Radiocarbon as a thermocline proxy for the eastern equatorial Pacific

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[1] An ocean model is used to test the idea that sea surface $\Delta^{14}C$ behaves as a thermocline proxy in the eastern equatorial Pacific. The ORCA2 model, which includes $\Delta^{14}C$ as a passive tracer, has been forced with reanalysis fluxes over 1948–1999, and the output is compared with a previously reported Galapagos $\Delta^{14}C$ record. The model reproduces the abrupt increase in the seasonally minimum $\Delta^{14}C$ in 1976/77 found in the data. This increase is associated with neither a shift of thermocline depth over the NINO3 region, nor a change in the relative proportion of Northern/Southern source waters. Rather, it is due to a decrease in the Sub-Antarctic Mode Water (SAMW) component of the upwelling water, thereby representing a decrease in entrainment of water from below the base of the directly ventilated thermocline. *INDEX TERMS:* 4231 Oceanography: General: Equatorial oceanography; 4255 Oceanography: General: Numerical modeling; 4808 Oceanography: Biological and Chemical: Chemical tracers. **Citation:** Rodgers, K. B., O. Aumont, G. Madec, C. Menkes, B. Blanke, P. Monfray, J. C. Orr, and D. P. Schrag (2004), Radiocarbon as a thermocline proxy for the eastern equatorial Pacific, *Geophys. Res. Lett.*, *31*, L14314, doi:10.1029/2004GL019764.

1. Introduction

[2] The abrupt shift in sea surface temperature (SST) in 1976/77 in the eastern equatorial Pacific was originally described by *Graham* [1994] and *Trenberth and Hurrell* [1994]. Later *Zhang et al.* [1997] argued that it represented only the most recent transition in a mode of variability that they referred to as the Pacific Decadal Oscillation (PDO). For each mechanisms which has thus far been proposed for the 1976/77 shift, the role of the Pacific thermocline is different [i.e., *Chang et al.*, 2001]. However, the paucity of pre-1980 measurements is an obstacle to describing a large-scale shift in circulation.

[3] Potentially important clues are provided by a high-resolution $\Delta^{14}C$ time series from a Galapagos (90°W, 0.5°S) coral [*Guilderson and Schrag*, 1998]. For the upper Pacific, $\Delta^{14}C$ needs to be understood within the context of the bomb-^{14}C transient introduced by the testing of atmospheric weapons in the 1950s and 1960s [*Toggweiler et al.*, 1989]. These tests led to elevated $\Delta^{14}C$ levels in atmospheric CO_2, with a peak in 1963. Over the ensuing decades, much of the atmospheric bomb-^{14}C crossed the sea surface via gas exchange, and then entered the subsurface ocean layers through subduction outside of the equatorial band. This invasion flux of bomb-^{14}C raised the spatial gradients of $\Delta^{14}C$ sufficiently to overwhelm the pre-bomb gradients in $\Delta^{14}C$. The transient $\Delta^{14}C$ signal at Galapagos records this dye-tracer eventually reaching the equatorial upwelling region through the thermocline circulation. As the 5-to-10 year air-sea equilibration timescale for $\Delta^{14}C$ is much longer than the residence time of surface mixed layer waters in the upwelling region of the eastern equatorial Pacific, $\Delta^{14}C$ in the mixed layer traces changes in circulation. The seasonal minima in $\Delta^{14}C$ recorded by the coral coincide with the period in boreal fall when the local SST and thermocline depth are minimum, and thus the coral record traces the $\Delta^{14}C$ of thermocline water entrained into the upwelling.

[4] Here we use a global ocean model to interpret the 1976/77 shift in $\Delta^{14}C$ measured on the Galapagos coral. As such, our study builds on the previous work of *Rodgers et al.* [1997] and *Rodgers et al.* [2000], as well as *Toggweiler et al.* [1991], which used GEOSECS $\Delta^{14}C$ measurements to argue for a presence of Sub-Antarctic Mode Water (SAMW) from the South Pacific in the lower Pacific thermocline.

2. Model Description

[5] We use the ORCA2 configuration of the Ocean Parallelisé (OPA) version 8.1 OGCM [*Madec et al.*, 1998]. Zonal resolution is 2°, and meridional resolution is 2° × cos(latitude), increasing to 0.5° at the equator, and there are 30 vertical levels. The model uses the *Gent and McWilliams* [1990] eddy parameterization scheme, and the turbulent kinetic energy (TKE) scheme of *Blanke and*

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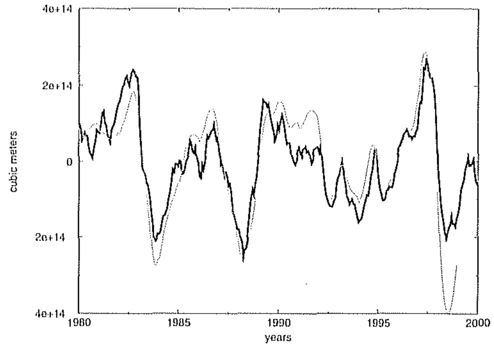

Figure 1. Warm water volume anomaly (WWVA) following the definition of *Meinen and McPhaden* [2000] in units of m^3; INT experiment (thick black line) and data (thing grey line) have correlation of 0.74.

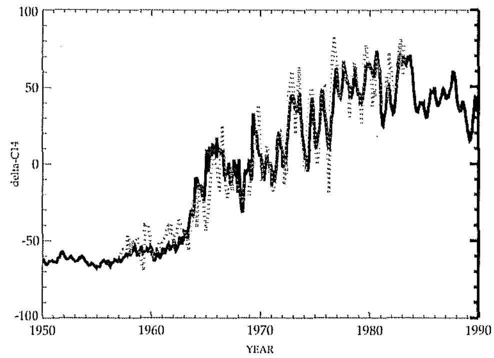

Figure 2. Δ^{14}C at Galapagos (90°W, 0.5°S), for the INT experiment (solid line), and for the data of *Guilderson and Schrag* [1998] (dashed line) in per mil units.

Delecluse [1993] is used for vertical mixing. Surface heat fluxes are calculated using bulk formulas, and surface salinity is restored to the *Boyer et al.* [1998] climatology. Δ^{14}C is treated online following the OCMIP protocol of *Orr et al.* [2001]. The initial Δ^{14}C distribution is taken from a 5000-year offline spin-up. In addition, the offline Lagrangian trajectory analysis tool of *Blanke and Raynaud* [1997] is used.

[6] Two model runs are described here, INT (interannual) and CLIM (climatological). Both are initialized with the *Boyer et al.* [1998] and *Antonov et al.* [1998] salinity and temperature climatologies, respectively, and spun up for 150 years using climatological forcing fields derived from the NCEP reanalysis [*Kalnay et al.*, 1996]. The INT experiment is then forced with the daily NCEP fields (without temporal smoothing) over 1948–1999, while the CLIM experiment continues to be forced with the same climatological fields over the period corresponding to 1948–1999.

3. Results

[7] The Warm Water Volume Anomaly, defined as the volume anomaly above the 20°C isotherm averaged over (120°E–80°W, 5°N–5°S), is shown for INT as well as for the observations [*Meinen and McPhaden*, 2000] in Figure 1. The very good agreement in amplitude suggests that despite a weak bias in the mean state of the NCEP trade winds [*Putnam et al.*, 2000], the anomalies are realistic in the equatorial Pacific. Thus we have decided not to apply any "bias correction", as have other studies [*Giese et al.*, 2002] by applying a multiplicative factor to the NCEP reanalysis fields.

[8] Next we compare the model's sea surface Δ^{14}C for Galapagos for the INT simulation (solid line) with the coral data of *Guilderson and Schrag* [1998] (dots) in Figure 2. The model output has been shifted forward in time by five months in order to correct for a lag in the model simulation of Δ^{14}C at Galapagos. In addition, the coral Δ^{14}C data has been shifted by +18 per mil in the figure to facilitate comparison of the temporal structure of the two curves. For the model, although both SST and thermocline depth are minimum in October for the region near Galapagos, there is upwelling of low-Δ^{14}C water from the lower thermocline along the equator at the eastern boundary. This low-Δ^{14}C water is then advected to the west in the South Equatorial Current (SEC), reaching the area near Galapagos five months later. Nevertheless, the INT simulation (Figure 2) qualitatively captures the temporal variability in Δ^{14}C, as well as the shift in its seasonally minimum Δ^{14}C in 1976/77.

[9] Figure 3 shows the Δ^{14}C for Galapagos from INT during the 1970s and the 20°C isotherm depth (Z20) averaged over the NINO3 region (5°N–5°S, 90°W–150°W). On interannual timescales, the local minima and maxima in Z20 are reflected as local minima and maxima in Δ^{14}C. During the strong 1972/73 El Niño event, Z20 was anomalously deep in the NINO3 region, and this is reflected in high Δ^{14}C values that are characteristic of warmer water from shallower in the pycnocline.

[10] However, the significant shift in seasonally minimum Δ^{14}C in 1976/77 does not reflect a shift in the seasonal minimum of Z20. To better understand the reason for the Δ^{14}C shift, we consider in Figure 4 the model EUC at 151°W for the INT and CLIM runs using Eulerian and Lagrangian diagnostics. As was done by *Rodgers et al.* [2003], the EUC is defined at 151°W as all water traveling eastward between 3°N and 3°S, from the base of the mixed layer to 612 m depth. For each 5-year period between 1951 and 1995 (i.e., 1951–1955, ..., 1991–1995) a monthly circulation climatology has been constructed. Figures 4a and 4b show the vertically integrated EUC transport across

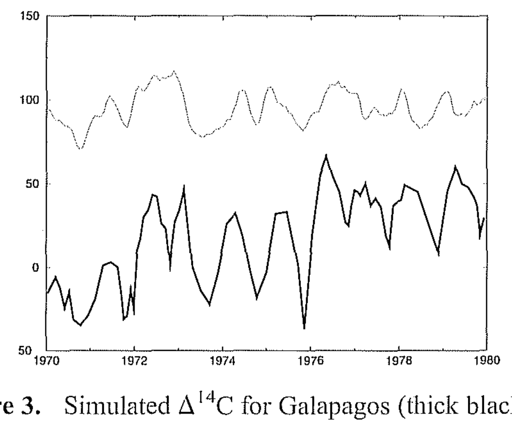

Figure 3. Simulated Δ^{14}C for Galapagos (thick black line) in per mil units, and thermocline depth averaged over the NINO region (thin grey line) in meters, both for INT.

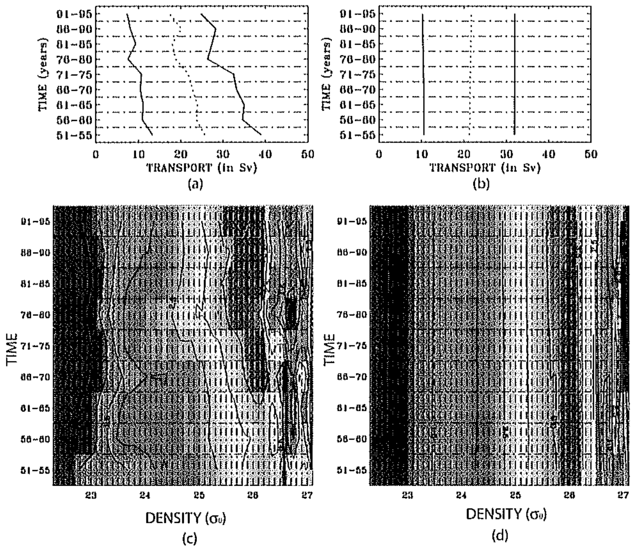

Figure 4. Model EUC transports at 151°W: (a) vertically integrated for INT in Sv; (b) vertically integrated for CLIM in Sv; (c) as a function of density for INT; and (d) as a function of density for CLIM. For (a) and (b), the thick solid line is the total, the dashed line represents the Southern Hemisphere component, and the thin solid line represents the Northern hemisphere component.

151°W for the INT and CLIM cases. Figure 4a reveals a weakening of the EUC between the 1960s and the 1980s from approximately 35 Sv to approximately 28 Sv. The fact that there is no such weakening of the EUC for CLIM (Figure 4b) indicates that for INT the weakening is not due to model drift. The weakening in Figure 4a is consistent with the weakening of the Pacific subtropical cells (STCs) reported by *McPhaden and Zhang* [2002].

[11] For each 5-year climatology, Lagrangian trajectory analysis has been used to determine the hemisphere of origin of the particles crossing 151°W. The proportional decrease in time of the Northern and Southern components is such that the Northern component comprises 30–35% and the Southern component comprises 65–70% of the EUC water. The relative stationarity of the hemispheric component proportions is inconsistent with the scenario proposed by *Rodgers et al.* [1999], according to which significant changes in the relative proportions of Northern/Southern component source waters in the EUC could help to account for decadal changes in SST.

[12] Figures 4c and 4d show the Eulerian transport of the EUC as a function of density (σ_0) across 151°W (again between 3°N and 3°S) from 1951–1995. For each of the 5-year periods, the colors/contours show the fraction of the transport for that period in each 0.1 density interval. Thus the changes in transport through time shown have been normalized. For INT (Figure 4c), the normalized transport maximum reveals a shoaling in time. During the 1960s, the transport is maximum at σ_0 = 26.2, with a secondary maximum at σ_0 = 26.6. During the 1980s, the transport is maximum between σ_0 = 25.5 and 26.2. Thus during the 1970s there was an abrupt drop in the mode water component of the EUC. The corresponding figure for CLIM (Figure 4d) reveals that under climatological forcing the model drift is small, and thus the shift in Figure 4c represents a model response to surface forcing.

4. Discussion

[13] We have found that the INT run captures the temporal structure of $\Delta^{14}C$ variability of the Galapagos coral over the period 1958 to 1983. The model is able to reproduce the 1976/77 shift in the seasonal minimum of the $\Delta^{14}C$ signal. Eulerian diagnostics reveal that this shift coincides with a shoaling of the EUC transport maximum in density space. The weakening of the EUC is consistent with a weakening of the Subtropical Cells (STCs). Importantly, the decrease in the fraction of water from below the σ_0 = 26.2 horizon suggests a decrease in the Sub-Antarctic Mode Water component of the upwelling.

[14] In order to understand how the INT simulation is able to capture the temporal structure of $\Delta^{14}C$ variability for Galapagos on interannual timescales, despite the five-month lag, it is important to consider that the equatorial Kelvin waves that cause Z20 variations at Galapagos will also cause Z20 variations (with the same sign) at the eastern boundary. For the shoaling of the EUC transport in density space, the reduced mode water (σ_0 > 26.2) component of upwelling waters after 1976/77 is reflected in the upwelled signal at both Galapagos and the eastern boundary. This model bias in the phase of the seasonal cycle of sea surface $\Delta^{14}C$ has associated with it an under-representation of the amplitude of the seasonal cycle in $\Delta^{14}C$, as can also be seen in Figure 2.

[15] It is our hope that this study will provide a dynamical framework for interpreting previously reported decadal shifts in coral $\Delta^{14}C$ [*Grottoli et al.*, 2004; *Druffel*, 1981]. For $\Delta^{14}C$, the 1976/77 shift is distinct from earlier shifts in that it coincides with the large spatial $\Delta^{14}C$ gradients associated with the bomb ^{14}C transient. The studies of *Grottoli et al.* [2004] and *Druffel* [1981] indicate that the prebomb-$\Delta^{14}C$ variability may be sufficiently large to record low frequency changes in circulation.

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