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Variability of surface ocean radiocarbon and stable isotopes in the southwestern Pacific

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Abstract. We present high-precision radiocarbon ($\Delta^{14}C$) results and stable isotope ($\delta^{18}O$ and $\delta^{13}C$) records for a coral from Heron Island (23°S, 152°E) and new stable isotope ($\delta^{18}O$ and $\delta^{13}C$) records for annual coral bands from Abraham Reef (22°S, 153°E) in the southern Great Barrier Reef studied earlier [*Druffel and Griffin*, 1993]. These tracers provide unique information on the regional water mass history, and together these data allow us to constrain the variability of circulation in the upper Pacific over the past four centuries. First, we observe decreases in $\delta^{18}O$ of coral from Abraham Reef and Heron Island, indicating an increase in sea surface temperature and/or a decrease in surface salinity since 1850. Second, the small Suess effect value ($\Delta^{14}C$ decrease from 1880 to 1955, due mostly to fossil fuel CO_2) observed previously at Abraham Reef[*Druffel and Griffin*, 1993] is confirmed in the measurements reported here from the Heron Island coral. This value is low compared to those observed in other areas of the ocean [*Druffel*, 1997; *Druffel and Linick*, 1978; *Nozaki et al.*, 1978] between 1880 and 1955. Third, we report alterations in the correlation between El Niño events and the occurrence of low $\Delta^{14}C$, which is indicative of long-term change(s) in circulation in the SW Pacific. The $\Delta^{14}C$ shifts reported here are not large, but even small temporal changes in prebomb $\Delta^{14}C$ suggest that important changes in the large-scale state of the ocean have occurred, such as a temporal change in circulation.

1. Introduction

Radiocarbon measurements of coralline aragonite reveal that the $^{14}C/^{12}C$ ratio in the dissolved inorganic carbon (DIC) of the seawater surrounding the coral is the same as that at the time of accretion. Multidecadal and multicentury radiocarbon records in corals from the surface regions of the Pacific Ocean have been reported previously [*Druffel*, 1987; *Druffel and Griffin*, 1993; *Konishi et al.*, 1982; *Toggweiler et al.*, 1991]. These records have been used to identify the source of thermocline ventilation in the South Pacific [*Toggweiler et al.*, 1991], to determine the seasonally variant transequatorial transport of waters in the Pacific [*Druffel*, 1987], to indicate the source of the Indonesian throughflow [*Moore et al.*, 1997], and to study the effect of El Niño on upper ocean circulation in the Pacific [*Druffel*, 1981; *Guilderson et al.*, 1998].

Oxygen isotope records in banded corals ($\delta^{18}O$) are mainly controlled by water composition (i.e., seawater salinity) and sea surface temperature (SST) at the time of accretion. These records can be used to identify changes in water mass properties, such as vertical or lateral mixing rates, that may have occurred simultaneously with observed $\Delta^{14}C$ changes. Stable carbon isotope ratios ($\delta^{13}C$) are influenced by several factors including water mass (DIC $\delta^{13}C$), incident light [*McConnaughey*, 1989b], and food availability [*Grottoli-Everett*, 1998].

Druffel and Griffin [1993] have shown that a high-precision $\Delta^{14}C$ record of biennial coral bands from the southern Great Barrier Reef (Abraham Reef in the Swain Reefs off the Queensland coast) contained variations on interannual timescales. These variations were associated with El Niño events and were particularly large between A.D. 1680 and 1730. *Druffel and Griffin* [1993] also found a small Suess effect (4‰), which is the decrease in $\Delta^{14}C$ between A.D. 1880 and 1955, mostly due to dilution of existing carbon with ^{14}C-free CO_2 from the burning of fossil fuels for energy. We identified changes in vertical mixing and in large-scale advective transport of source waters to the western Coral Sea region as likely processes that could account for the $\Delta^{14}C$ patterns.

We report new data from coral collected near Heron Island located 60 km from the Queensland coast and 150 km SW of Abraham Reef. Like previously published results from Abraham Reef [*Druffel and Griffin*, 1993], we see a small Suess effect in the surface waters off Heron Island (2‰). Short-term variability of $\Delta^{14}C$ appears to be, in part, linked to El Niño events; however, this correlation is interrupted during the period 1870 to 1915. These pieces of evidence (small Suess effect and El Niño noncorrelation) occur coincidentally with the decreases we observe in $\delta^{18}O$ values for the late 1800s and early 1900s in both the Heron Island and Abraham Reef corals and point to a large-scale change in the state of the SW Pacific Ocean over the past century.

2. Methods

The coral samples used for this project were *Porites australiensis*; they are from the Heron Island (23°S, 152°E) coral core whose postbomb $\Delta^{14}C$ results were reported earlier [*Druffel and Griffin*, 1995]. Heron Island is located on a shallow reef flat near the inner Great Barrier Reef (GBR). We compare and contrast Heron Island data with previously published data from Abraham Reef, located on the Swain Reefs at the southeastern edge of the outer GBR (22°S, 153°E) where periodic upwelling on a seasonal basis has been documented [*Andrews and Gentien*, 1982]. Coral cores were taken in November 1983 from small colonies (1-2 m diameter) in the region outside the lagoon. In addition to the 350-year, biennial $\Delta^{14}C$ record reported earlier [*Druffel and Griffin*, 1993], nine more biennial samples were measured from the Abraham Reef coral core (see Table 1). Methods used to clean, X-ray, map, and section the annual growth

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Table 1. Δ14C Measurements for all Heron Island and Additional Abraham Reef Samples Not Reported Earlier by *Druffel and Griffin* [1993]

Year	Technique	$\Delta^{14}C$ ‰
	Heron Island	
1848.8	1537	-49.4
1848.8	(1053)	-47.8
1850.8	1515	-52.7
1852.8	1520	-51.3
1854.8	1513	-56.1
1856.8	1538	-48.6
1856.8	(1054)	-53.3
1858.8	1521	-48.5
1860.8	1536	-51.6
1860.8	(1052)	-46.8
1862.8	1535	-44.5
1864.8	1534	-49.6
1866.8	1539	-51.5
1866.8	(1055)	-50.5
1868.8	1519	-53.7
1870.8	902	-50.4
1872.8	1516	-51.3
1874.8	903	-47.9
1876.8	1510	-44.4
1878.8	1303	-45.9
1880.8	801	-51.7
1882.8	1514	-47.3
1884.8	795	-49.0
1886.8	1517	-49.1
1888.8	1302	-47.5
1890.8	802	-49.9
1892.8	1518	-55.5
1894.8	796	-49.5
1896.8	1307	-48.5
1898.8	1301	-42.7
1900.8	814	-48.2
1900.8	826	-51.7
1902.8	1309	-46.0
1904.8	(1063)	-53.9
1906.8	1512	-49.9
1908.8	1300	-47.6
1910.8	794	-41.0
1912.8	1511	-49.1
1914.8	1299	-47.0
1916.8	798	-48.0
1918.8	1298	-46.6
1920.8	538	-54.9
1922.3	536	-45.5
1923.3	1574	-46.6
1923.3	(1059)	-47.1
1924.3	532	-57.8
1925.8	396	-52.9
1927.3	(1058)	-52.2
1928 8	1297	-52.7
1930.8	387	-46.9
1932.3	(1069)	-55.5
1933.8	1296	-46.7
1935.8	406	-41.2
1937.3	1569	-55.9
1938.8	1295	-48.1
1940.8	392	-52.8
1942.8	1290	-49.4
1944.8	389	-45.3
1946.8	1289	-54.8
1948.8	1288	-53.9
1950.3	360	-52.4
1951.3	379	-47.5
1952.8	334	-44.9
1954.3	383	-47.7
1955.3	403	-56.3
	Abraham Reef	
1815.3	(1064)	-44.9
1881.3	1572	-43.6
1893.3	(1374)	-43.7
1927.3	1568	-54.6
1937.3	1565	-53.8
1945.3	1566	-54.1
1948.3	1571	-49.0
1952.3	1570	-56.3
1956.3	1567	-52.5

Technique refers to sample number analyzed using gas proportional beta counting. Numbers in parentheses were obtained using AMS techniques at the University of Arizona.

bands are described elsewhere [*Druffel and Griffin*, 1993; *Griffin and Druffel*, 1985].

Two-year bands were taken from the Heron Island coral that grew prior to 1950 (1-year bands from 1950 to 1983) and subjected to radiocarbon analysis. High-density bands accreted from about October to December of each year. The bands were cut on the leading edge of the high-density bands (October), hence the midpoint of each 2-year band was approximately early spring of the year reported (19XX.8), and the midpoint of each 1-year band was approximately early autumn of the year reported (19XX.3). Approximately 10% of each annual band was lost during cutting.

Most of the samples were analyzed for $\Delta^{14}C$ using conventional gas counting techniques [*Griffin and Druffel*, 1985] at our Woods Hole Oceanographic Institution (WHOI) or University of California, Irvine (UCI) radiocarbon laboratories. Ten samples were analyzed using exclusively accelerator mass spectrometry (AMS) techniques at the University of Arizona tandem AMS facility. Six of the Abraham Reef samples were measured using both gas counting and AMS techniques and results from each method agreed within 3‰. The $\Delta^{14}C$ measurements from conventional counting were of high precision, with 1 σ total uncertainty (counting statistics and laboratory reproducibility) of 2.5-3.0‰ [*Druffel and Griffin*, 1995]. Total uncertainties (1 σ) for the AMS measurements ranged from 2.8 to 4.0‰ [*Druffel et al.*, 1995]. Radiocarbon measurements are reported as $\Delta^{14}C$ values (for age-corrected geochemical samples) according to standard techniques [*Stuiver and Polach*, 1977]. The $\delta^{13}C$ measurements were made on all samples and used to correct the $\Delta^{14}C$ results.

All $\delta^{13}C$ and $\delta^{18}O$ measurements reported here were performed according to standard techniques [*Druffel and Griffin*, 1993] except that the samples were crushed dry (not crushed in methanol and roasted at 375°C for 1 hour prior to isotopic analyses as had been done previously). Biennial samples were cut from slabs of the cores using a band saw, then crushed with a mortar and pestle. Seasonal (seven samples per year based on size of annual growth bands) and annual samples were drilled using a diamond-tipped drill bit mounted on a Dremel tool. Stable isotope results had total uncertainties (1 σ) of $\pm$ 0.1‰ for both $\delta^{13}C$ and $\delta^{18}O$; analyses were performed at either William Curry's or at Lloyd Keigwin's laboratory at WHOI.

3. Results and Discussion

3.1. Stable Oxygen and Carbon Isotopes

3.1.1. Abraham Reef: Annual (1635-1989). Annual $\delta^{18}O$ and $\delta^{13}C$ results for the period 1635-1989 from the Abraham Reef coral are shown in Figure 1. The annual $\delta^{18}O$ results range from -3.8 to -4.6‰ and the least squares fit shows a decrease of 0.41‰ over the entire 354-year period. The earlier biennial isotope measurements from this core reported by *Druffel and Griffin* [1993] were ~ 0.5‰ lower than the annual results reported here (crushed dry and not baked), most likely because of alteration to calcite of the aragonitic samples during baking of the earlier samples.

The annual Abraham Reef $\delta^{13}C$ results (Figure 1) display considerable variability and a general increase of ~ 1.0‰ from 1635 to 1700 and a decrease of ~ 0.5‰ from the mid-1800s to 1955. Between 1955 and 1983, $\delta^{13}C$ values appeared to increase by 0.4 ‰. These annual results agree with the previously published $\delta^{13}C$ values for biennial coral samples, which were crushed in methanol and baked [*Druffel and Griffin*, 1993]. Therefore the baking and methanol appear to have affected only the $\delta^{18}O$ results.

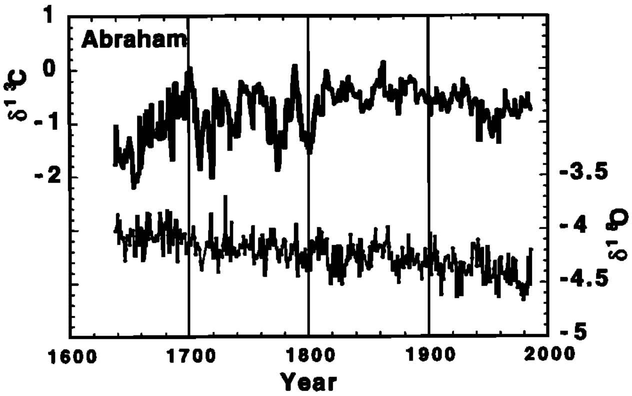

Figure 1. The $\delta^{13}C$ (top curve) and $\delta^{18}O$ (bottom curve) measurements of the annual Abraham Reef coral sample. These samples were run with no pretreatment. Earlier $\delta^{18}O$ and $\delta^{13}C$ values (not shown) reported by *Druffel and Griffin* [1993] were pretreated by precrushing with methanol and heating under vacuum to 375°C for 1 hour to rid samples of organic carbon. These measurements were 0.5‰ lower than those shown here for $\delta^{18}O$ and identical to the $\delta^{13}C$ values shown.

Periodicities within the individual isotope records were investigated using spectral and cross spectral techniques described elsewhere [*Druffel and Griffin*, 1993; *Howell*, 1998]. The spectral density for the detrended annual $\delta^{18}O$ results exhibits significant variance at the 14- and 6.5-year cycles, whereas that for the annual $\delta^{13}C$ results exhibits variance at the 14- and 5-year cycles. A cross-spectral analysis was performed, and the 14-year period was coherent at the 95% confidence level. This coherency may represent similar forcing for both the $\delta^{18}O$ and $\delta^{13}C$ records, though coherency does not necessarily indicate a causal link.

3.1.2. Heron Island: Seasonal (1922-1939). The seasonal $\delta^{18}O$ results from Heron Island coral (Figure 2) display a seasonal range in amplitude from 0.3‰ in 1927 and 1928 to 1.2‰ in 1932 and 1933. Using the empirical relationship between monthly Abraham Reef coral $\delta^{18}O$ values and observed range in monthly SST (-0.16‰ per 1°C rise) at this site from 1981-1993 with advanced very high resolution radiometer (AVHRR), MultiChannel sea surface temperature (MCSST) data [*Bian and Druffel*, 1999], the Heron Island data imply a range in SST of 2°-7°C. Resampling the AVHRR data set at the same sample resolution as the Heron Island coral data (7 samples/yr), a range of 4°-7°C is expected. There is no apparent long-term trend in the seasonal $\delta^{18}O$ data.

The seasonal $\delta^{13}C$ record also shows a recognizable seasonal signal, with amplitude of 0.3 to 1.2‰ (Figure 2). There appears to be a significant trend towards higher values with time, from an average of -0.9‰ in 1922 to 0.0‰ by 1939. There is a direct correlation between the $\delta^{18}O$ and $\delta^{13}C$ seasonal results, where r is 0.56 (number of samples n=123) and is significant to the 99.9%

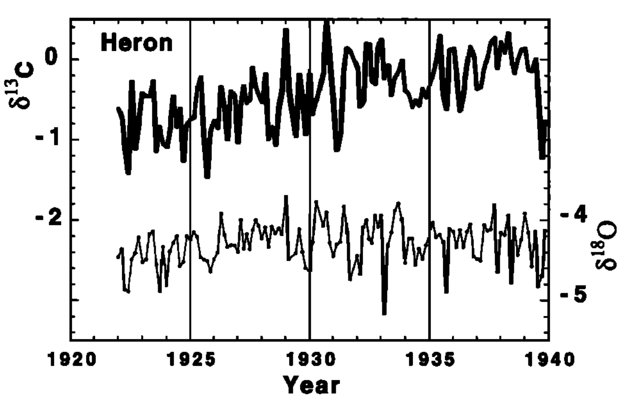

Figure 2. Seasonal $\delta^{13}C$ (top curve) and $\delta^{18}O$ (bottom curve) measurements (~7 per year) for the Heron Island coral.

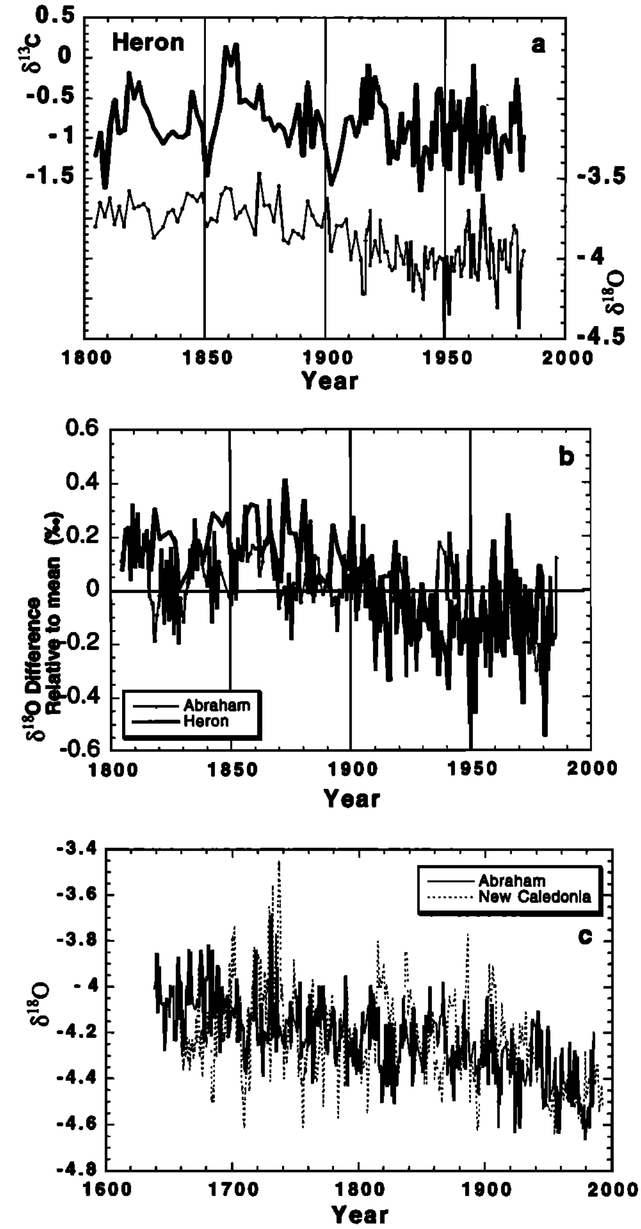

Figure 3. (a) Same as Figure 2, but for annual and biennial Heron Island coral. Samples from 1805-1914 are biennial bands, and those from 1915-1983 are from annual bands, (b) Difference of $\delta^{18}O$ values relative to mean values for Abraham and Heron corals (see Figures 1 and 3a for original $\delta^{18}O$ data sets), (c) The $\delta^{18}O$ values for annual coral bands from Abraham Reef (from Figure 1) and annually averaged (from biannual $\delta^{18}O$ measurements) values from a New Caledonia coral [from *Quinn et al.*, 1998].

confidence level. This is curious because an inverse correlation was observed between monthly $\delta^{18}O$ and $\delta^{13}C$ in Abraham corals for the 1980s [*Bian and Druffel*, 1999]. The probable reason for this difference is that Abraham Reef is influenced by upwelling, which brings low $\delta^{13}C$ DIC with cold waters (that would cause high $\delta^{18}O$ values in corals).

A cross-spectral analysis was performed of the detrended, seasonal $\delta^{18}O$ and $\delta^{13}C$ data; the 1-year period was coherent to the 95% confidence level. These results indicate coherency of the seasonal stable isotope measurements in the Heron Island coral.

3.1.3. Heron Island: Biennial and Annual Samples (1805-1983). The Heron Island coral $\delta^{18}O$ and $\delta^{13}C$ results for biennial samples from 1805 to 1914 and annual samples from 1915 to 1983 are shown in Figure 3a. There are three distinct regions of $\delta^{18}O$ variability. Prior to 1878, the biennial $\delta^{18}O$ values averaged -3.69 ± 0.09 ‰ (standard deviation) (n=34). Between 1878 and 1958, $\delta^{18}O$ values decreased by 0.34‰ (from -3.76‰ to -4.10‰, n=52). Finally, the $\delta^{18}O$ values for the most recent, annual coral bands of this core (> 1958) had an average value of -3.97 ± 0.19‰ (n=32). There was interannual and decadal variability in the $\delta^{18}O$ values.

The $\delta^{13}C$ values (Figure 3a) in the first half of the record revealed oscillations with decadal variability; values rose quickly in the 1810s, 1850s, and 1900s-1910s and decreased slower after each rise. In the years after 1930, variability of $\delta^{13}C$ values was on interannual timescales and was more variable than earlier results. The average $\delta^{13}C$ value for all of the Heron biennial samples was -0.85 ± .38 ‰ (n=119). There was a significant, positive correlation between all of the $\delta^{13}C$ and $\delta^{18}O$ values shown in Figure 3a; a least squares fit has a correlation coefficient r of 0.41 (n=119)(confidence level 99.9%).

3.1.4. Comparisons. When the Heron and Abraham $\delta^{18}O$ records (annual and biennial) are plotted as the difference relative to their means (Figure 3b), both records display gradual trends toward lower values from 1805-1983. During four short periods in the 1810s, 1840s, 1870s and 1890s, Abraham values were significantly lower than those at Heron. Only during the short period of 1935-1945 were values in the Heron coral less than those at Abraham. The source of the differences are likely associated with variations in the water masses that lave the two sites (see section 3.2.4.).

The general trend toward lower $\delta^{18}O$ values in the Australian corals follows that seen in a coral from New Caledonia [*Quinn et al.*, 1998] (Figure 3c). Though not all of the decadal variability is similar between the New Caledonia and Abraham Reef records, it is noteworthy that some of them are the same.

The decrease of $\delta^{18}O$ values by 0.4‰ (Figures 3b and 3c) in Abraham, Heron, and New Caledonia corals represents a temperature rise of ~ 2°C, a change in the water $\delta^{18}O$ (i.e., lower salinity), or a combination of both. Salinity did not change by more than 0.2‰ during the 1960s, but it is impossible to say how it changed over the longer timescale [*Rochford*, 1968]. Several coral records show $\delta^{18}O$ decreases over the past century [*Cole*, 1996; *Quinn et al.*, 1998].

3.2. Radiocarbon

3.2.1. Heron Island - Biennial. The $\Delta^{14}C$ results for the prebomb Heron Island coral (Figure 4 and Table 1) displayed significant variability (5 σ) which ranged from -41‰ in 1911 to -58‰ in 1924. The average value for the 59 $\Delta^{14}C$ results between 1848.9 and 1955.3 was -49.8 ± 3.8‰ (standard deviation). These results are similar to the prebomb Abraham

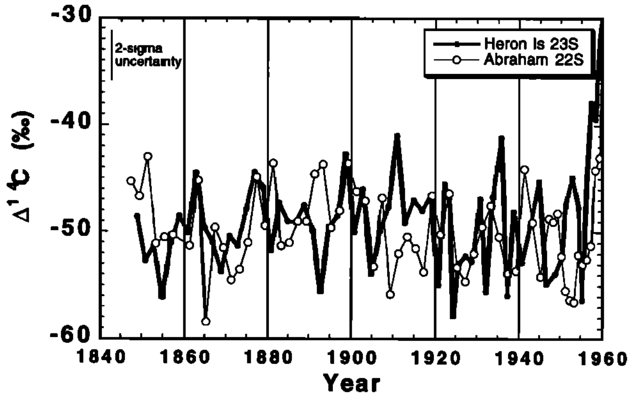

Figure 4. The $\Delta^{14}C$ measurements of biennial coral samples from Heron Island (see Table 1 for actual values) and Abraham Reef [*Druffel and Griffin*, 1993]. Average values are plotted for duplicate measurements listed in Table 1.

Reef results, which ranged from -43‰ in 1851 to -58‰ in 1865 (also shown in Figure 4); the average of 57 Abraham results within this time period was -50.2 ± 3.6‰. The Heron and Abraham $\Delta^{14}C$ records agree within 10‰ (~3 σ) except during five short time periods: 1864-1865, 1892-1993, 1908-1911, 1936-1937 and 1952-1953. After 1956, $\Delta^{14}C$ values at both sites rose owing to the input of bomb-produced ^{14}C to the surface ocean from the atmosphere [*Druffel and Griffin*, 1995].

A least squares fit of the prebomb Heron Island results reveals a decrease of 2.2‰ from 1879 to 1955. Likewise, a decrease of 4.1‰ was calculated for the Abraham Reef data over the same time period [*Druffel and Griffin*, 1993]. It appears that the short-term variability (14-17‰) is much larger than the long-term trend (2-4‰) in each of the GBR $\Delta^{14}C$ records.

The range of $\Delta^{14}C$ results (15‰) is only half of that (30‰) observed in the prebomb $\Delta^{14}C$ records reported from Florida and Bermuda corals for the same time period [*Druffel*, 1997]. Rapid pulses in the rate of ventilation of mode water in the northern Sargasso Sea (a process different from those that control mixing in the SW Pacific) is believed to be the reason for the high variability of prebomb $\Delta^{14}C$ in the North Atlantic [*Druffel*, 1997].

3.2.2. Suess effect: Why is it so small? Using a box diffusion model, the calculated ^{14}C Suess effect in a typical mixed layer from 1880-1950 in a subtropical ocean location would be ~ -11‰ [*Stuiver et al.*, 1986]. This matches $\Delta^{14}C$ measurements of Florida coral for this time period [*Druffel*, 1997; *Druffel and Linick*, 1978]. However, we measure a decrease of only 2‰ in the Heron Island coral and 4‰ in the Abraham Reef coral.

This is despite the presence of a ^{13}C Suess effect in seawater DIC from the South Pacific, measured between 1970 and 1990 [*Quay et al.*, 1992]. These facts are not mutually exclusive, however, as excess CO_2 may have been measurable in the South Pacific Ocean only after the late 1950s, when bomb ^{14}C had masked any evidence of a ^{14}C Suess effect. As there was no ^{13}C produced during the explosions of thermonuclear bombs, the reduction in $\delta^{13}C$ (fossil fuel and biomass burning CO_2 $\delta^{13}C$ = -28‰) of the DIC in the upper few hundred meters of the Pacific Ocean was observable subsequent to 1955 whereas the $\Delta^{14}C$ decrease was masked by bomb ^{14}C. *Quay et al* [1992] reported the presence of excess CO_2 in the ocean from both the burning of fossil fuels and biomass burning as recently as A.D. 1990.

Why is the decrease of $\Delta^{14}C$ in these corals less than that predicted from a box diffusion model? *Toggweiler et al.* [1991] hypothesized that steady state production of Subantarctic Mode Water in the southwestern Pacific (7°-10°C) was the main process responsible for thermocline ventilation of the South Pacific,

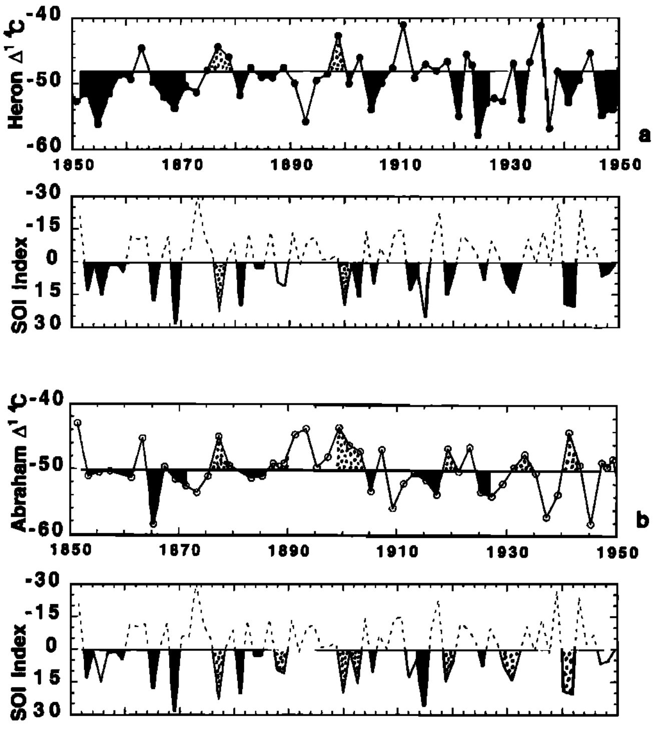

Figure 5. (a) The Wright Southern Oscillation Index (Darwin minus Tahiti air pressure) for December-February [*Wright*, 1989] and the $\Delta^{14}C$ record from the Heron Island corals, (b) Wright SOI and the $\Delta^{14}C$ record from the Abraham Reef coral.

equatorial 13°C water, and the Peru upwelling water. They describe a large-scale overturning of South Pacific water that upwells south of New Zealand and involves the northward diversion and pumping downward of this water into the thermocline. This downward pumping would cause dilution of the fossil fuel CO_2 signal and explain why the ^{14}C Suess effect was small in the southwestern Pacific.

Another explanation of the small Suess effect involves a temporal change in circulation so as to overprint or mask the Suess effect expected in the Australian corals. Reasons to favor this explanation are the coincident decrease in coral $\delta^{18}O$, presumably due to a trend toward warmer temperatures, and the noncorrelation between low $\Delta^{14}C$ values in GBR corals in response to El Niño.

The $\delta^{13}C$ decrease predicted from a box diffusion model for DIC in the subtropical mixed layer is ~ -0.4 to -0.5‰. The observed $\delta^{13}C$ decreases from 1850-1950 are smaller in the GBR corals (Heron Island, -0.1‰ (Figure 3a); Abraham Reef, -0.35‰ (Figure 1)). However, the $\delta^{13}C$ signal in corals is not simply correlated with the $\delta^{13}C$ in DIC. Dependence of coral $\delta^{13}C$ on ambient light, zooplankton abundance [*Grottoli-Everett*, 1998], and other factors [*McConnaughey*, 1989a, b] has been shown to dominate the $\delta^{13}C$ signal.

3.2.3. Correlation Between Low $\Delta^{14}C$ and El Niño Events in the SW Pacific. *Druffel and Griffin* [1993] reported a correlation between El Niño events as recorded for the eastern tropical Pacific region [*Quinn et al.*, 1987] and low $\Delta^{14}C$ measurements in biennial coral samples from Abraham Reef. There is a similar correlation between Heron Island coral $\Delta^{14}C$ and the Southern Oscillation Index (SOI) from the western tropical Pacific [*Wright*, 1989] (see Figure 5a). The correlation is not as good between 1870 and 1920, where there are two El Niño events accompanied by average $\Delta^{14}C$ (1888-1889, 1915, unshaded areas), two by high $\Delta^{14}C$ values (1877 and 1900, stippled areas), and two periods of low $\Delta^{14}C$ with no accompanying El Niño events (1892 and 1929, unshaded areas).

Figure 5b shows the $\Delta^{14}C$ record for the Abraham coral [*Druffel and Griffin*, 1993] plotted with the SOI from the western tropical Pacific [*Wright*, 1989]. The correlation between a high SOI and low $\Delta^{14}C$ values is seen only for the period prior to 1870, whereas $\Delta^{14}C$ tends to be high (stippled areas) during many of the El Niño events after 1870. *Druffel and Griffin* [1993] reported that the decoupling of low $\Delta^{14}C$ and El Niño occurrence (from *Quinn et al.'s* [1987] eastern equatorial Pacific record) ended around 1925. Comparing the Abraham $\Delta^{14}C$ values with the Wright SOI record (Figure 5b) is likely a more reliable portrayal of the southwestern Pacific climate shifts than that determined by *Quinn et al.* [1987] for the eastern Pacific. One does not expect to find exactly the same record of El Niño in the western and eastern tropical Pacific.

3.2.4. Circulation in the SW Pacific: Changes in the large-scale state of the ocean? Changes in advection and vertical transport in the SW Pacific gyre may alter the water mass composition at our coral site and contribute to the variability in coral stable isotope and radiocarbon changes we observe with time. The main source of water to the southern Great Barrier Reef is the westward, subtropical East Australian Current (EAC), and to a lesser extent, the westward, tropical South Equatorial Current (SEC) (see Figure 6). An important feature of the EAC is its intrusion during the summer season onto the shelf where our coral sites lie [*Andrews and Furnds*, 1986; *Tranter et al.*, 1986]. The upwelling caused by this cross-shelf intrusion has a positive relationship with the speed of the EAC [*Ridgway and Godfrey*, 1997]. Also, hydrographic data show that westward flow from the western Coral Sea feeds the EAC [*Church*, 1987].

Druffel and Griffin [1993] hypothesized that lower $\Delta^{14}C$ values measured in corals that grew during El Niño events were the result of a higher relative input of ^{14}C-depleted SEC waters that entered the southern Great Barrier Reef region during these events. From expendable bathythermograph (XBT) results [*Meyers and Donguy*, 1984] and sea level height measurements [*Wyrtki*, 1984], the westward SEC was shown to have flowed in the 10°-20°S latitude band during the 1982-83 El Niño, about 10° further south than during normal years (0°-10°S). This would presumably bring low $\Delta^{14}C$ equatorial waters (influenced by equatorial divergence) from the SEC directly to our coral sites at 22° and 23°S during El Niño.

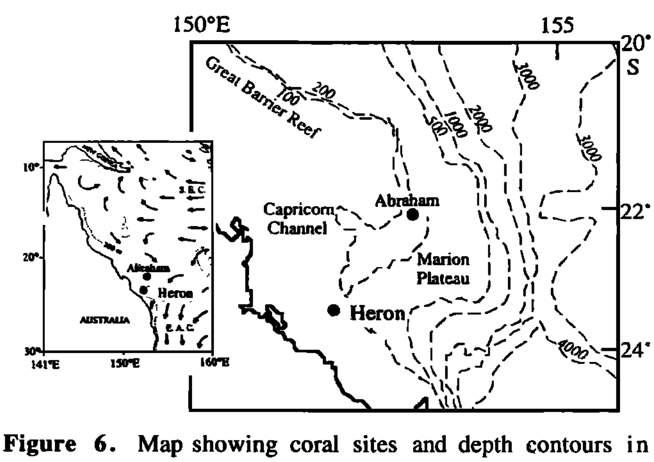

Figure 6. Map showing coral sites and depth contours in meters. (inset) Map of mean circulation of surface currents in the Coral Sea and area farther south during winter (adapted from *Pickard et al.* [1977, Figure 63].

However, the $\Delta^{14}C$ values in GBR corals were not lower during the severe El Niño of 1982-1983 [*Druffel and Griffin*, 1995]. This could be because ^{14}C levels in the SEC were not low due to the cessation of upwelling for a prolonged period during this unusually strong event [*Druffel and Griffin*, 1995]. Few of the strong El Niño events after the 1870s were accompanied by low $\Delta^{14}C$ in the GBR corals (Figure 5a and 5b). The small Suess effects measured in the Abraham and Heron coral records are coincident with the interruption of the El Niño/^{14}C correlation and is likely associated with a long-term (multidecadal) change in surface circulation that occurred in the SW Pacific during the late 1800s and early 1900s.

The post 1870s period was also accompanied by lower $\delta^{18}O$ values at both the Heron Island and Abraham Reef locations (Figure 3b), indicating higher SST and/or lower salinity during this period. Upwelling in the area of the western Coral Sea north of Abraham Reef occurs between 100- and 700-m depth, with a maximum upwelling velocity at 150-200 m [*Godfrey*, 1973; *Tranter et al.*, 1986]. Salinity of the seawater rises from 35.0‰ at surface to 35.6‰ at 150 m depth in the western Coral Sea [*Pickard et al.*, 1977]. This 0.6‰ salinity increase could explain a 0.35‰ maximum increase of skeletal $\delta^{18}O$, which was attributed by *Bian and Druffel* [1999] to the enrichment observed in monthly coral samples from Abraham Reef during El Niño events. A decrease in upwelling from the late 1800s to the early 1900s could explain the observed decrease in $\delta^{18}O$.

4. Summary

We present three observations that indicate changes in the large-scale state of the ocean dominated in the SW Pacific during the late 1800s and early 1900s: (1) decreases in $\delta^{18}O$ of coral from Abraham Reef and Heron Island over the past two centuries, (2) a poorer correlation between low $\Delta^{14}C$ and El Niño events during the period 1870 to 1930 at both of our GBR sites, and (3) a smaller than expected Suess effect at both Abraham Reef and Heron Island.

The general observation of low $\Delta^{14}C$ in GBR coral during most El Niño events is in contrast to the trend seen at locations east of the GBR. The western tropical Pacific is plagued by drought during El Niño, as opposed to higher rainfall found at tropical locations farther east. Our coral isotope records provide an interesting antithesis that could be used in larger-scale studies of the timing of El Niño and the heat and water budgets that are disrupted during these events.

The Suess effect is small in the GBR region, suggesting a possible change in the surface circulation over multidecadal timescales. Long-term changes in $\delta^{18}O$ in the GBR and the similarity with published coral $\delta^{18}O$ records at New Caledonia [*Quinn et al.*, 1998] suggest a long-term warming or freshening of the surface waters of the southwestern Pacific during the past several centuries.

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