

CHANGES OF SUBTROPICAL NORTH PACIFIC RADIOCARBON AND CORRELATION WITH CLIMATE VARIABILITY

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ABSTRACT. We show that high-precision radiocarbon ($\Delta^{14}\text{C}$) measurements from annual bands of a Hawaiian surface coral decreased by 7‰ from AD 1893 to 1952. This decrease is coincident with the Suess Effect, which is mostly due to the dilution of natural levels of ^{14}C by ^{14}C -free fossil fuel CO_2 . This decrease is equal to that expected in surface waters of the subtropical gyres, and indicates that the surface waters of the North Pacific were in steady state with respect to long term mixing of CO_2 during the past century. Correlation between $\Delta^{14}\text{C}$ and North Pacific gyre sea surface temperatures indicates that vertical mixing local to Hawaii and the North Pacific gyre as a whole is the likely physical mechanism to result in variable $\Delta^{14}\text{C}$. Prior to 1920, this correlation starts to break down; this may be related to the non-correlation between biennial $\Delta^{14}\text{C}$ values in corals from the southwest Pacific and El Niño events observed during this period as well.

INTRODUCTION

Knowledge of the various mixing modes in the ocean, and understanding of how they have changed in the past as a function of climate are key to understanding future climate. Radiocarbon (^{14}C) in DIC (dissolved inorganic carbon) is one of the most useful transient tracers for understanding ocean circulation on a range of time scales. ^{14}C is produced naturally in the upper atmosphere and as a result of thermonuclear bomb testing in the late 1950s and early 1960s. Century-long ^{14}C records in corals provide critical tests for ocean circulation models and for hypotheses of the role of the oceans in past climate change. Long time series ^{14}C records (>100 yr) are available from only three locations in the Atlantic (Florida, Bermuda, Belize) and one in the Pacific (Great Barrier Reef). From these and shorter records, researchers have uncovered evidence of past vertical changes in surface-subsurface mixing and horizontal current shifts. ^{14}C provides important constraints for modeling various chemical cycles in the ocean (Broecker et al. 1978, 1980; Druffel 1989; Toggweiler et al. 1989), especially when combined with records of stable isotopes (Gagan et al. 2000), Sr/Ca (Linsley et al. 2000), and trace metals (Linn et al. 1990; Shen et al. 1992) in coral bands. We report natural levels of ^{14}C , Sr/Ca ratios and stable oxygen and carbon isotope ratios in a Hawaiian coral that grew prior to 1953. Guilderson et al. (2001) report $\Delta^{14}\text{C}$ of seasonal samples from the same coral core for the period 1946–1992.

Annual density bands are present in the calcium carbonate (aragonite) skeletons of many species of hermatypic, massive corals (for example: Hudson et al. 1976; Knutson et al. 1972). Upon X-radiographic analysis of a thin slab of coral (7–10 mm thick), alternating light and dark bands are revealed; one band pair represents one year's growth at most locations. Because aragonite is saturated in the surface ocean, it does not dissolve. There is also no evidence that aragonite exchanges with carbonate from the seawater. Thus, unaltered coralline aragonite retains a permanent record of various isotopic and chemical signatures that existed in past ocean waters.

The ^{18}O values in banded corals correlate inversely with sea surface temperature (SST) (Dunbar and Wellington, 1981; McConnaughey 1989; Weber and Woodhead 1972) and directly with salinity (or ^{18}O) of seawater (Cole and Fairbanks 1990; Swart and Coleman 1980). Due to kinetic effects, hermatypic corals precipitate calcium carbonate depleted in the heavy isotopes, ^{18}O and ^{13}C , relative to that expected for aragonite-sea water equilibrium. Nonetheless, the displacement from equilib-

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rium is constant and the annual range in ^{18}O (corrected for changes in $^{18}\text{O}_{\text{water}}$) is equal to the annual range predicted by equilibrium ^{18}O -temperature fractionation (Dunbar and Wellington 1981; Fairbanks and Dodge 1979). Aragonite ^{13}C values are also lower than those expected from equilibrium precipitation. The three environmental factors that are believed to control ^{13}C in coral skeletons are the DIC ^{13}C value, light and heterotrophy (Grottoli and Wellington 1998; McConaughey 1989; Weber 1974; Weil et al. 1981).

Climatic Shifts – ENSO and the Pacific Decadal Oscillation

The ocean is not at steady state with respect to SST or mixing on interannual timescales (Davis 1976; Namias 1969; Trenberth and Hurrell 1994). Interannual variability of climate known as ENSO (El Niño/Southern Oscillation) is caused by interaction of the tropical ocean and atmosphere (Bjerknes 1969). There are also decadal shifts of climate centered in the North and tropical Pacific (Mantua et al. 1997), known as the Pacific Decadal Oscillation (PDO). For example, the Pacific climatic shift of 1976 was a well documented, decadal warming of tropical waters (Gu and Philander 1995). Four hypotheses have been posed to explain these decadal shifts: 1) coupled instability that involves ocean-atmosphere interactions in the subtropics (Latif and Barnett 1994), 2) exchanges of subtropical and tropical subsurface waters that cause changes to the thermal structure with subsequent air-sea interactions in the tropics (Gu and Philander 1997), 3) stochastic atmospheric forcing that causes longer term oceanic responses (Hasselmann 1976; Zorita and Frankignoul 1997), and 4) global warming that could contribute to more frequent and more severe El Niño events and fewer La Niña (cooling) events since the onset of the 1976 climate shift (Trenberth and Hoar 1997).

Lysne et al. (1997) use an ocean general circulation model to show that the extratropical regions of the North Pacific ($>20^\circ\text{N}$) generate decadal variability of SST on the equator, that precedes the affect at the tropics by 4–5 years. Indices of heat storage anomalies to 400 m from observations made between 1968 and 1984 (White 1995) show that tropical anomalies lag those in the North Pacific by 4–5 years (Lysne et al. 1997). Whether the tropical Pacific index lagged the subtropical North Pacific index prior to 1968 has not been established because of incomplete data records at both locations. A recent study of a Rarotonga (21°S , 160°W) coral concluded that several large decadal changes of Sr/Ca ratios and the PDO are in phase, suggesting that the origin of the decadal variability is in the tropics and transmitted rapidly via the atmosphere (Linsley et al. 2000).

The effect of decadal mixing changes in the upper ocean on climate is not known because there are too few observations. Long-term records of water mass changes are needed from strategic sites in the subtropical and tropical Pacific in order to determine the long term origin of climatic events in the tropical and subtropical Pacific. Here we present $\Delta^{14}\text{C}$, Sr/Ca, ^{18}O and ^{13}C records in annual and seasonal coral bands in a Hawaiian coral.

Oceanographic Setting of Hawaii and Collection of the Coral

Our coral core was collected from the Kona (western, dry) coast of the Big Island of Hawaii ($19^\circ 31'\text{N}$, $155^\circ 58'\text{W}$) in April 1992. The site was 2.5 km south of Keauhou Bay, 50 m from shore, at 8 m water depth, and exposed to open ocean waters (Cathcart 1996). The 10-cm diameter core was taken from a head of *Porites lobata* that was 4 m high and 6.8 m in diameter. The average range of SST at 19°N , 159°W is 2.5–3 °C over the past 40 years from COADS surface marine observations (http://ferret.wrc.noaa.gov/fbin/climate_server). Seasonal oxygen isotope values for this coral ranged from about -4.2‰ to -4.6‰ , lower than expected from the relationship between ^{18}O and SST ($0.2\text{‰}/1^\circ\text{C}$ [Druffel 1985]). Therefore, salinity changes likely occurred at our site on an annual basis (Cathcart 1996). The surface circulation is toward the west-southwest direction (North Equatorial

Current [NEC]), and is controlled by the northeasterly winds. Surface waters at this site are mainly low in nutrients because of its mid-gyre location where most of the water is being downwelled. There are eddies caused by the interaction between the NEC and the southern tip of the Big Island of Hawaii (Ka Lae).

METHODS

Methods used to clean, X-ray, map, and section the annual growth bands are described elsewhere (Druffel and Griffin 1993; Griffin and Druffel 1985). This coral was analyzed for monthly Sr/Ca (Schrag 1999), ^{18}O and ^{13}C for bands 1915–1992 at Harvard; the uncertainties for each analysis were 0.15%, 0.07‰, and 0.05‰, respectively. These data, the density band patterns and $\Delta^{14}\text{C}$ measured in bands from selected post-bomb years of late 1950s and early 1960s (Table 1) were used to construct an age model back to 1915. High-density bands that accreted from about October to December of each year were used to assign ages to bands <1915 because there were no Sr/Ca or stable isotope data for this time period. We estimate that the error of the age of the oldest band (1893) was at most ± 2 years.

Annual composites for samples older than 1953 were mixed from the monthly samples and prepared for ^{14}C analysis at UCI. Coral samples were converted to CO_2 , then to graphite on Co catalyst with H_2 gas as the reductant (Vogel et al. 1987).

All of the samples were analyzed for $\Delta^{14}\text{C}$ using accelerator mass spectrometry (AMS) techniques at the Center for AMS Research at Lawrence Livermore National Laboratory. The $\Delta^{14}\text{C}$ measurements were of high precision, with 1 σ total uncertainty (counting statistics and laboratory reproducibility) of 2.5–3.5‰. ^{14}C measurements are reported as $\Delta^{14}\text{C}$ values (for age-corrected geochemical samples) according to standard techniques (Stuiver and Polach 1977). The ^{13}C measurements were made on all CO_2 samples prior to graphitization and used to correct the $\Delta^{14}\text{C}$ results.

Duplicate sets and splits of AMS analyses were performed to determine total uncertainty of the $\Delta^{14}\text{C}$ measurements. Seven duplicate coral samples (separate acidifications of subsamples of the crushed coral) revealed an average $\Delta^{14}\text{C}$ difference and standard deviation of $1.9 \pm 0.8\%$, respectively. Splits (same CO_2 gas produced from a single acidification) revealed an average $\Delta^{14}\text{C}$ difference of $2.2 \pm 1.7\%$ ($n=4$). AMS analyses of 12 replicates of our modern coral standard (CSTD) conducted over a 15-month period revealed an average $\Delta^{14}\text{C}$ of $-58.4 \pm 2.5\%$, which agreed with our average of $-56.3 \pm 2.4\%$ for 18 results from conventional counting at WHOI and UCI from 1992 to 1997.

Separate seasonal coral samples (8–11 samples per year) were drilled for the time period 1920.8 to 1923.0 using a diamond-tipped drill bit mounted on a Dremel tool. Seasonal $\Delta^{14}\text{C}$ analyses were measured on these 21 samples as described above. All ^{13}C and ^{18}O measurements reported for these samples were performed according to standard techniques by Ms Dorinda Ostermann in the laboratory of Dr William Curry at Woods Hole Oceanographic Institution (crushed dry, individual acid baths for each sample) (Druffel and Griffin 1999). These stable isotope results had total uncertainties (1 σ) of $\pm 0.05\%$ for ^{13}C and $\pm 0.03\%$ for ^{18}O .

RESULTS

The ^{18}O trend shows general agreement with SST and Sr/Ca for most of the record (Figure 1a). The seasonal signal (0.2–0.4‰) is smaller in amplitude than the expected range if SST (2.5–3.0 °C) was the only control (0.5–0.6‰), likely due to changes in salinity or ^{18}O of seawater at this site. There is interannual variability prior to 1940 in both the Sr/Ca and ^{18}O records; the ^{18}O variability

(0.4‰, 2 °C) is about equal to that expected from the change in Sr/Ca ratios (0.00010, 2 °C; Schrag 1999). The ^{13}C record displays a seasonal signal even prior to 1940, which made it helpful for assigning ages to the earlier bands (Figure 1b). The large interannual variability evident prior to 1930 is 3‰ and may reflect changes in DIC ^{13}C , light and/or heterotrophy (McConnaughey 1989).

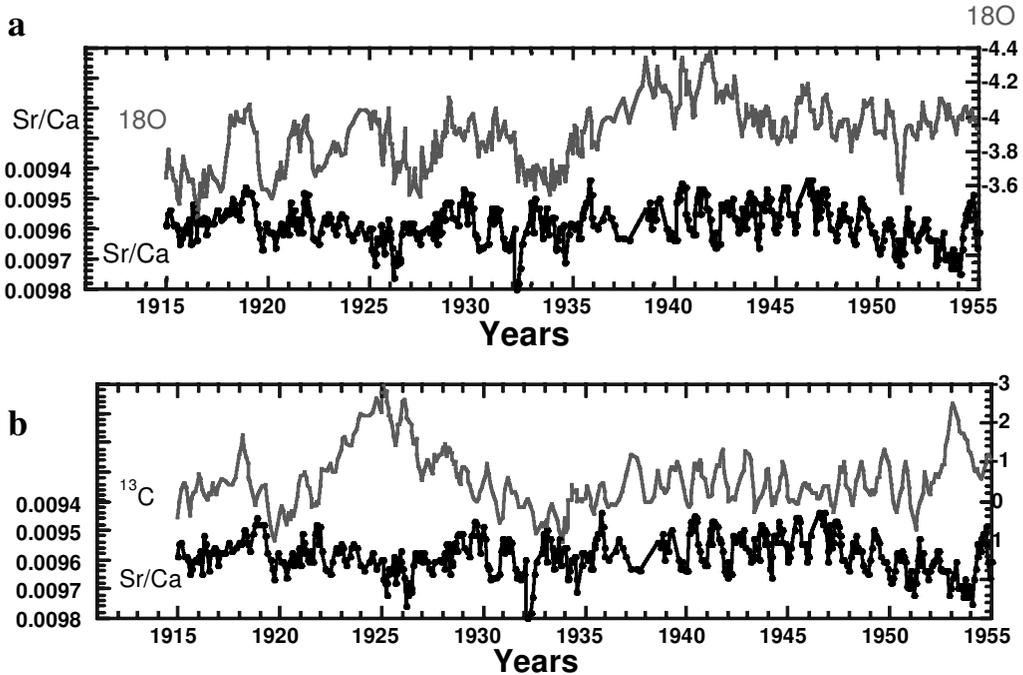


Figure 1 High-resolution (12 samples/year) records: a) Sr/Ca (mole ratio) and ^{18}O (‰), and b) Sr/Ca (mole ratio) and ^{13}C for the Hawaiian coral from the west coast of the Big Island of Hawaii.

The average of 62 annual $\Delta^{14}\text{C}$ measurements (duplicate analyses and splits were averaged to a single point) from the Hawaiian coral averaged $-47.9 \pm 3.7\text{‰}$. The range of values was 14.9‰, from -55.0‰ to -40.1‰ (Table 1). The range and average of $\Delta^{14}\text{C}$ values for the first half of the record (<1921 , 14.9‰; -45.6 ± 3.6 n=29) were larger than for the second half (9.6‰; -49.9 ± 2.5 n=33). There was also a significant decrease of 7‰ from 1893 to 1952 as illustrated by a least squares fit of the $\Delta^{14}\text{C}$ measurements (line shown in Figure 2).

$\Delta^{14}\text{C}$ measurements were made for 20 seasonal samples that were drilled alongside the original transect for the time period 1920.8–1923.0 (8–11 samples/year). The annual measurements for 1921.19 and 1922.19 (-46.6‰ , -48.1‰ , see Table 1) agreed within 1 sigma with the averages of the seasonal samples for these years ($-45.5 \pm 2.3\text{‰}$ n=9, $-47.4 \pm 2.6\text{‰}$ n=8, respectively). The seasonal $\Delta^{14}\text{C}$ data between 1921.8 and 1922.2 have a slightly lower average ($-49.0 \pm 1.2\text{‰}$ n=5) than those before (average $-45.3 \pm 2.2\text{‰}$ n=10) and after ($-46.0 \pm 1.2\text{‰}$ n=5) this time (Figure 3a). This seasonal variability of $\Delta^{14}\text{C}$ ($\sim 3\text{--}4\text{‰}$) is small compared with some other sites where seasonal changes in mixing, e.g. upwelling, is prevalent (Guilderson and Schrag 1998).

The ^{18}O measurements of the monthly samples display a 0.4‰ range (Figure 3a). The values are lowest between the period 1921.4 and 1921.9. They increase by about 0.3‰ between 1922.0 and

Table 1 High-precision $\Delta^{14}\text{C}$ measurements of annual Hawaiian coral bands

UCID#	Year	$\Delta^{14}\text{C}$	\pm
3987	1893.33	-48.5	2.6
3982	1894.29	-49.4	2.5
3985A	1895.29	-45.6	2.9
3985B	1895.29	-47.1	2.7
3279	1896.21	-46.0	2.8
3284	1897.25	-47.7	3.2
3275	1898.26	-41.9	2.7
3304	1899.09	-43.2	2.6
3282	1900.07	-41.4	2.8
3276	1901.21	-49.6	2.8
3285	1902.06	-47.5	3.1
3283	1902.73	-45.8	2.8
3280	1903.57	-40.1	2.8
3274	1904.33	-42.9	3.3
3281	1904.33	-44.0	2.4
3277	1905.21	-45.6	2.8
3984	1905.21	-43.3	2.5
3273	1906.18	-42.1	2.7
4429	1907.13	-49.2	2.3
4432	1908.32	-46.4	2.1
3308	1909.28	-49.4	2.8
3299	1910.23	-42.5	2.7
3297A	1911.20	-44.7	3.1
3297B	1911.20	-41.0	2.8
3306	1912.71	-44.9	2.7
3298	1913.18	-54.1	2.7
3988	1913.18	-55.8	2.5
3301	1914.15	-49.3	2.7
3197	1915.10	-40.4	2.3
3190	1916.34	-44.5	2.7
3196	1917.46	-42.5	2.6
3270	1918.24	-50.3	2.7
3195	1919.10	-40.3	2.2
3983	1919.10	-43.3	3.0
3278	1920.09	-46.1	2.6
3189	1921.19	-46.6	2.7
3194	1922.19	-48.1	2.6
3186	1922.98	-48.6	2.7
4431	1923.86	-52.2	2.3
3198	1925.26	-45.3	2.6
3193	1926.34	-45.4	2.7
3265	1927.17	-49.4	2.8
3266	1928.17	-51.6	2.7
3188	1929.30	-50.7	2.9
3179	1930.24	-51.0	2.6
3271	1931.23	-54.0	2.7
3182	1932.14	-46.5	2.8
3173	1933.06	-50.8	2.6
3180	1934.22	-50.7	2.1
3181	1934.22	-47.9	2.6
3183	1935.19	-49.1	2.8
3178	1936.17	-50.6	2.6
3994	1936.17	-49.2	2.2
3993	1937.20	-49.5	2.2
4430	1937.50	-54.1	2.3
3992	1938.20	-45.0	2.6
3991	1939.70	-51.6	2.5
3175	1940.12	-52.7	3.7
3184A	1940.12	-50.0	2.7
3184B	1940.12	-53.4	2.6
3174	1941.10	-54.6	2.8
3172	1942.17	-50.4	2.6
3191	1943.15	-48.4	2.3
3986	1943.20	-49.3	2.8
3201	1944.18	-49.1	2.6
3267	1945.24	-51.0	2.7
3200	1946.22	-47.9	2.5
3187	1947.07	-48.9	2.7
3272	1948.05	-48.4	2.7
3199	1949.10	-52.3	2.6
3192	1950.10	-52.1	2.9
3185	1951.10	-50.8	2.4
3309A	1951.90	-53.5	2.8
3309B	1951.90	-53.3	2.7
3168	1958.01	-25.6	2.7
3167	1961.90	11.0	2.8
3170	1964.92	100.1	3.0
3169	1966.04	114.4	3.1
Data in boxes indicate duplicates or splits (A/B)			
of the same samples			

1923.0; the first part of this increase coincides with the period of low $\Delta^{14}\text{C}$ values. The ^{13}C values range from -1.6 to -0.5‰ (Figure 3b) and are lowest during the time period 1921.4 to 1922.1, also coincident in part with the low $\Delta^{14}\text{C}$ period.

The monthly Sr/Ca measurements (from the original transect) averaged 0.0096 during 1920.8–1921.6 and then decreased to 0.0095 from 1921.7 to 1922.0. After that, there was an increase to 0.00965 by 1922.2, indicating cool waters. This trend in the first half of 1922 was coincident with the low $\Delta^{14}\text{C}$ values. Thereafter, Sr/Ca ratios decreased to 0.00956.

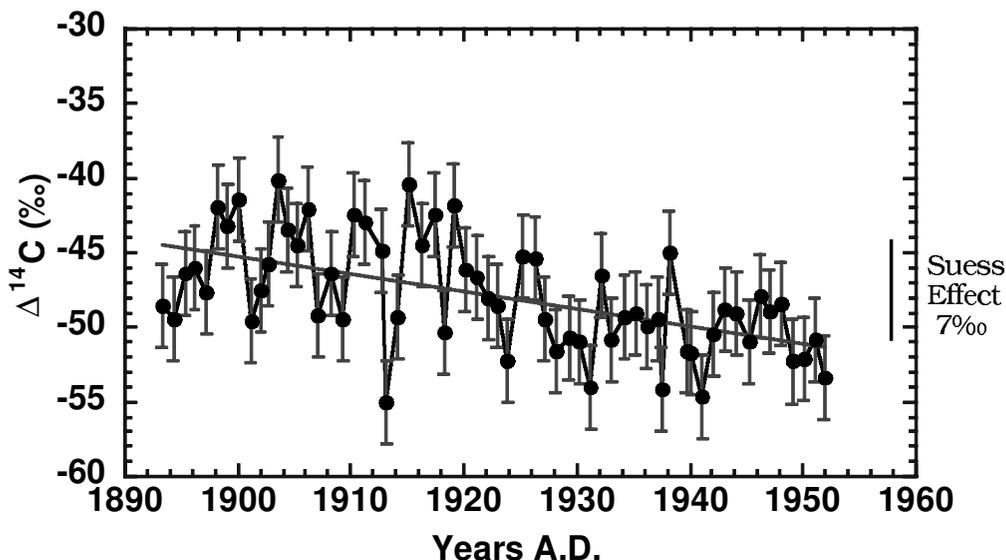


Figure 2 $\Delta^{14}\text{C}$ analyses of annual coral bands from the Hawaiian coral. Note that the Suess Effect is 7‰ from the least squares fit drawn through the points.

DISCUSSION

An average of all Hawaiian seasonal $\Delta^{14}\text{C}$ analyses was $-46.4 \pm 2.3\text{‰}$ for the time period 1920.8–1923.0 (Table 2) and the range of individual values was 7‰ . This range agrees with the seasonal range of 6 to 15‰ observed for $\Delta^{14}\text{C}$ values from a 6-year pre-bomb period (1946–1952) in this coral (Guilderson et al. 2001) and a coral from the subtropical South Pacific (Rarotonga, 21°S , 160°W , $8\text{--}15\text{‰}$; Guilderson et al. 2000). The range of seasonal $\Delta^{14}\text{C}$ in Hawaiian coral is less than that in two other seasonal pre-bomb records: Nauru (0° , 166°E , $10\text{--}20\text{‰}$; Guilderson et al. 1998) and Galapagos (1°S , 91°W , $15\text{--}25\text{‰}$; Guilderson and Schrag 1998). Lower seasonal ranges of $\Delta^{14}\text{C}$ are observed at Hawaii and Rarotonga because of the stable mixed layers at these subtropical locations. Galapagos is influenced by equatorial upwelling, where ^{14}C -depleted waters come to the surface on a seasonal basis.

The low $\Delta^{14}\text{C}$ values during 1921.8–1922.2 coincide with both increasing Sr/Ca ratios and ^{18}O values, but the seasonal signal in $\Delta^{14}\text{C}$ ($3\text{--}4\text{‰}$) is too low to be significant. In 1.5 month resolved data between 1946 and 1952, Guilderson et al. (2001) find that $\Delta^{14}\text{C}$ is low when SSTs are colder and high when SSTs are warmer.

Inter-annual variability is apparent in the annual ^{14}C record, however, with changes ranging from 5 to 15‰ every few years (Figure 2). The question remains as to the possible physical mechanisms

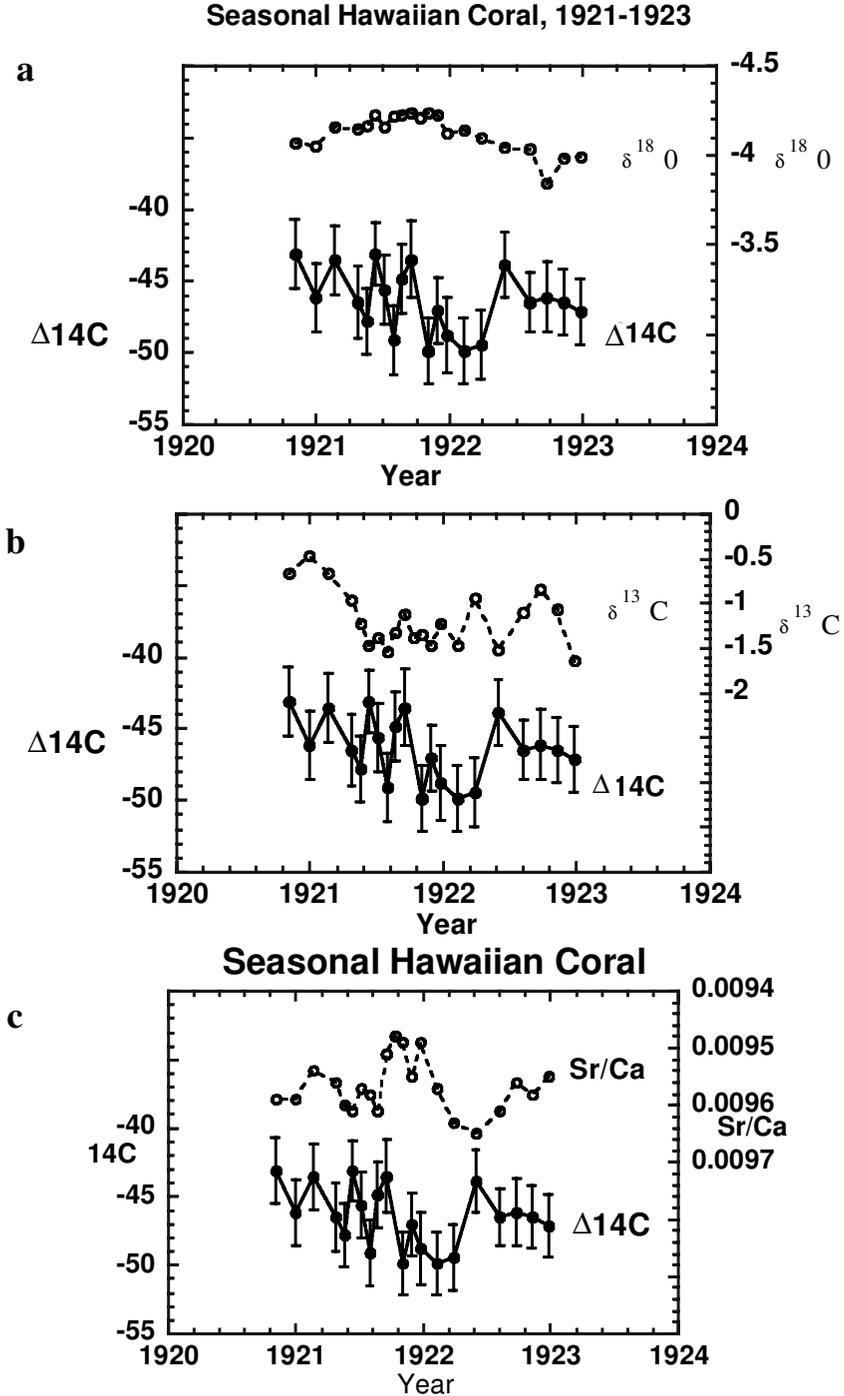


Figure 3 $\Delta^{14}\text{C}$ and a) ^{18}O , b) ^{13}C , and c) Sr/Ca in the high-resolution samples (8–11 per year) for the Hawaiian coral during the period 1920.8–1923.0.

that can result in variable $\Delta^{14}\text{C}$. Two possibilities are: 1) vertical mixing local to Hawaii and/or the North Pacific gyre as a whole, and 2) advected components from the Eastern Equatorial Pacific/Panama Basin or the North Equatorial Current.

First, the general decrease with time in the $\Delta^{14}\text{C}$ record is likely due to the Suess Effect. The absence of a long-term change in Sr/Ca indicates that SST has not changed over the 60 years of coral records (Figure 1a). The Suess Effect in the atmosphere was a 20‰ decrease of $\Delta^{14}\text{C}$ (as measured in tree rings) from the late 1800s to 1955 (Stuiver and Quay 1981) and was caused mainly by the input of ^{14}C -free CO_2 from fossil fuel burning into the atmosphere (Suess 1953). The expected Suess Effect in the surface waters of the ocean was 10‰ in subtropical regions (Stuiver and Quay 1981). This trend is approximately equal to that found in the Hawaiian coral (Figure 2), and indicates that the surface waters of the North Pacific were in steady state with respect to long term mixing of CO_2 during the past century.

Second, it appears that winter SST anomalies in the North Pacific (30–50°N, 180–120°W [Kaplan et al. 1998]) are a reasonable proxy for the annual Hawaiian North Pacific gyre SSTs (Figure 4). Over most of the record, $\Delta^{14}\text{C}$ varies in a manner consistent with vertical mixing; when SSTs are colder (primarily due to winter vertical pumping) $\Delta^{14}\text{C}$ is lower and when SSTs are warmer, $\Delta^{14}\text{C}$ is higher (Figure 4). Prior to 1920, this correlation starts to break down and may be due to vertical gradients in $\Delta^{14}\text{C}$, poor spatial-temporal-coverage in the COADS data or uncertainties in the coral age-model (± 2 years at 1893). Prior to 1920, there is still a 5–6 year “beat” in the SST record that matches the latter portion. If we change our age-model by +1 year, then we can use the same physical mechanism (vertical mixing) to match $\Delta^{14}\text{C}$ with SST. We also note that between 1880 and 1920, $\Delta^{14}\text{C}$ values in an Abraham reef coral were not low during El Niños as they had been both before and after (Druffel and Griffin 1993, 1999). It is possible that these two phenomena are related and indicate a link between North and South Pacific basin climatology.

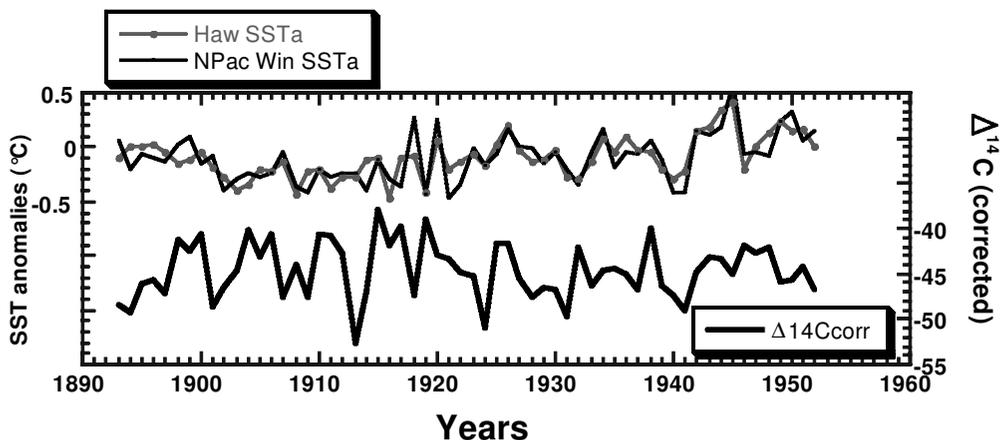


Figure 4 Winter SST anomalies in the North Pacific (30–50°N, 180–120°W), Hawaiian mean annual SST and the annual $\Delta^{14}\text{C}$ record for the Hawaiian coral (with the Suess Effect subtracted out) (from Table 1).

Table 2 Monthly Hawaiian coral data

Sample	UCID#	Year	$\Delta^{14}\text{C}$	\pm	$\pm\text{mos}$	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	Sr/Ca
H14	4091	1922.98	-47.1	2.3	1	-1.65	-3.99	0.00955
H15	4081	1922.85	-46.5	2.3	1	-1.06	-3.99	0.00958
H16	4082	1922.73	-46.1	2.5	1	-0.83	-3.84	0.00956
H17	4083	1922.6	-46.5	2.1	1	-1.11	-4.04	0.00961
H18-19	4084	1922.41	-43.9	2.3	2	-1.53	-4.05	0.00965
H20	4085	1922.23	-49.4	2.4	1	-0.94	-4.10	0.00963
H21	4086	1922.1	-49.9	2.3	1	-1.46	-4.14	0.00957
H22	4087	1921.98	-48.8	2.6	1	-1.23	-4.12	0.00949
H23	4088	1921.91	-47	2.3	1	-1.47	-4.23	0.00955
H24	4089	1921.84	-49.9	2.3	1	-1.34	-4.24	0.00949
H25	4090	1921.78	ND	ND	1	-1.38	-4.21	0.00948
H26	4100	1921.71	-43.5	2.7	1	-1.12	-4.23	0.00951
H27	4098	1921.64	-44.8	2.4	1	-1.32	-4.23	0.00961
H28	4095	1921.58	-49.1	2.4	1	-1.53	-4.22	0.00958
H29	4097	1921.51	-45.6	2.4	1	-1.38	-4.16	0.00957
H30	4099	1921.44	-43.1	2.2	1	-1.46	-4.23	0.00961
H31	4101	1921.38	-47.8	2.3	1	-1.23	-4.16	0.00960
H32-33	4093	1921.31	-46.5	2.5	2	-0.95	-4.15	0.00956
H34-35	4092	1921.14	-43.5	2.4	2	-0.66	-4.16	0.00954
H36-37	4094	1921	-46.1	2.4	2	-0.47	-4.05	0.00959
H38-39	4096	1920.84	-43.1	2.4	2	-0.67	-4.07	0.00959
ND = No Data								

Regarding the vertical mixing scenario we have presented here, there are undoubtedly complications due to advected signatures. Guilderson et al. (2001) demonstrate that $\Delta^{14}\text{C}$ at Hawaii is affected by non-local sources because the plateau of the bomb $\Delta^{14}\text{C}$ peak is short-lived. There is low $\Delta^{14}\text{C}$ water being advected into the area, otherwise there would have been a longer plateau of high values before decreasing in the early-to-mid 1970s (Guilderson et al. 2001).

Spectral analyses (Arand software, Phillip Howell, personal communication, 1999) showed spectral density in the 6-year cycle for the $\Delta^{14}\text{C}$, ^{18}O and Sr/Ca records. Cross spectral analyses (CSA) of the $\Delta^{14}\text{C}$ and ^{18}O records reveal that the 6-year period is coherent to the 80% confidence level, as is the CSA of the $\Delta^{14}\text{C}$ and Sr/Ca records. Interestingly, the CSA of the $\Delta^{14}\text{C}$ and the wind speed records (COADS) at Hawaii reveal the 6-year period is coherent (80% CL). These results imply that the origin of the ^{14}C change at Hawaii is initiated by wind variability (high wind, low $\Delta^{14}\text{C}$), supporting our simple view that vertical mixing controls $\Delta^{14}\text{C}$. There was no coherent period, without an unreasonable lag, in the CSA of the $\Delta^{14}\text{C}$ record and the sea level pressure difference between Tahiti and Darwin (ENSO index).

A picture is emerging whereby change in vertical mixing appears to have been the initiator of annual $\Delta^{14}\text{C}$ change in the subtropical North Pacific. This scenario indicates that winds are driving circulation in the subtropical North Pacific and that $\Delta^{14}\text{C}$ correlates with these climate changes.

Our results do not confirm or deny Lysne et al.'s (1997) observations of a several year lag between warming and cooling periods in the subtropics prior to arrival of these events in the tropics. The subtropics and tropics are linked by wind-driven meridional circulation. Water in the eastern regions of the subtropical oceans are subducted and transported southwest to the equatorial thermocline, where

they are subsequently upwelled to the surface. Distributions of transient tracers such as tritium (Fine et al. 1987) and ^{14}C (Quay et al. 1983) have shown that subtropical-to-tropical transport of water parcels occurs and this mechanism is maintained by the easterly trade winds. Gu and Philander (1997) surmised that interdecadal climate fluctuations are linked between the tropics and subtropics and that continual changes between the tropics and subtropics can be the result of these processes. However, longer records of $\Delta^{14}\text{C}$, Sr/Ca and ^{18}O in corals from the tropical and subtropical Pacific are needed before we can show whether there is a link over long timescales.

CONCLUSIONS

Our $\Delta^{14}\text{C}$ record shows a long term change reflective of the Suess Effect, which indicates that the surface waters of the subtropical North Pacific were in steady state with respect to long term mixing of CO_2 during the past century. It appears that winds are driving circulation in the subtropical North Pacific, which causes variability in SST and $\Delta^{14}\text{C}$. Longer $\Delta^{14}\text{C}$ records are needed for the subtropical and tropical regions to determine if there are correlations between tropical and subtropical climate and whether they have changed in nature or periodicity.

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