DECADAL TIMESCALE SHIFT IN THE ^{14}C RECORD OF A CENTRAL EQUATORIAL PACIFIC CORAL

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ABSTRACT. Coral skeletal radiocarbon records reflect seawater Δ^{14}C and are useful for reconstructing the history of water mass movement and ventilation in the tropical oceans. Here, we reconstructed the inter-annual variability in central equatorial Pacific surface water Δ^{14}C from 1922–1956 using near-monthly ^{14}C measurements in a *Porites* sp. coral skeleton (FI5A) from the windward side of Fanning Island (3°54′32″N, 159°18′88″W). The most pronounced feature in this record is a large, positive shift in the Δ^{14}C between 1947 and 1956 that coincides with the switch of the Pacific Decadal Oscillation (PDO) from a positive to a negative phase in the mid-1940s. Although the absolute Δ^{14}C values from 1950–1955 in FI5A differ from the Δ^{14}C values of another coral core collected from the opposite side of the island, both records show a large, positive shift in their Δ^{14}C records at that time. The relative increase in the Δ^{14}C of each record is consistent with the premise that a common mechanism is controlling the Δ^{14}C records within each coral record. Overall, the Fanning Δ^{14}C data support the notion that a significant amount of subtropical seawater is arriving at the Equator, but does not allow us to determine the mechanism for its transport.

INTRODUCTION

Coral proxy records offer a viable means of recovering pre-instrumental climate and oceanographic information. Scleractinian corals deposit a calcium carbonate skeleton (aragonite) in distinct, annual bands and can grow for several hundred yr. X-radiographic analysis of a thin slab of coral skeleton typically reveals alternating light and dark bands, each pair of which represents 1 yr of coral growth (e.g., Weber et al. 1975; Hudson et al. 1976; Barnes and Lough 1993). Corals incorporate the dissolved inorganic carbon (DIC) of the surrounding seawater as aragonite, a crystalline form of calcium carbonate, into their exoskeleton.

Coral radiocarbon (Δ^{14}C) values have been shown to reflect the Δ^{14}C of seawater DIC (Druffel 1997) and are useful for determining the history of water mass movement and ventilation in the subtropical and tropical oceans (e.g., Nozaki et al. 1978; Druffel 1987; Guilderson and Schrag 1998a; Druffel et al. 2001). ^{14}C is produced naturally in the stratosphere and was also created anthropogenically as a result of thermonuclear weapons explosions in the stratosphere in the late 1950s and early 1960s.

Δ^{14}C in seawater DIC is a natural tracer of upper-ocean circulation. Ocean waters from below the thermocline have lower Δ^{14}C values than surface waters because they are isolated from the atmosphere, the source of ^{14}C. Therefore, surface concentrations of seawater Δ^{14}C are sensitive to upwelling and vertical mixing. Prior to 1957, non-polar surface water ^{14}C values ranged from –38‰ in mid-gyre regions to –72‰ in the eastern equatorial Pacific, where upwelling is very strong (Table 1). Here, we reconstructed the pre-bomb record of Δ^{14}C in surface waters of the central equatorial Pacific from 1922–1956 from Fanning Island and found a profound change in ^{14}C that correlated with a shift in the climate of the North Pacific Ocean.

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METHODS

FI5A is a longitudinal slab cut from a 75-yr-long coral core taken from a live head spanning 1922–1997. The coral core was collected at 11 m depth in October 1997 outside the lagoon on the eastern (windward) side of Fanning Island (3°54′32″N, 159°18′88″W). The core length is 126.0 cm with an average yearly growth of 1.7 cm. This study focuses on the natural variation in coral Δ^{14}C in the pre-bomb (pre-1955) period. We chose to focus our study on the pre-bomb period, when ambient atmospheric Δ^{14}C levels are believed to be relatively constant.

Samples were collected for ^{14}C analysis every millimeter by milling a 14-mm-wide by 4-mm-deep trough along the major axis of growth using a Dremel tool with a diamond-tipped drill bit. Care was taken to follow the curvature of the horizontal growth lines within each sample in order to minimize chronological smearing. For 1922–1956, the ^{14}C value was measured on every other sample, yielding an average of 8 sub-annual samples per yr. Approximately 50 mg of coral was acidified overnight at room temperature in a small glass 50 ml volume acidification vessel under vacuum and a 1.6 ml subsample of the resultant CO_2 gas was reduced with hydrogen gas on an iron metal catalyst to produce a 0.8 mg graphite target (Vogel et al.). δ^{13}C was measured on a second subsample of CO_2 gas using a Finnigan MAT 251 at NOSAMS (δ^{13}C = the ‰ deviation of ^{13}C:^{12}C relative to the v-Peedee Belemnitel Limestone Standard). The ^{14}C content of the graphite was measured using accelerator mass spectrometry (AMS) techniques and reported as Δ^{14}C (the per mil deviation of ^{14}C/^{12}C of the sample relative to that of the 95% Oxalic Acid-1 standard) (Stuiver and Polach 1977). All Δ^{14}C values were corrected for fractionation to a δ^{13}C of –25‰. One-third (104 samples) of the AMS measurements were performed at NOSAMS (National Ocean Sciences AMS Facility at the Woods Hole Oceanographic Institution) with an uncertainty of ±5.0–7.0‰. Two-thirds (209 samples) of the AMS measurements were made at CAMS (Center for Accelerator Mass Spectrometry at Lawrence Livermore National Laboratories) with an uncertainty of ±3.0‰ or less. Numerous analyses of an internal laboratory reference coral standard averaged –59.0 ± 5.4‰ (1 standard deviation, 14 samples) at NOSAMS and averaged –60.3 ± 3.6‰ (1 standard deviation, 20 samples) at CAMS. The average Δ^{14}C values of the internal laboratory reference standard did not significantly differ between laboratories (t-test: t = 0.86, degrees of freedom = 32, p = 0.4), though the precision of measurements at CAMS is higher than that at NOSAMS.

RESULTS

Figure 1 shows Δ^{14}C values measured at roughly 6-week intervals for the 35-yr pre-bomb period of the record from 1922–1956. Overall Δ^{14}C values range from –71‰ in 1944 to –17‰ in 1951. Pre-1947 values range from –71‰ to –39‰ and post-1947 values range from –52‰ to –17‰. Seasonal-to-interannual variations in the Δ^{14}C record are not statistically detectable by spectral analysis. The most pronounced feature in this record is the noticeable positive shift in the Δ^{14}C beginning in 1947 that lasts until 1956. The overall yearly average Δ^{14}C increased by 17‰ from –55 ± 5‰ (1 standard deviation, sample size = 217) to –38 ± 8‰ (1 standard deviation, sample size = 78) after 1947. Differences in the Δ^{14}C measurement precision between NOSAMS and CAMS do not interfere with our ability to interpret our results because the 17‰ shift in Fanning coral Δ^{14}C is much greater than the reported measurement errors.

An element of complexity enters our interpretation of the Δ^{14}C when we compare the results for FI5A with those from another coral from the leeward side of Fanning Island (CTFN). An annual resolution Δ^{14}C time series is available from the CTFN coral, located 11 km to the west of the FI5A site, on the opposite side (leeward side) of the island at ~10 m depth (3°55′N; 159°24′W) (Druffel 1987). Both FI5A and CTFN Δ^{14}C records show an increase in the Δ^{14}C in the 1940s to 1950s, but
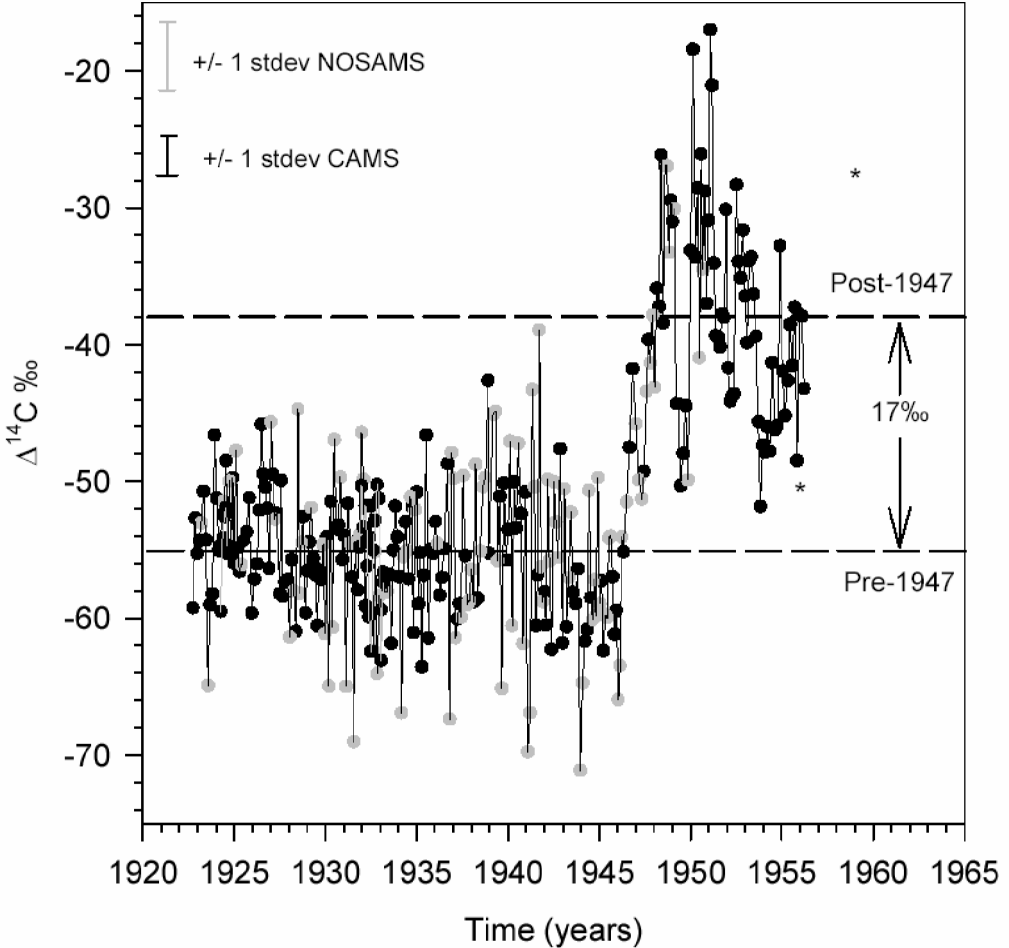
Figure 1 Near-monthly skeletal Δ^{14}C (‰) of Fanning Island *Porites* coral (FI5A) from 1922–1956. Δ^{14}C averaged –55‰ from 1922–1947 and –38‰ from 1947–1956, as indicated by dashed lines. Individual Δ^{14}C values measured at CAMS (•) and NOSAMS (•) are plotted and connected with a solid black line. Two bulk annual Δ^{14}C measurements for 1956 and 1959 (*) are plotted. Areas of overlap between 2 adjacent drilling paths (light gray lines) closely match and are barely visible.

differ in timing and absolute Δ^{14}C values (Figure 2). Differences in the timing of the Δ^{14}C increases may be due to uncertainties in the age model of the CTFN record of ±1–2 yr. The 20–25‰ differences in absolute Δ^{14}C values in the two Fanning cores may be because the leeward side of the island, where the CTFN core was collected, may have a shallower mixed layer and experience more upwelling than the windward location, where FI5A was collected. Alternatively, the difference in the absolute Δ^{14}C may indicate that there are larger uncertainties in coral Δ^{14}C records, at least at some sites, than had originally been believed (Druffel et al. 1989) and demonstrates the need to analyze Δ^{14}C in multiple corals at some sites. Nonetheless, similar changes in the Δ^{14}C values of each record are consistent with the premise that a common mechanism is controlling the Δ^{14}C records within each coral record.
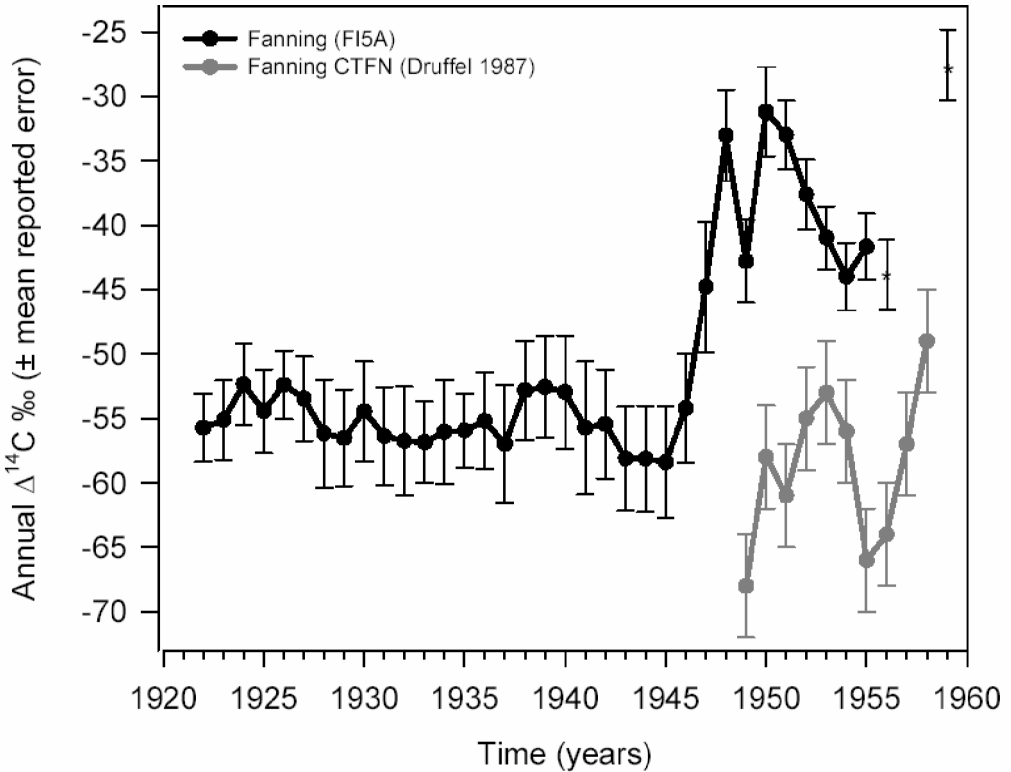
Figure 2 Annual coral skeletal $\Delta^{14}C$ in FI5A Fanning Island (solid black line) and CTFN Fanning Island (solid dark gray line) (Druffel 1987). * Indicates a single bulk sample for the entire year for FI5A in 1959.

DISCUSSION

The positive shift in $\Delta^{14}C$ at Fanning Island beginning in 1947 indicates that a shift in the $\Delta^{14}C$ composition of the surface waters began at that time. Two potential sources for the high $\Delta^{14}C$ to the central equatorial Pacific are discussed below. In addition, we suggest that the phase switch in the Pacific Decadal Oscillation (PDO) beginning in 1942 may have been the mechanism that triggered the rise in $\Delta^{14}C$.

Bomb Source of ^{14}C

Thermonuclear bombs exploded in the stratosphere between 1955 and 1963 and produced bomb ^{14}C, which was incorporated into seawater and then into coral as it grew. Limited testing of ground-level nuclear weapons took place between 1952 and 1954 that was about 6% of the total bomb ^{14}C produced (950×10^{26} atoms). Druffel (1987) suggested that it was possible for a small amount of close-in fallout of locally-produced bomb ^{14}C from early bomb tests to have caused increases in ^{14}C as early as 1952. High $\Delta^{14}C$ values can also occur if post-bomb carbonate recrystalizes in pre-1955 coral skeleton. However, normal to high coral Sr/Ca ratios from 1947–1956 indicate that there was no detectable evidence of recrystalization from aragonite to calcite (D Schrag, unpublished data). The large $\Delta^{14}C$ shift at Fanning starting in 1947 appears inconsistent with expected $\Delta^{14}C$ changes due to bomb sources and, therefore, seems likely to have had an oceanic origin.

Table 1 Mean pre-bomb reconstructed Pacific Ocean surface seawater Δ^{14}C values (± 1 standard deviation of the mean), as reconstructed from corals and shells, and as estimated from the distribution of post-bomb dissolved inorganic carbon (DIC) Δ^{14}C values.

Region or Current Specific location (lat[1], long[2]) (years)	Δ^{14}C (‰) ± 1 SD[3] (sample size)	Reference and source material[4]
Central Equatorial Pacific		
Fanning (4′N, 159′W) (1949–1955)	–60 ± 7 (7)	(Druffel 1987); c
Fanning (4′N, 159′W) (1922–1946.8)	–55 ± 5 (217)	This paper; c
Fanning (4′N, 159′W) (1947–1956)	–38 ± 8 (78)	This paper; c
Northwestern Subtropical Pacific (KC)[5]		
Okinawa (26′N, 127′E) (1912–1954)	–38 ± 4 (19)	(Konishi et al. 1981); c
Northwestern Tropical Pacific		
South China Sea (3–17′N, 104–120′E) (1863–1945)	–51 ± 9 (10)	(Southon et al. 2002); c, b, g
North Pacific Gyre		
Mid-Gyre Region (25′N, 160′W)	–40 (1)	(Druffel 1985);[6] DIC
Hawaii (20′N, 156′W) (1893–1952)	–48 ± 4 (74)	(Druffel et al. 2001); c
Northeastern Tropical Pacific		
Panama (7′N, 81′W) (1950–1955)	–58 ± 5 (4)	(Druffel 1987); c
Southeastern Pacific (SEC/PC)[7]		
Galapagos (0′S, 90′W) (1930–1954)	–72 ± 5 (21)	(Druffel 1981); c
South Pacific Gyre		
Mid-Gyre Region (20′S, 160′W)	–45 (1)	(Druffel 1985);[6] DIC
Rarotonga (21′S, 160′W) (1950–1956)	–52 ± 5 (124)	(Guilderson et al. 2000); c
Fiji (18′S, 179′E) (1930–1955)	–61 ± 7 (14)	(Toggweiler et al. 1991); c
Fiji (18′S, 179′E) (1950–1955)	–62 ± 8 (4)	(Toggweiler 1983); c
Southwestern Pacific		
East Australia (22′S, 153′E) (1950–1955)	–54 ± 2 (5)	(Druffel and Griffin 1993); c
East Australia (23′S, 153′E) (1950–1956)	–49 ± 5 (6)	(Druffel and Griffin 1993); c

[1]lat = latitude
[2]long = longitude
[3]SD = standard deviation of the mean
[4]c = coral, b = bivalve, g = gastropod
[5]KC = Kuroshio Current
[6]estimates of pre-bomb seawater DIC Δ^{14}C values, as reported by Druffel 1985
[7]SEC = South Equatorial Current, PC = Peru Current

North Pacific Gyre Source of High ^{14}C

Estimates of pre-bomb (pre-1955) Δ^{14}C values (± 1 standard deviation) for surface seawater DIC in the Pacific Ocean are listed in Table 1. The lowest value of –72‰ was found at the Galapagos Islands in the eastern South Equatorial Current (SEC) (Druffel 1981) and the highest values of –38‰ were found in the Kuroshio Current (Konishi et al. 1981). Elsewhere in the western Pacific and within the subtropical gyres, Δ^{14}C values ranged between –40‰ and –65‰. In our FI5A core, pre-

1947 $\Delta^{14}C$ values averaged $-55 \pm 5\%o$, falling between the estimated $\Delta^{14}C$ values in the SEC and the estimated $\Delta^{14}C$ values in the mid-northern and mid-southern gyres. After 1947, FI5A $\Delta^{14}C$ rose to an average of $-38 \pm 8\%o$, approximately matching values in the North Pacific Gyre (NPG) surface waters of $-40\%o$ and the Kuroshio current of $-38\%o$.

High $\Delta^{14}C$ NPG surface water is subducted and transported southwestward towards the western equatorial Pacific, where it becomes one of the major sources both for the Kuroshio Current and the eastward flowing North Equatorial Counter Current (Liu et al. 1994; Gu and Philander 1997; Zhang et al. 1998; Cai and Whetton 2001; Zhang et al. 2001). Therefore, an increase (decrease) in the proportion of NPG waters contributing to equatorial surface waters through vertical advective processes could account for an increase (decrease) in the surface ocean $\Delta^{14}C$ at Fanning. However, other recent modeling studies indicate that seawater temperature anomalies arriving at the Equator from the subtropics via this mechanism are likely to be relatively small (Schneider et al. 1999; Pierce et al. 2000; Hazeleger et al. 2001).

Variation in the lateral advection of water could also have contributed to the shift in seawater $\Delta^{14}C$ at Fanning. Fanning is situated at the boundary between the eastward flowing North Equatorial Counter Current (NECC) and the westward flowing SEC. Expendable bathythermograph (XBT) observations from the 1970s and 1980s suggest that Fanning may be supplied by either the NECC or the SEC depending on the season and on interannual variations in the equatorial Pacific flow (Taft and Kessler 1991). Modeling results suggest that, although seasonal variability in $\Delta^{14}C$ in the central equatorial Pacific is relatively small, interannual variability could potentially be large due to the advection of high $\Delta^{14}C$ water from further west (Rodgers et al. 1997). High $\Delta^{14}C$ NPG surface waters contribute to the more southerly NEC waters and further become entrained in the NECC in the western equatorial Pacific. In addition, relatively low $\Delta^{14}C$ water in the portion of the SEC north of the Equator feeds the NECC by turning northward near the date line (Rodgers et al. 1997). A shift in the relative contributions of the SEC and NEC to the NECC could result in a shift in the $\Delta^{14}C$ values of surface seawater at Fanning Island. Our Fanning $\Delta^{14}C$ data indicate that a significant amount of subtropical seawater is arriving at the Equator, but do not allow us to determine the mechanism for the transport of subtropical seawater to the central equatorial Pacific.

Possible Trigger for the Shift in ^{14}C

Most importantly, the dramatic positive shift in the Fanning $\Delta^{14}C$ record beginning in 1947 coincides with the negative shift in the mid-1940s of the Pacific Decadal Oscillation (PDO) (Figure 3) (Mantua et al. 1997). The PDO is an index of slow, large-scale changes in the mode of Pacific Ocean variability with timescales of 20–30 yr, derived from the leading principle component of monthly sea surface temperature (SST) anomalies in the North Pacific Ocean poleward of 20°N (Mantua et al. 1997; Minobe and Mantua 1999). When the PDO is in a warm phase (i.e., the PDO index is positive), the North Pacific has cooler SSTs and higher sea level pressure (SLP) than normal, while in the central and eastern equatorial Pacific, SSTs are warmer and the SLP is lower than normal. The reverse is true during a PDO cool phase. Variations in the PDO correspond to decadal variability generated by ocean-atmosphere processes within the equatorial Pacific (Kirtman and Schopf 1998; Timmermann and Jin 2002). The mid-1940s shift in PDO polarity has been linked to decadal scale changes in a broad range of instrumental and proxy data (Mantua et al. 1997; Zhang et al. 1998; Minobe and Mantua 1999; Biondi et al. 2001; D'Arrigo et al. 2001) but not specifically in the equatorial Pacific. Our Fanning Island $\Delta^{14}C$ record suggests that there is a link between the PDO shift in the 1940s and a shift in the water masses arriving in the central equatorial Pacific at that time. Our $\Delta^{14}C$ record shifts to a more positive mean value, approximately 4.5 yr after the onset of the negative

PDO phase (Figure 3). This is consistent with the estimated 3.5–6 yr travel time from the NPG to the Equator reported in modeling studies (Liu et al. 1994; Zhang et al. 2001) and is a bit fast relative to the modeled 6–9 yr travel time calculated by Cai and Whetton (2001).
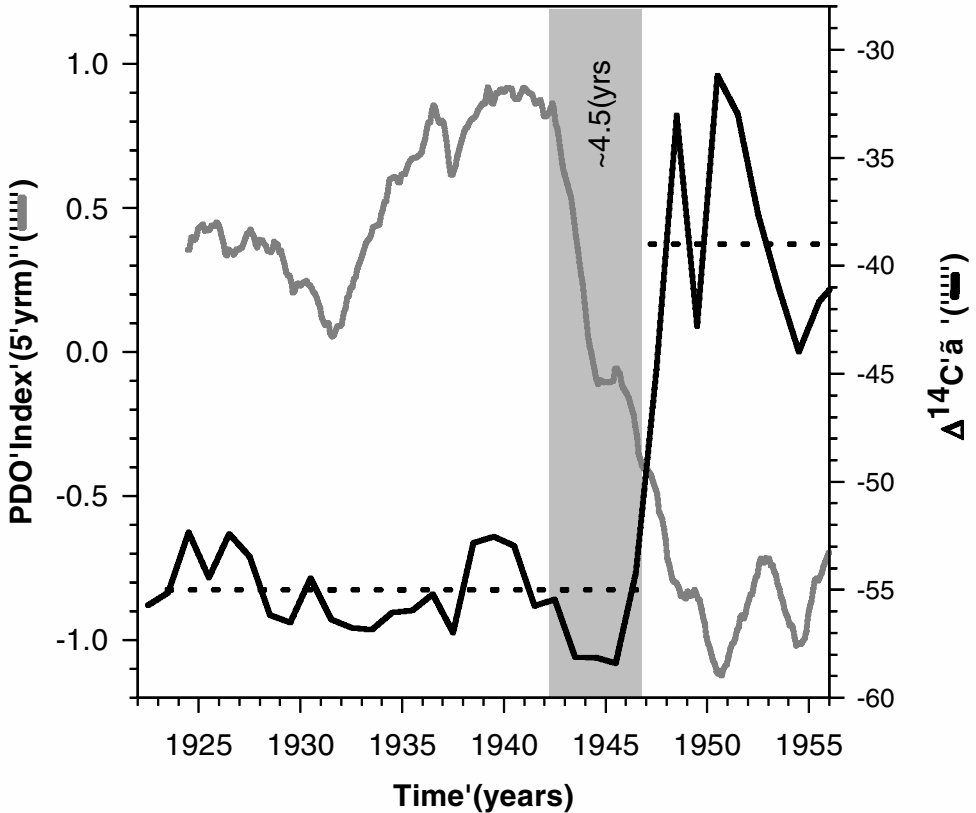

Figure 3 Yearly mean Δ^{14}C and PDO Index from 1922–1956. The PDO Index has been smoothed with a 5-yr (60-month) running mean. Time from the onset of the negative phase in the PDO Index to the increase in yearly mean Δ^{14}C above the long-term pre-1947 mean is approximately 4.5 yr and is indicated by the shaded bar. Dashed lines correspond to the overall mean coral Δ^{14}C from 1922–1947 and 1947–1955. PDO Index data was obtained with permission from the University of Washington's Joint Institute for the Study of the Atmosphere and Oceans, http://jisao.washington.edu/pdo.

Synthesis

Of the few pre-bomb time series of Δ^{14}C available from the tropical Pacific, most show no major changes associated with the PDO. For example, no sharp transition in Δ^{14}C occurs in the mid-1940s at Hawaii (Druffel et al. 2001), Australia (Druffel and Griffin 1993), Nauru (Guilderson and Schrag 1998b), or the Galapagos (Druffel 1981). However, in the south Pacific gyre, the 140-yr δ^{18}O record from a Moorea (17°30′S, 149°50′W) coral shows a strong interdecadal signal with warmer and/or less saline water in the SPG between 1940 and 1960 (Boiseau et al. 1999). A Δ^{14}C record from this site is not available. In a Rarotonga coral (21°4′S, 159°49′W), the Sr/Ca-derived sea surface temperature record indicates a strong interdecadal component that is synchronous with the PDO Index (Linsley et al. 2000). The limited 1950–55 Δ^{14}C record from the same Rarotonga coral is too short to discern a clear signal prior to 1955 (Guilderson et al. 2000). In the available Δ^{14}C coral records,

only Fanning indicates a mid-1940s shift. In other available proxy records, only Fanning, Rarotonga, and Moorea indicate a mid-1940s shift in Sr/Ca or δ^{18}O (Boiseau et al. 1999; Linsley et al. 2000; Dunbar, unpublished). These findings suggest that decadal scale climate fluctuations may be more easily detected in the central Pacific than at its periphery.

Overall, near-monthly isotopic measurements in a *Porites* sp. coral skeleton from Fanning Island revealed a positive shift in the mean Δ^{14}C of 17‰ beginning in 1947 and lasting through at least 1956. The discussion presents intriguing evidence that the positive shift in the Δ^{14}C values of the Fanning Island coral during 1947–1956 is consistent with a decadal timescale introduction of high Δ^{14}C NPG water in the central equatorial Pacific and/or a shift in the proportionate contribution of SEC and NEC surface waters to the NECC. These results suggest that the distinct switch to a negative PDO phase in the mid-1940s could act as the trigger for a shift in ocean circulation. Pre-1950 Δ^{14}C measurements from additional central equatorial corals would enhance our ability to confirm the presence of, and to fully resolve, the extent of the mid-1940s circulation shift in the central Pacific.

ACKNOWLEDGEMENTS

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