Fingerprints of climate change in North Pacific oxygen

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[1] Oxygen gas dissolved in seawater has been studied as a tracer of physical and biological processes in the ocean for nearly half a century. Analysis of historical oxygen data has revealed widespread changes in subsurface oxygen concentrations over the past few decades, providing important constraints on the impact of late 20th century climate change on the circulation and biological productivity of the ocean. We report results from a hind cast ocean circulation/biogeochemical model that reproduces the spatial patterns of observed subsurface O_2 variability in the North Pacific, where inferred O_2 trends are strongest. We find that decadal North Pacific O_2 variations in the lower ventilated thermocline primarily reflect changes in the basin's large-scale circulation. A southward expansion of the model subtropical gyre explains the observed subtropical O_2 increase from the 1980s to the 1990s, while the simultaneous O_2 decreases seen throughout the midlatitude Pacific are driven largely by reduced communication between the atmosphere and the ocean interior. Similar O_2 decreases are pervasive among mid- and high-latitude water masses, but further research is needed to determine whether these changes reflect a global response to 20th century climate change. **Citation:** Deutsch, C., S. Emerson, and L. Thompson (2005), Fingerprints of climate change in North Pacific oxygen, *Geophys. Res. Lett.*, *32*, L16604, doi:10.1029/2005GL023190.

1. Introduction

[2] Large-scale changes in oceanic O_2 concentrations during the latter decades of the 20th century have now been detected in nearly every basin of the world's oceans [*Bindoff and McDougall*, 2000; *Emerson et al.*, 2001; *Garcia et al.*, 1998; *Matear et al.*, 2000; *Ono et al.*, 2001; *Shaffer et al.*, 2000; *Watanabe et al.*, 2001; G. C. Johnson and N. Gruber, Decadal water mass variations along 20°W in the northeastern Atlantic Ocean, submitted to *Progress in Oceanography*, 2005]. Concentrations of O_2 measured in different decades at nearby locations have consistently been found to have decreased in shallow subsurface waters of the mid- and high-latitude oceanic thermocline (100–1000 m depth). Such O_2 changes provide unique clues to the response of the ocean's circulation and biological productivity to climate forcing, and their similarity across ocean basins suggests that the response may be global in scale. The degree to which observed O_2 changes reflect long-term trends has not been established however, and the relative role of physical and biological processes in causing the O_2 changes has not been quantified.

[3] Throughout most of the surface ocean, air-sea gas exchange causes O_2 concentrations to remain close to thermodynamic equilibrium with the atmosphere. Once surface waters are subducted into the ocean interior, O_2 concentrations decrease from near their saturated values due to the respiration of organic matter, which consumes O_2 along the path of circulation. At any location in the ocean interior, variations in O_2 through time therefore reflect: 1) changes in the thermodynamic O_2 saturation and/or air-sea O_2 disequilibrium of surface waters, 2) changes in the ventilation of the ocean interior through the subduction or sinking of O_2-rich surface waters, 3) changes in the rate of subsurface biological O_2 consumption, and 4) changes in the speed and/or pathway by which a water parcel is circulated through the field of O_2 consumption. The total change in oxygen, ΔO_2, is the sum of anomalies from the respective processes:

$$\Delta O_2 = \Delta O_2^{sat} + \Delta O_2^{vent} + \Delta O_2^{bio} + \Delta O_2^{circ} = \Delta O_2^{sat} - \Delta AOU \quad (1)$$

where AOU (Apparent Oxygen Utilization), defined as the difference between saturated and observed oxygen (AOU = $O_2^{sat} - O_2$), integrates the effects of ventilation, biology, and circulation.

[4] The largest amplitude and most thoroughly documented O_2 changes have been observed in the North Pacific (see review by *Emerson et al.* [2004]), and are reflected in AOU changes, indicating a dominant role for physical and biological mechanisms, rather than thermodynamic changes (ΔO_2^{sat}). Variability in the physical conditions of the North Pacific is a prominent feature of the global climate system [*Trenberth and Hurrell*, 1994]. Patterns of decadal sea surface temperature and atmospheric pressure variability (the Pacific Decadal Oscillation, or PDO) have been linked to changes in the intensity of the Kuroshio current [*Deser et al.*, 1999], and in mixed layer depth [*Polovina et al.*, 1995], which can strongly effect basin-scale circulation and ventilation rates. Changes in the marine biosphere of the North Pacific have also been observed over the last several decades. Trends in chlorophyll and primary productivity in the subtropical gyre [*Karl et al.*, 2001; *Venrick et al.*, 1987] and ecological oscillations among higher trophic levels have been observed, and appear to be associated with transitions in the physical climate system [*Chavez et al.*, 2003]. Quantifying the combined effects on oceanic O_2 of biological and physical variability occurring over a range of space and time scales requires a 3-dimensional model of ocean circulation and biogeochemistry.

2. Methods

[5] We use an isopycnal-coordinate General Circulation Model (GCM) configured for a North Pacific domain
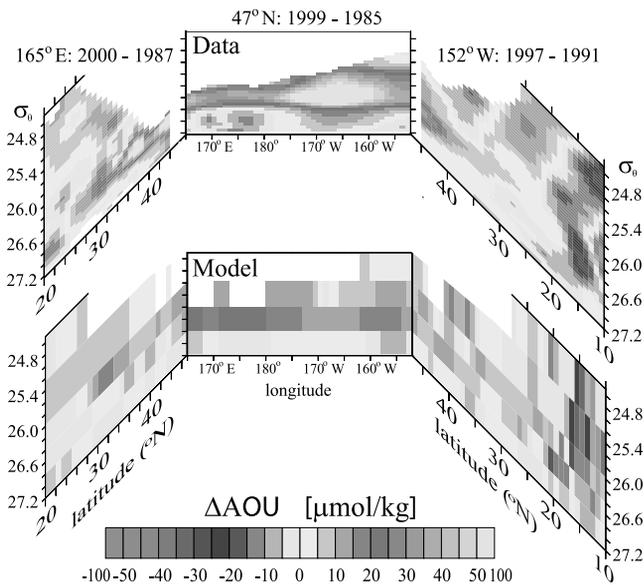
Figure 1. AOU difference (μmol/kg) between two time periods representing approximately the 1980s and 1990s, along three sections through the North Pacific from (a) measurements on repeated cruise transects and (b) a hind cast general circulation/biogeochemistry model. AOU data sources are as referenced by *Emerson et al.* [2001, 2004]. Sections are mapped for each cruise at regular density intervals (y-axis) using a "loess" interpolation and then subtracted as $AOU_{later} - AOU_{earlier}$. Model output are annual mean values taken along each section from corresponding cruise years. Values poleward of the wintertime outcrop are not shown. See color version of this figure in the HTML.

extending from 20°S to 60°N and from 90°E to 70°W [*Hallberg and Rhines*, 1996; *Ladd and Thompson*, 2001]. The model is forced at the sea surface by atmospheric winds and air-sea heat fluxes computed from bulk formulae and surface air temperature. Sea surface salinity is restored toward its climatological distribution [*Levitus and Boyer*, 1994]. In addition, temperature, salinity, and biogeochemical tracers are restored toward climatological values within 5° of the domain's open boundaries.

[6] The sources and sinks of biogeochemical tracers are computed following the Ocean Carbon Model Intercomparison Project [*Doney et al.*, 2004], in which the export of biological organic matter from surface waters into the ocean interior is diagnosed by restoring model PO_4 toward climatological values [*Conkright et al.*, 2002] in the upper 75 m. The spatially variable organic matter export flux is instantly remineralized with depth, decreasing downward with an effective depth scale of ~300 m [*Martin et al.*, 1989]. Oxygen utilization rates (OUR) in the interior are proportional to the rate of PO_4 remineralization, with a constant stoichiometric ratio of 170 moles of O_2 per mole of PO_4. Surface O_2 concentrations are set to saturated values.

[7] Model nutrient and O_2 distributions are integrated to a steady state using climatological mean atmospheric forcing. The resulting O_2 distributions are in close agreement with observed annual mean values [*Levitus and Boyer*, 1994], suggesting that the physical supply and biological consumption of O_2 in the North Pacific are adequately represented in the model.

[8] Steady state tracer distributions are used as initial conditions for a hind cast simulation, in which changes in ocean circulation are forced at the surface by historical atmospheric winds and surface air temperatures from NCEP reanalyses between 1948–2000. Changes in sea surface salinity (or equivalently, net air-sea fresh water flux) are not represented. Surface PO_4 is restored toward the same monthly climatology throughout the simulation; however, biological export flux and the associated OUR, varies in proportion to physically driven changes in nutrient supply. Since there is little reliable data about whether or how surface nutrient distributions changed in the late 20th century, we make the conservative assumption that they have not.

3. Results

[9] Model O_2 variability, which is nearly identical to AOU variability but of opposite sign, can be tested by comparing observed AOU changes along three repeated cruise transects [*Emerson et al.*, 2004] to model output sampled during the same years. These ocean "snapshots" are separated by an average of a decade and correspond roughly to the 1980s and 1990s. Although the cruises used in these comparisons occurred during different months, the seasonal O_2 cycle appears to be unable to account for the phasing or magnitude of intercruise O_2 differences [*Emerson et al.*, 2001].

[10] The pattern of simulated AOU differences over this period bears a strong resemblance to the available observations (Figure 1). Across most of the subarctic section, model AOU rises by 5–30 μmol/kg from the 1980s to the 1990s, with the largest increases in waters with densities of $\sigma_\theta \sim$ 26.6, which represent the densest waters to outcrop during winter in the North Pacific (we therefore refer to $\sigma_\theta \sim$ 26.6 as the lower ventilated thermocline). The magnitude of elevated AOU becomes smaller above and below this density, and toward the subtropics. In the eastern subtropics near 25°N, AOU decreases (O_2 increases) appear throughout the upper water column. The primary features in the simulated AOU difference sections can be identified in the observations, although the model underestimates the magnitude of the observed AOU rise in the subpolar region along 47°N, by a factor of 2–3.

[11] Decadal mean AOU differences between the 1980s and 1990s in the model's lower ventilated thermocline (σ_θ 26.6) suggest that the AOU changes inferred from individual cruise lines during specific years are representative of decadal and basin-scale AOU differences. Modeled AOU increases along 47°N reflect a basin-wide loss of O_2 throughout most of the lower ventilated thermocline in the subpolar Pacific, while the subtropical AOU decrease seen in the eastern subtropical section near 25°N is part of a broad, decadal O_2 increase extending across the basin.

[12] Changes in O_2 driven by biological and physical processes are reflected in AOU, which can be written as the sum of a "preformed" value (AOU_0, defined as the AOU at the base of the mixed layer), and the O_2 utilization rate (OUR) integrated along each displacement, ds, of the water

Table 1. Numerical Experiments

Experiment Design	Causes of AOU Variability
Full variability	$\Delta AOU_{bio} + \Delta AOU_{vent} + \Delta AOU_{circ}$
Climatological OUR	$\Delta AOU_{vent} + \Delta AOU_{circ}$
Climatological OUR and AOU_o	ΔAOU_{circ}

parcel as it is transported with velocity, v, into the ocean interior:

$$AOU = AOU_o + \int OUR * ds/v \qquad (2)$$

AOU_o, which is equal to AOU at each grid point where an isopycnal layer intersects the base of the mixed layer, depends on the degree of isopycnal exchange with (i.e., ventilation by) the mixed layer. The total change in AOU (ΔAOU) results from changes in its preformed value, changes in the speed or pathway of the circulation (denoted "circ"), or changes in OUR, so that the derivative of equation 2 can be written schematically as:

$$\Delta AOU = \Delta AOU_o + \Delta OUR*circ + OUR*\Delta circ \qquad (3)$$

The terms on the right hand side can be identified as AOU changes due to ventilation ($\Delta AOU_{vent} = \Delta AOU_o$), biology ($\Delta AOU_{bio} = \Delta OUR*circ$), and circulation ($\Delta AOU_{circ} = OUR*\Delta circ$), respectively.

[13] On the basis of equation 3, we perform two additional simulations with the same time-dependent circulation fields, designed to isolate and characterize the contribution of individual processes to total AOU changes (see Table 1). In a second simulation, we remove the effect of changes in biology by fixing the OUR to its climatological pattern. In the third simulation, we remove the influence of changes in both biology and ventilation on the distribution of AOU. In addition to holding OUR constant, we force the AOU on each isopycnal surface to remain at its initial value wherever that surface intersects the mixed layer either in the model's climatological mean state or at any time during the variable circulation simulation. This manipulation, which ensures a constant "preformed" AOU at the base of the mixed layer, removes the effect of changes in the location and/or strength of ventilation on the AOU distribution. While additional processes, such as changes in air-sea heat flux and biological O_2 production could also alter preformed AOU, by setting surface O_2 at its saturated concentration, we effectively assume that the influence of these surface processes on O_2 are rapidly removed by air-sea gas exchange. Using all three simulations we can map the contribution of biology, circulation, and ventilation to the total AOU change. Each term in equation 3 is equal and opposite of the corresponding ΔO_2 term in equation 1 and can be added to the small thermodynamic term (ΔO_2^{sat}) to yield the total ΔO_2.

[14] Model biological productivity undergoes large regional changes, with decadal differences between the 1980s and 1990s exceeding 50% of the mean value (C. Deutsch et al., Physical-biological interactions in North Pacific oxygen variability, submitted to *Journal of Geophysical Research*, 2005) Because the respiration of sinking organic matter attenuates rapidly with depth however, productivity changes have little influence on decadal O_2 variability in the lower ventilated thermocline, where biologically driven AOU changes are very small (<5 μmol/kg, Figure 2b). Instead, AOU anomalies at this density are caused primarily by physical processes. Circulation variability constitutes the largest direct cause of AOU changes in the tropical-subtropical latitudes (Figure 2c), while ventilation changes are the primary cause of widespread AOU increases in the subpolar region (Figure 2d).

[15] The specific causes of simulated AOU differences between the 1980s and 1990s in the lower ventilated thermocline can be traced to both decadal trends and to episodic perturbations in the physical state of the North Pacific. Ventilation rates on σ_θ 26.6 undergo a long-term decrease due largely to the shrinking area of its wintertime outcrop, which reduces the communication of atmospheric properties to the lower ventilated thermocline. Because the observed freshening of waters formed in the Northwest Pacific [*Wong et al.*, 1999] is not represented in the model, the simulated reduction in ventilation on this surface may be an underestimate, causing subpolar AOU increases to be weaker than observed. The subtropical AOU decrease is caused by a southward expansion of the subtropical gyre from the 1980s to the 1990s, whereby low-O_2/high-AOU waters in the North Equatorial Countercurrent are displaced along $\sim$15°–20°N by relatively well oxygenated waters in the gyre, a shift that is also reflected in salinity. Circulation related AOU changes are also observed in the Kuroshio Extension whose strength and location shift in response to atmospheric forcing, qualitatively consistent with hydrographic shifts associated with the PDO [*Deser et al.*, 1999].

[16] Episodic physical perturbations also contribute significantly to model decadal O_2 variability. Two notable perturbations occurring in the early 1970s in the Kuroshio
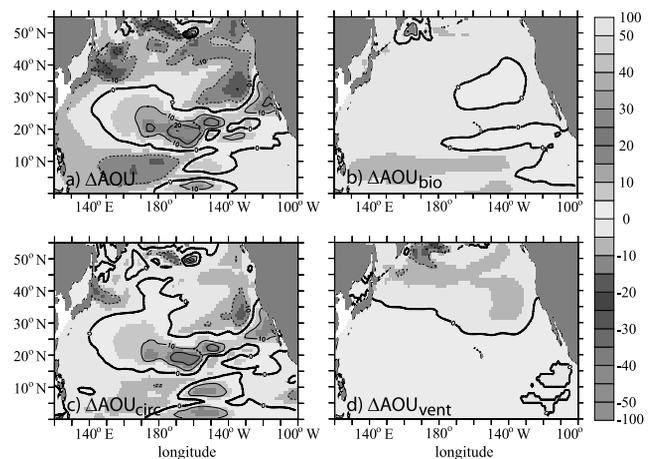
Figure 2. Difference between decadal mean AOU (μmol/kg) in the 1990s and the 1980s along the isopycnal surface σ_θ 26.6. Total AOU difference (a) is the sum of AOU anomalies due to biological changes (b), changes in circulation (c), and ventilation changes (d). The decadal AOU difference is nearly equal but opposite of the total decadal O_2 change, since thermodynamically driven O_2 changes are negligible. Contour lines are shown at intervals of 5 μmol/kg, with dashed lines indicating negative AOU changes. See color version of this figure in the HTML.

Extension and Eastern Tropical North Pacific generate O_2 anomalies of sufficient scale (widths of $\sim 20°$ longitude) and intensity ($\Delta O_2 \sim 50$ μmol/kg) that they remain coherent features for over a decade. The multi-decadal trend toward reduced ventilation on σ_θ 26.6 is punctuated by a brief period of increased ventilation in the early 1970s. This episodic ventilation event injects a large bolus of high O_2 water into the permanent thermocline that travels eastward and then southward into the subtropical gyre, producing an increase in O_2 from the 1970s to the 1980s, followed by an O_2 decrease from the 1980s to the 1990s. Similarly, part of the O_2 increase in the center of the subtropical gyre (Figure 2a) reflects a return to background O_2 levels following the passage of a transient low-O_2 anomaly transported westward from the Eastern Tropical North Pacific in the early 1970s, reminiscent of recent observations by *Lukas and Santiago-Mandujano* [2001].

4. Summary

[17] Observed changes in oceanic O_2 provide a unique fingerprint of the response of ocean circulation and biology to climate change, and therefore a stringent test for ocean circulation/biogeochemistry models. We have presented an ocean hind cast simulation forced by historical climate change that reproduces the overall patterns of observed O_2 variability in the North Pacific, and allows the biological and physical causes to be identified.

[18] Within the lower ventilated thermocline, O_2 is most sensitive to changes in the rates and pathways of circulation and ventilation, making it a reliable indicator of changes in the physical state of the oceanic thermocline. The observed O_2 decreases in the subpolar to subtropical North Pacific are consistent with a long-term reduction in ventilation, while the O_2 increases observed along the southern flank of the subtropical gyre are well simulated by a shift in the gyre boundary. Additional subtropical O_2 variability is non-local in origin, involving pathways of water mass transport from the gyre edges, where O_2 gradients are strong. The transient nature of such anomalies complicates the determination of long-term O_2 trends based on a single pair of ocean snapshots in different decades. Continued analysis of historical O_2 data and ongoing measurement programs are necessary to determine whether O_2 changes observed over the past several decades in the North Pacific and elsewhere reflect long-term anthropogenic climate trends, or transient fluctuations associated with decadal climate variability.

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