term eastward shift in the decadal mean position of the SSS front and SPCZ since the mid 1800s implies a transition to more La Niña-like conditions in the southwest Pacific. This scenario is consistent with the tropical ocean thermostat theory articulated by *Clement et al.* [1996] and *Cane et al.* [1997]. These studies describe a mechanism where adding heat to the tropical atmosphere in a coupled ocean-atmosphere system results in greater oceanic heating in regions with no upwelling and a deep thermocline, like the WPWP. In areas of upwelling, such as the eastern Pacific, cooling via vertical advection offsets or minimizes surface heating. If this theory is correct, under conditions of tropical atmospheric warming, the mean state of the tropical Pacific ocean-atmosphere system shifts toward a long-term mean La Nina-like pattern with a larger east-west SST gradient. This mechanism is consistent with SST observations [*Cane et al.*, 1997]. Our results showing an eastward movement of the SSS front and SPCZ expansion since the mid 1800s are also consistent with a shift to more La Niña-like conditions, and thus provide additional support for the *Cane et al.* [1997] scenario of the response of the tropical Pacific ocean-atmosphere system to global warming.

[29] There also appears to be a relationship between at least one decadal change in global temperature [*Chenoweth*, 2001; *Moberg et al.*, 2005] and shifts in the SPCZ-SSS front system. We note that synchronous with the decrease in tropical and global surface temperatures associated with the Tambora volcanic eruption in 1815 the coral $\delta^{18}O$

records from Rarotonga and New Caledonia indicate that the SPCZ shifted north and/or contracted for almost a decade (Figure 6). This is consistent with the results of *Mann et al.* [2005], who found an El Niño-like response in the Zebiak-Cane model to large volcanic eruptions and further supports the thermostat mechanism.

[30] Our interpretations of the South Pacific coral $\delta^{18}O$ records is also consistent with the prediction that rising global temperatures will lead to enhanced convection and rainfall in the oceanic sectors of the ITCZ [e.g., *Held and Soden*, 2000; *Chou and Neelin*, 2004; *Kumar et al.*, 2004]. The initiation of the most recent SPCZ southeastward expansion in the mid 1800s began just after Northern Hemisphere temperatures started to increase [*Moberg et al.*, 2005]. Despite wide differences in regional precipitation changes in the tropics, there is a consensus from climate model simulations that the atmosphere warms and becomes moister in response to increases in anthropogenic greenhouse gases [*Held and Soden*, 2000; *Chou and Neelin*, 2004; *Kumar et al.*, 2004]. Although not all instrumental rainfall records from the SPCZ region show trends of increasing precipitation in the twentieth century and are spatially quite variable [*Salinger et al.*, 1995], it is possible that the SPCZ has increased in surface area but not in overall rainfall intensity since the mid 1800s. This would explain both the coral $\delta^{18}O$ results and the variable island precipitation records.

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