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Journal
Geophysical Research Letters, 42(5)

ISSN
0094-8276

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Publication Date
2015-03-16

DOI
10.1002/2014gl062997

AGU PUBLICATIONS

Geophysical Research Letters

RESEARCH LETTER

10.1002/2014GL062997

Key Points:

- 14C and 18O values in near monthly Galapagos corals are inversely correlated
- Intense upwelling is identified in the east equatorial Pacific in early 1800s
- Central Mode Water from North Pacific supplied the EUC

Supporting Information:

- Texts S1 and S2 and Figures S1–S5

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Citation:
Druffel, E. R. M., S. Griffin, D. Vetter, R. B. Dunbar, and D. M. Mucciarone (2015), Identification of frequent La Niña events during the early 1800s in the east equatorial Pacific, *Geophys. Res. Lett.*, *42*, 1512–1519, doi:10.1002/2014GL062997.

Received 29 DEC 2014
Accepted 11 FEB 2015
Accepted article online 16 FEB 2015
Published online 13 MAR 2015

Identification of frequent La Niña events during the early 1800s in the east equatorial Pacific

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Abstract We report measurements of near monthly $\Delta^{14}C$ and $\delta^{18}O$ during selected decades from an east equatorial Pacific coral that grew during the past four centuries. We find that El Niño events occurred regularly during the late 1700s. During the early 1800s, El Niño events occurred less often, and La Niña conditions prevailed, which were accompanied by unprecedented, low cool season $\Delta^{14}C$ values and high cool season $\delta^{18}O$ values. These results indicate that shallow overturning water (e.g., Central Mode Water) from the North Pacific was likely an important source of water to the Galapagos area during the early 1800s.

1. Introduction

Radiocarbon (^{14}C) in dissolved inorganic carbon (DIC) in seawater is primarily a tracer of water mass circulation and upwelling. The replacement time of the water in the mixed layer with subsurface water is fast (months), yet it would take 10 years to exchange all of the DIC in the mixed layer with CO_2 from the atmosphere by gas exchange. Subsurface DIC contains lower $\Delta^{14}C$ values than those in the surface, because of isolation from the atmosphere and radioactive decay. Thus, the $\Delta^{14}C$ value of the mixed layer (−40 to −70‰ in nonpolar oceans) more closely resembles that in subsurface waters (−50 to −80‰) than that in the atmosphere (0‰).

From a model study of surface $\Delta^{14}C$ values from the tropical Pacific, *Toggweiler et al*. [1991] showed that the source of low $\Delta^{14}C$ water at the Galapagos (−72‰) was from Subantarctic Mode Water (SAMW) formed in the convergence zone at 40°S–50°S. It was reported that two thirds of the Equatorial Undercurrent (EUC) water arriving in the east equatorial Pacific was from the Southern Hemisphere [*Rodgers et al*., 2003]. A second source of low ^{14}C water to the east equatorial Pacific is shallow overturning water from the subpolar North Pacific [*Druffel et al*., 2014; *Rodgers et al*., 2004], which has lower $\Delta^{14}C$ values than waters from SAMW.

For a short-lived coral from the same region off western Isabella Island, *Dunbar et al*. [1994] showed that monthly $\delta^{18}O$ values were inversely correlated with Galapagos instrumental sea surface temperature (SST) between 1965 and 1982. They reported that annual $\delta^{18}O$ measurements for the Galapagos coral sequence used for this current study were a very good indicator of El Niño–Southern Oscillation (ENSO) events. Annual $\Delta^{14}C$ values in this same Galapagos coral [*Druffel et al*., 2007] (Figure 1) exhibited variability on ENSO (3–7 years), decadal, and multidecadal timescales. There were two periods, the early 1600s and the early 1800s, with unusually low annual $\Delta^{14}C$ [*Druffel et al*., 2007] and high $\delta^{18}O$ [*Dunbar et al*., 1994] values (Figure 1). The near monthly $\Delta^{14}C$ and $\delta^{18}O$ records during these and other time periods are the topic of this paper.

In a companion paper, *Druffel et al*. [2014] report calibration of the isotopic signals in this Urvina Bay coral sequence for the period 1939–1954. We found that $\delta^{18}O$ and $\Delta^{14}C$ values are significantly correlated, with $\Delta^{14}C$ lagged by ~2 months. We also found that both the $\Delta^{14}C$ and $\delta^{18}O$ records correlate significantly with Niño 3.4 SST [*Kao and Yu*, 2009] and that the cool season $\Delta^{14}C$ and $\delta^{18}O$ values correlate with the Pacific Decadal Oscillation (PDO) index.

Here we report $\Delta^{14}C$ and $\delta^{18}O$ values for near monthly samples of coral from Urvina Bay for the periods of 1602–1611, 1637–1647, and 1783–1835. We report that El Niño events during the late 1700s occurred with a similar frequency to that in the twentieth century but that La Niña conditions dominated during the early 1800s. We show that having seasonally resolved measurements allows tracking of the seasonal minimum of $\Delta^{14}C$ and $\delta^{18}O$, and that this is obscured by measurements that only represent annual averages.

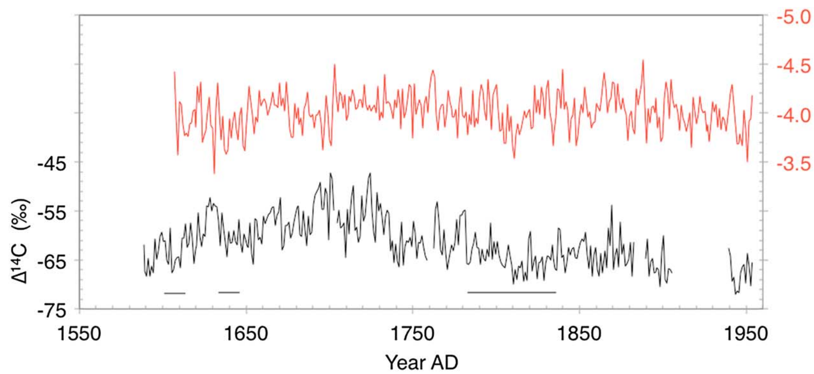

Figure 1. Records of $\Delta^{14}C$ (black) [*Druffel et al*., 2007] and $\delta^{18}O$ [*Dunbar et al*., 1994] (red) values for annual Galapagos coral bands from the UR-86 coral sequence from Urvina Bay. Horizontal black bars show time periods analyzed for near monthly sampling in this work.

2. Methods

The coral sequence used in this study (UR-86) was the same as that used in previous studies reporting annual stable isotope [*Dunbar et al*., 1994] and $\Delta^{14}C$ [*Druffel et al*., 2007] measurements, including a calibration of near monthly data for 1939–1954 [*Druffel et al*., 2014]. Briefly, the coral (*Pavona clavus*) was collected in 1986 from an uplifted reef on the west coast of Isabella Island (0°15′S, 91°22′W) and had been exposed to open ocean waters during its lifetime. The coral sections contained pristine aragonite with no secondary aragonite or calcite based on microscopic and X-ray diffraction analyses. The estimated absolute age assignment error for growth bands at A.D. 1800 is ±3 years and that for the oldest part of the coral core at A.D. 1587 is ±5 years [*Dunbar et al*., 1994]. Our measurements were performed on samples that had been drilled every 1 mm of linear growth with a Dremel tool and diamond bit. Annual bands were assigned mainly using $\delta^{13}C$ seasonality (Figure S1 in the supporting information).

For radiocarbon analyses, coral samples were acidified and the CO_2 was reduced with hydrogen gas on iron powder to produce graphite and analyzed at the Keck Carbon Cycle accelerator mass spectrometry laboratory at the University of California, Irvine [*Druffel et al*., 2014]. Radiocarbon results are reported as Δ values that are corrected for known age to 1950 [*Stuiver and Polach*, 1977]. Total uncertainty of $\Delta^{14}C$ measurements was ±1.4‰. To determine if additional uncertainty was associated with sampling from different areas of the same annual band, eight annual samples were drilled separately from the 1815 band and each sample was analyzed for ^{14}C. The average of the individual $\Delta^{14}C$ values was -69.1 ± 1.1 standard deviation ‰ demonstrating that there was no additional uncertainty associated with sampling from different areas of this coral band.

Stable oxygen and carbon isotope analyses were performed using a Finnigan MAT252 mass spectrometer coupled to a Kiel III carbonate device at Stanford University. Precision on National Institute of Standards and Technology Standard Reference Materials 8544 and internal laboratory standards is better than 0.06‰ and 0.03‰ for $\delta^{18}O$ and $\delta^{13}C$, respectively.

3. Results

3.1. Near Monthly $\Delta^{14}C$ Results

Near monthly $\Delta^{14}C$ values range from a low of −79.5‰ in early 1812 to a high of −51.3‰ in late 1644 (Figure 2). The average of near monthly $\Delta^{14}C$ values is similar for the 1600s (−63.7 ± 2.2‰), 1783–1805 (−63.2 ± 2.9‰), and 1806–1835 (−65.4 ± 2.2‰). In contrast, the average of near monthly $\Delta^{14}C$ values for the 1637–1647 period was significantly higher (−59.4 ± 1.8‰). We note that tree ring $\Delta^{14}C$ values were about 4‰ higher from 1637–1647 than that during the 1600s [*Stuiver and Quay*, 1981] (Figure S2 in the supporting information). A discussion of atmospheric influences on the coral record is presented in section 4.1.

Seasonality is evident during most years, with seasonal magnitudes ranging from 4 to 19‰. Seasonality is nearly absent during the years of 1798–1800, 1819–1820, and 1830. Excluding these years, the average

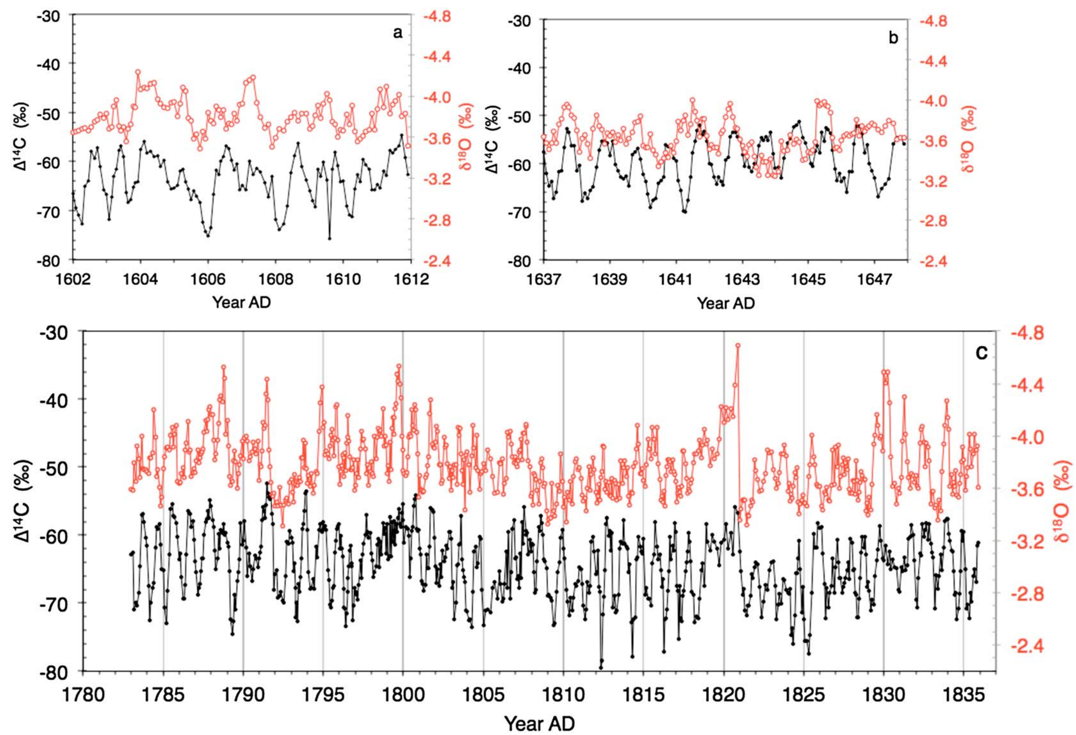

Figure 2. $\Delta^{14}C$ (black circles) and $\delta^{18}O$ (open red circles) values in the Galapagos coral (UR-86) for near monthly (a) 1602–1611, (b) 1637–1647, and (c) 1783–1835 samples. Tic marks on the *x* axes represent 1 January.

seasonal ranges were consistent for the four periods studied (1600s, 12 ± 4‰ $n = 10$; 1637–1647, 11 ± 3‰ $n = 11$; 1783–1805, 12 ± 3‰ $n = 21$; and 1806–1835, 13 ± 4‰ $n = 27$).

Warm season (approximately January to March) high $\Delta^{14}C$ values (the highest $\Delta^{14}C$ value in a given year) were present in nearly all years, and most were within a narrow 3–4‰ range (1600s and 1783–1805, −56 to −59‰; 1637–1647, −52 to −56‰; and 1806–1835, −57 to −61‰) (Figures 2a and 2b and Figure S3 in the supporting information). Cool season (approximately midcalendar year) low $\Delta^{14}C$ values (the lowest $\Delta^{14}C$ value in a given year) were present during most years and were significantly more variable than the warm season high ranges. The cool season values ranged from −65‰ to −72‰ for 50–60% of the years. The remaining cool season low $\Delta^{14}C$ values during the 1600s and 1783–1835 (40 and 38%, respectively) were nearly all lower than −72‰. In contrast, the remaining 46% of the cool season low $\Delta^{14}C$ values during 1637–1648 were all higher than −65‰, in line with the higher $\Delta^{14}C$ baseline for this time period.

3.2. Near Monthly $\delta^{18}O$ Results

Stable oxygen ($\delta^{18}O$) isotope measurements from the near monthly coral samples ranged from four high values of −3.24‰ in 1643 (Figure 1b) to a low of −4.69‰ in late 1820 (Figure 1c). Averages of $\delta^{18}O$ values for the four time periods are similar (−3.81 ± 0.16‰, −3.63 ± 0.17‰, −3.85 ± 0.22‰, and −3.72 ± 0.21‰). Seasonality is evident during most years, where $\delta^{18}O$ values range from 0.3 to 0.8 ‰. Seasonality was low or absent during the years 1643, 1646, and 1792; interestingly, there was normal seasonal variability of $\Delta^{14}C$ values during these years. Excluding these years, the average seasonal ranges were approximately equal for the four time periods (0.4 ± 0.2‰, 0.4 ± 0.1‰, 0.5 ± 0.2‰, and 0.5 ± 0.2‰).

Warm season low $\delta^{18}O$ values were present in nearly all years, and most were within a 0.4‰ range (1600s, −3.8 to −4.2‰; 1637–1647, −3.6 to −4.0‰; 1783–1805, −4.0 to −4.4‰; and 1806–1835, −3.8 to −4.2‰). Cool season high $\delta^{18}O$ values were present during most years and ranged consistently from −3.4 to −3.7‰ for most of the time periods studied. Most of the remaining cool season high $\delta^{18}O$ values during the late 1700s were < -3.7‰ and those during the early 1800s were > -3.4‰, indicating warm and cool SST, respectively.

10.1002/2014GL062997

4. Discussion

We organize the discussion into three parts. First, we describe the influence of the atmospheric CO_2 $\Delta^{14}C$ on the near monthly coral $\Delta^{14}C$ results. Second, we present the results of linear regressions and cross spectral analyses between the $\Delta^{14}C$ and $\delta^{18}O$ records and between the isotopic records and selected climate indices. Third, we discuss how these results relate to circulation in the east equatorial Pacific and source regions of these waters. We identify periods marked by very low, cool season $\Delta^{14}C$ values and discuss the possible influence of volcanic aerosol content in the atmosphere as a driver.

4.1. Influence of Atmospheric CO_2 $\Delta^{14}C$ on the Surface Ocean $\Delta^{14}C$

The $\Delta^{14}C$ record for the Northern Hemisphere atmosphere [*Stuiver et al.*, 1986] is shown in Figure S2, along with the near monthly coral $\Delta^{14}C$ records from Galapagos. If the atmosphere had been a main influence on the surface ocean $\Delta^{14}C$, then the surface ocean should have mimicked the changes in the atmosphere, and been damped to about half the magnitude [*Oeschger et al.*, 1975]. Instead, we see variable differences between atmospheric and coral $\Delta^{14}C$ values that range from 55‰ to 75‰. For example, we observe high atmospheric $\Delta^{14}C$ values during the early 1800s, when the coral $\Delta^{14}C$ values were lowest (Figure S2 in the supporting information). It is clear that gas exchange with the atmosphere was not the primary driver of surface ocean $\Delta^{14}C$ changes during the time periods studied here.

4.2. Correlation Between $\Delta^{14}C$ and $\delta^{18}O$ Results

A linear regression of $\Delta^{14}C$ and $\delta^{18}O$ values reveals significant inverse correlations that are maximized when $\Delta^{14}C$ values are lagged by 2 time steps, ~2 months (1602–1611 $r = -0.49$, $p < 0.0001$, $n = 113$; 1637–1647 $r = -0.24$, $p = 0.004$, $n = 140$; and 1783–1835 $r = -0.53$, $p < 0.0001$, $n = 662$). These correlations support our conceptual understanding of isotopic changes in the east equatorial Pacific that low $\Delta^{14}C$ values (increased upwelling) cooccur with high $\delta^{18}O$ values (low SST). We observed the same correlation and lag in our calibration study from this coral sequence between 1943 and 1954 [*Druffel et al.*, 2014]. Several large shifts from high to low $\Delta^{14}C$ values cooccur with large shifts from low to high $\delta^{18}O$ values (e.g., 1604–1605, 1791, and 1820–1821) (Figure 2). There are a few years when there is no correlation between the two isotope records, and most of these show a small seasonal change in $\delta^{18}O$ values when $\Delta^{14}C$ changes are large (e.g., 1646–1647, 1792–1793, and 1813).

The observed lag indicates that SST changes precede $\Delta^{14}C$ changes by ~2 months. As noted earlier, $\Delta^{14}C$ values in coral are primarily controlled by the $\Delta^{14}C$ value of the source water, whereas $\delta^{18}O$ in coral is primarily a function of SST in this region [*Dunbar et al.*, 1994] and can change on shorter timescales, i.e., by solar heating or evaporative cooling of surface waters. It is notable that when there are clusters of low, cool season $\Delta^{14}C$ values (< -72‰) in the 1783–1835 bands, cooccurring high $\delta^{18}O$ values (> -3.6‰) are seen 80% of the time. Thus, the lowest cool season $\Delta^{14}C$ events are associated with the coldest water (with a ~2 month lag).

There is significant periodicity in the isotope data, other than at the 1 year period. Wavelet analyses [*Torrence and Compo*, 1998] (http://ion.researchsystems.com/IONScript/wavelet/) of the isotope records for the two decades in the seventeenth century are quite different from one another (Figures S4a–S4d in the supporting information). Whereas variance is significant at the 2–3 year period for both the $\Delta^{14}C$ and $\delta^{18}O$ measurements from 1602–1611, there is no such variance in either isotope data set from 1637–1647. Cross spectral analyses [*Howell et al.*, 2006] of the $\Delta^{14}C$ and $\delta^{18}O$ measurements from these two time periods confirm this difference in periodicity (Figures S5a and S5b in the supporting information).

Wavelet analyses of the $\Delta^{14}C$ and $\delta^{18}O$ measurements for the 1783–1835 bands reveals variance concentrated at the 3–7 year periods, similar to ENSO variability (Figures S4e and S4f in the supporting information). Cross-spectral analyses [*Howell et al.*, 2006] of the $\Delta^{14}C$ and $\delta^{18}O$ measurements reveals significant coherency for the 3–4.5 year periods (95% confidence level (CL)) (Figure S5c in the supporting information). Iterative cross spectral analyses (three iterations) reveals that these periods are coherent only for the first and third iterations (1783–1801 95% CL and 1818–1835 80% CL) but not for the middle iteration (1801–1818) (Figures S5d–S5f in the supporting information). Thus, it appears that the coherency of the two isotope records breaks down during the early 1800s, which is when the lowest $\Delta^{14}C$ values appear more frequently.

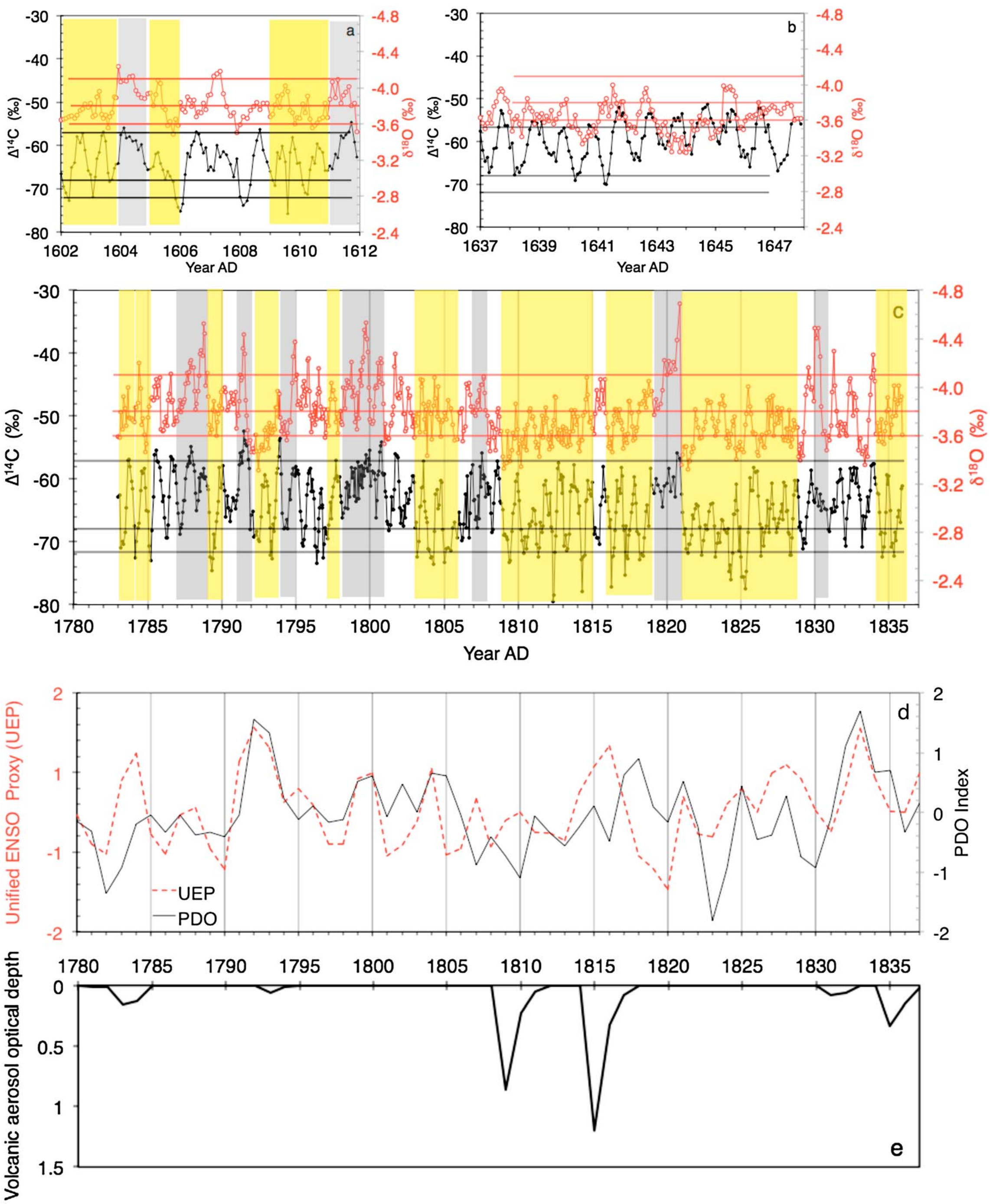

Figure 3. Near monthly Galapagos coral $\Delta^{14}C$ (black circles) and $\delta^{18}O$ (red open circles) values for (a) 1602–1611, (b) 1637–1647, and (c) 1783–1835 with El Niño (grey shading) and La Niña (yellow highlight) events identified by a set of criteria that are based on our calibration study [*Druffel et al.*, 2014] (see text). The thresholds for the $\Delta^{14}C$ values was adjusted by +2‰ to account for the Suess Effect on the calibration data (1939–1954). The values mentioned in the text are indicated by lines on the graphs (red for $\delta^{18}O$ and black for $\Delta^{14}C$). (d) The annual PDO record [*MacDonald and Case*, 2005] (black) and the annual UEP record [*McGregor et al.*, 2010] (red); (e) volcanic aerosol optical depth at the equator [*Robertson et al.*, 2001].

4.3. ENSO Records and Correlations of $\Delta^{14}C$ and $\delta^{18}O$ Values With ENSO and PDO Indices

El Niño events in the Galapagos coral are identified by a set of criteria that are based on data from our calibration study (Figure 3). Briefly, at least one $\Delta^{14}C$ value within a given year was >-57‰, all values were >-68‰, and the annual average was >-64‰; at least one $\delta^{18}O$ value was <-4.1‰, all values were

<-3.6‰ and the annual average was <-3.8‰. Five of six of these criteria were satisfied for data from 1604, 1611, 1787–1788, 1791, 1794, 1798–1800, 1807, 1819–1820, and 1830 (grey shading in Figures 3a–3c).

La Niña events in the coral are also identified based on data from our calibration study (Figures 3a–3c). Briefly, at least one $\Delta^{14}C$ value within a given year was <-72‰, all values were <-57‰, and the annual average was <-64‰; at least one $\delta^{18}O$ value was >-3.6‰, all values were >-4.1‰, and the annual average was >-3.8‰. Five of six of these criteria were satisfied for data from 1602–1603, 1605, 1609–1610, 1783–1784, 1789, 1792–1793, 1797, 1803–1805, 1809–1814, 1816–1818, 1821–1828, and 1834–1835 (yellow highlighted areas in Figures 3a–3c). The incidences of La Niña events are higher during the 1600s and 1803–1835 than for the other time periods. This latter prolonged period of cool, upwelling waters suggests a period of frequent La Niña events (section 4.4).

Moderate, strong and extreme El Niño events reported by *Gergis and Fowler* [2009] cooccur within 3 years of those identified in our coral record for all but two (1611 and 1787) (not shown). Additionally, moderate, strong and extreme La Niña events [*Gergis and Fowler*, 2009] cooccur within 3 years of those identified in our coral record for all but one (1834–1835) (not shown), though they occur at half the rate (10 events from 1803 to 1835, 30%) that we observe in the coral for this time period (70%) (Figure 3c). Because the error of the age model is ±3 years in the 1800s and younger bands, and ±5 years in the 1600s, this agreement is reasonable. From these results, and the significant coherency of the 3–4 year periods (section 4.2), it appears that the $\delta^{18}O$ and $\Delta^{14}C$ data from this coral are reasonable recorders of El Niño and La Niña events.

We present linear regressions of our isotope records from 1783 to 1835 with records of ENSO and PDO. We do not include our earlier isotope records in this comparison, because they are too short (10–11 years) and their age assignments have higher uncertainty.

The ENSO record we use for comparison to our isotope records is the Unified ENSO Proxy (UEP) that consolidates previously defined proxy reconstructions into one record of ENSO [*McGregor et al.*, 2010]. A linear regression of UEP values and cool season $\Delta^{14}C$ values reveals a significant correlation ($r = 0.30$, $p = .03$, and $n = 50$). Likewise, a linear regression of UEP values and warm season $\Delta^{14}C$ values reveals a significant correlation ($r = 0.28$, $p = 0.05$, $n = 50$). We obtained significant inverse correlations for linear regressions of UEP values and cool season $\delta^{18}O$ values ($r = -0.41$, $p = 0.002$, $n = 50$) and UEP values and warm season $\delta^{18}O$ values ($r = -0.49$, $p = 0.0009$, $n = 50$). Each of these correlations is maximized when the isotopic records are lagged by 3–4 years behind the UEP values.

It appears that both cool and warm season $\Delta^{14}C$ and $\delta^{18}O$ values are significantly correlated with the UEP record. Similar results are obtained between our isotope records and the Southern Oscillation Index (see Text S1 in the supporting information). These results agree with correlations in our calibration study between near monthly $\Delta^{14}C$ and $\delta^{18}O$ [*Druffel et al.*, 2014] and a monthly SST NINO 3.4 record [*Kao and Yu*, 2009]. The record of ENSO from our Galapagos record between 1783 and 1835 is consistent with previous ENSO proxies, though our record has more frequent La Niña events during the early 1800s than the other proxies.

The annual PDO index is available for the period A.D. 993–1996 using tree ring chronologies from North America [*MacDonald and Case*, 2005] (Figure 3d). A linear regression of the annual PDO index and cool season $\Delta^{14}C$ values from 1783 to 1835 yields a positive correlation ($r = 0.36$, $p = 0.008$, $n = 51$). This does not agree with the results of our calibration study, which showed a weak inverse correlation ($r = -0.35$, $p = 0.2$, $n = 15$) [*Druffel et al.*, 2014]. A positive correlation was also obtained for a linear regression of the annual PDO and warm season $\Delta^{14}C$ values between 1783 and 1835 ($r = 0.27$, $p = 0.04$, $n = 52$). The annual PDO index and cool season $\delta^{18}O$ values from 1783 to 1835 are negatively correlated ($r = -0.41$, $p = 0.002$, $n = 51$). This agrees with results from the calibration study ($r = -0.80$, $p = 0.006$, $n = 11$) [*Druffel et al.*, 2014]. Similar results were obtained using a separate PDO record [*Biondi et al.*, 2001] (see Text S2 in the supporting information).

The PDO index and the UEP record are positively correlated ($r = -0.36$, $p = 0.007$, $n = 53$) during this period, indicating that periods of high NINO 3.4 SST (e.g., El Niño) cooccur with periods of high PDO (general warming in the East Pacific).

4.4. Prolonged La Niña Conditions During the Early 1800s

Our results reveal that the early 1800s was anomalous, because it was marked by prolonged cool, upwelling conditions. There are 19 years during the 1803–1835 period when cool season $\Delta^{14}C$ values were lower than −72‰ (~60% of the time) and over half of those years had more than one sample lower than −72‰ (1804, 1809, 1812, 1814, 1816–1818, 1824, 1825, and 1828). For comparison, there were 4, 0, and 6 years during the 1602–1611, 1637–1647, and 1783–1802 periods when cool season low $\Delta^{14}C$ values were lower than −72‰ (40%, 0%, and 30% of the time) and only 4 years that had more than one sample lower than −72‰ (1606, 1608, 1789, and 1793).

A similar trend is observed for the cool season $\delta^{18}O$ values, which were higher than −3.6‰ during 24 years from 1803 to 1835 (75% of the time). For comparison, there were 5 and 7 years during the 1602–1611 and 1783–1802 periods, respectively, when cool season low $\delta^{18}O$ values were higher than −3.6‰ (50% and 35% of the time). As noted previously, the 1637–1647 period was anomalously cool (high $\delta^{18}O$) with 8 of the 11 cool season high $\delta^{18}O$ values >−3.6‰, whereas $\Delta^{14}C$ values were relatively high (>−70‰).

The abundance of low cool season $\Delta^{14}C$ values and high cool season $\delta^{18}O$ values during the early 1800s is unique compared to the other time periods studied. Low cool season $\Delta^{14}C$ values and high cool season $\delta^{18}O$ values are characteristic of the cold expression of ENSO, or La Niña, such as during 1949 in the calibration study [*Druffel et al*., 2014].

A recent, multiproxy reconstruction of SST variability in the Niño 3.4 region similarly shows a sharp decline in SST in the early 1800s [*Emile-Geay et al*., 2013a; *Emile-Geay et al*., 2013b]. This cooling has been attributed to the increased frequency of volcanic events between 1803 (Cotopaxi, Ecuador) and 1835 (Coseguina, Nicaragua) [*Crowley et al*., 1997]. Volcanic activity releases aerosols into the atmosphere that cool the Earth's surface and is a likely driver of high $\delta^{18}O$ values and low $\Delta^{14}C$ values indicative of increased La Niña influence in the Galapagos coral [*Druffel et al*., 2007; *Dunbar et al*., 1994]. A record of volcanic aerosol optical depth for the equator [*Robertson et al*., 2001] (Figure 3e) shows that two large peaks, at 1809 and Tambora at 1815, coincide with the period of frequent La Niña events in the coral record (1809–1829). It does not, however, identify the origin of the exceptionally low $\Delta^{14}C$ waters (−80‰) that we observe in the early 1800s. Our La Niña event rate during this period (Figure 3c) is twice that found by other studies [*Gergis and Fowler*, 2009; *McGregor et al*., 2010]. It seems likely that our high rate of La Niña events is, at least in part, influenced by changes in circulation in the Galapagos region.

4.5. Origin of Water Upwelling at Galapagos

To explore the effect of circulation changes on our isotope records, we review the sources of water to the Galapagos region. The Equatorial Undercurrent (EUC) surfaces in the western Galapagos Islands (location of coral) during April and May when the surface flow along the equator is eastward [*Kessler*, 2006]. This is coincident with the cool season $\delta^{18}O$ and $\Delta^{14}C$ values.

Rodgers et al. [2003] reported that the sources of the densest waters to the EUC (sigma theta 26.4–26.9) originate from 40°S to 50°S and 35°N to 42°N. The origin of the EUC water is partially from SAMW that has a $\Delta^{14}C$ value of −72‰ [*Toggweiler et al*., 1991]. Additionally, a northern source of water to the EUC, shallow overturning water from the subpolar north Pacific, has been identified as a major source of low $\Delta^{14}C$ water (~ −87 to −101‰) [*Bien et al*., 1960; *McNeely et al*., 2006; *Robinson and Thompson*, 1981] during the 1940s that is even lower than that of SAMW [*Druffel et al*., 2014]. The waters formed in the North Pacific are subtropical mode water, central mode water, eastern subtropical mode water, and North Pacific Intermediate Water. Central mode water (CMW) has a density (σ_Θ 26.0–26.5) [*Nakamura*, 1996] that is similar to Galapagos waters that come from the less dense part of the EUC. The $\Delta^{14}C$ values of the subpolar North Pacific are lower than our lowest coral values from the early 1800s (−80‰) (Figure 2c). Thus, these waters likely mix with waters that have higher $\Delta^{14}C$ values during their transit to the equator. The unusually low cool season $\Delta^{14}C$ (and high $\delta^{18}O$) values of the early 1800s likely indicate that there was a larger contribution of shallow overturning waters (CMW) from the north into the EUC, and to the western Galapagos, though we cannot demonstrate here that there is a clear link.

Radiocarbon is a potentially unique tracer for the reemergence of mode waters in the equatorial Pacific. The record of variable cool season $\Delta^{14}C$ values appears to be tracking water mass variations, and that the North Pacific (CMW) contributions offer a strong potential to be important, given that the southern sources of EUC water (SAMW) are presumably less dominant for the waters upwelling near the Galapagos because they are too dense.

5. Implications for Future Work

During the late 1700s, El Niño and La Niña periodicity was similar to that observed in the twentieth century, every 3–7 years; however, this was not the case during the early 1800s. La Niña periodicity is less well known, though our data show frequent events during the early 1600s and early 1800s. These cool events are coincident with periods of increased volcanic aerosols though a causal link has yet to be demonstrated.

The data presented here is only one fourth of the available coral bands for the seventeenth through nineteenth centuries from the Galapagos. Records from other locations in the mid-Pacific and western Pacific would help to establish basin-wide trends in ENSO and shed light on the two types of El Niño events, i.e., central-Pacific versus eastern-Pacific [*Kao and Yu*, 2009]. Additionally, radiocarbon can be used as a unique tracer of water origin that can assist in establishing the circulation changes that are associated with climate change, both natural and anthropogenic.

Acknowledgments
We thank the following people for their support and assistance during this project: Mitch Colgan and Gerry Wellington for assistance in collecting the coral, John Southon, Kevin Druffel-Rodriguez, and Cyril McCormick for technical help, two anonymous reviewers for their insightful comments, the W.M. Keck Foundation, the Greg and Donna Jenkins Foundation, the Kavli Foundation, and NSF Chemical Oceanography Program (OCE-0137207 and OCE-0526463 to E.R.M.D.), Earth System History Program (EAR-0402619 to E.R.M.D.) and Climate Dynamics Program (GEO/ATM 0402562 to R.B.D.) provided funding. The data reported in this paper are available at the NOAA National Climatic Data Center under Paleoclimatology data sets.

The Editor thanks two anonymous reviewers for their assistance evaluating this paper.

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