Climate variability in the North Pacific thermocline diagnosed from oxygen measurements: An update based on the U.S. CLIVAR/CO\textsubscript{2} Repeat Hydrography cruises

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[1] New observations of oxygen variability in the North Pacific Ocean are reported on the basis of comparison of the U.S. Climate Variability and Predictability and Carbon (CLIVAR/CO\textsubscript{2}) Repeat Hydrography sections conducted along 30\textdegree N (2004) and 152\textdegree W (2006) with the earlier World Ocean Circulation Experiment (WOCE) data and other cruises along these sections. The largest changes in apparent oxygen utilization (AOU) continue to occur, as found in earlier North Pacific repeat section analyses, within the thermocline on $\sigma_\theta = 26.6$ kg m\textsuperscript{-3}, which is the densest isopycnal to outcrop in the open North Pacific in climatological data. In the northeastern North Pacific along 152\textdegree W, where a total of five cruises (1980, 1984, 1991, 1997, and 2006) spanning a period of 26 years are available, the AOU changes correspond to an overall increase in AOU on $\sigma_\theta = 26.6$ kg m\textsuperscript{-3} from the 1980s/early 1990s to 2006. However, from 1997 to 2006 a decrease in AOU is observed within the boundary region between the subtropical and subpolar gyres at 40\textdegree–45\textdegree N. Along the center axis of the subtropical gyre at 30\textdegree N, where two cruises are available (1994 and 2004), AOU has also substantially increased on $\sigma_\theta = 26.6$ kg m\textsuperscript{-3} from 1994 to 2004 in the eastern part of the section. The repeat section data along 152\textdegree W and 30\textdegree N are consistent with a pattern of decadal-scale ventilation anomalies that originate in the northwestern Pacific, possibly through variability (including cessation) of the $\sigma_\theta = 26.6$ kg m\textsuperscript{-3} outcrop, travel eastward along the subtropical-subpolar gyre boundary, and enter the northern portion of the subtropical gyre along the way. For the 152\textdegree W AOU data within the gyre boundary region (40\textdegree–45\textdegree N), good agreement exists with the close-by time series data from Ocean Station P (50\textdegree N, 145\textdegree W) where a bidecadal cycle in AOU has been observed. In contrast, a sensible correlation with the Pacific Decadal Oscillation could not be found.


1. Introduction

[2] Among the objectives of the Climate Variability and Predictability and Carbon (CLIVAR/CO\textsubscript{2}) Repeat Hydrography Program (see http://ushydro.ucsd.edu) was to determine the large-scale variability of biogeochemical tracers as well as of pathways of ocean ventilation. Oxygen (O\textsubscript{2}) measurements are particularly useful in addressing these objectives because a large historical record of high-quality O\textsubscript{2} data exists that exceeds that of any other biogeochemical tracer and because of the close relation of oxygen concentrations to other biogeochemical properties through stoichiometric ratios [Redfield et al., 1963; Anderson and Sarmiento, 1994]. The purpose of this paper is to examine O\textsubscript{2} changes along the two North Pacific U.S. CLIVAR/CO\textsubscript{2} sections, P2 and P16N (Figure 1), conducted in 2004 and 2006, respectively, and to compare them with earlier observations of O\textsubscript{2} variability in the North Pacific Ocean.

[3] To account for variations in the O\textsubscript{2} equilibrium concentrations with the atmosphere (O\textsubscript{2,equi}), O\textsubscript{2} concentrations are often converted to apparent oxygen utilization (AOU), which is defined as AOU = O\textsubscript{2,equi}(θ, S) – O\textsubscript{2,meas} where O\textsubscript{2,meas} are the measured O\textsubscript{2} concentrations in the ocean and θ and S are potential temperature and salinity. In the

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subsurface ocean, below the mixed layer and the euphotic zone, AOU gives a measure of how much $O_2$ has been consumed because of respiration since a water parcel has left the surface layer. Temporal variations in AOU can occur in the subsurface ocean because of changes in respiration rates (biological effects) or changes in the age of a water parcel (physical effects) which changes the time the water parcel has been exposed to respiring organisms. Since temporal variations in $O_2$, equil are usually small when AOU variations are examined on isopycnal surfaces (where variations in $\theta$ and $S$ are comparatively small), temporal changes in AOU are about equivalent to $O_2$ changes of equal size but opposite in sign.

Several papers have examined temporal AOU (or $O_2$) variability in the North Pacific over the past several years. Investigations of repeated hydrographic sections [Watanabe et al., 2001; Emerson et al., 2001, 2004; Kumamoto et al. 2004] (see Figure 1) have shown that there has been an increase in AOU in the subpolar gyre and the northern subtropical gyre from the 1980s/early 1990s to the late 1990s/early 2000s. Modeling studies [Deutsch et al., 2005, 2006] as well as examination of chlorofluorocarbon (CFC) age data in conjunction with AOU data [Watanabe et al., 2001; Mecking et al., 2006] suggest that most of these changes are due to physical processes rather than changes in biology. Since the observed AOU differences consistently are the largest near the same density surface, $\sigma_\theta = 26.6$ kg m$^{-3}$, it has also been suggested that variations in the outcropping of this specific isopycnal (which is the densest isopycnal to outcrop in the open North Pacific in climatological data; see green line in Figure 1 for outcrop location) may be contributing to the observed AOU increase [Emerson et al., 2004; Mecking et al., 2006] in addition to a possible reduction in vertical mixing [Ono et al., 2001] and an increase in the inflow of Alaskan gyre waters into the western subarctic [Andreev and Kusakabe, 2001].

The late winter outcrop of $\sigma_\theta = 26.6$ kg m$^{-3}$ (green line, annually averaged acceleration potential [Montgomery, 1937; Reid, 1965] contours on $\sigma_\theta = 26.6$ kg m$^{-3}$ relative to 1500 m (streamlines; blue lines), and annually averaged apparent oxygen utilization (AOU) on $\sigma_\theta = 26.6$ kg m$^{-3}$ (color shading) are also shown. The grid spacing of the map projection is $15^\circ$ and $30^\circ$ of latitude and longitude, respectively. The outcrop, streamline, and AOU maps were calculated using the World Ocean Atlas 1998 climatology [Antonov et al., 1998; Bayer et al., 1998; O'Brien et al., 1998]. Acceleration potential and AOU data to the west/northwest of the $\sigma_\theta = 26.6$ kg m$^{-3}$ outcrop in late winter are omitted.

**Figure 1.** Location of CLIVAR/CO$_2$ Repeat Hydrography cruises in the North Pacific (black circles) together with previous observations of AOU (or $O_2$) changes based on reoccupations of hydrographic cruises (small gray symbols) and on compilations of longer-term time series records (large gray symbols). The late winter outcrop of $\sigma_\theta = 26.6$ kg m$^{-3}$ (green line), annually averaged acceleration potential [Montgomery, 1937; Reid, 1965] contours on $\sigma_\theta = 26.6$ kg m$^{-3}$ relative to 1500 m (streamlines; blue lines), and annually averaged apparent oxygen utilization (AOU) on $\sigma_\theta = 26.6$ kg m$^{-3}$ (color shading) are also shown. The grid spacing of the map projection is $15^\circ$ and $30^\circ$ of latitude and longitude, respectively. The outcrop, streamline, and AOU maps were calculated using the World Ocean Atlas 1998 climatology [Antonov et al., 1998; Bayer et al., 1998; O'Brien et al., 1998]. Acceleration potential and AOU data to the west/northwest of the $\sigma_\theta = 26.6$ kg m$^{-3}$ outcrop in late winter are omitted.
sections to earlier data and that there is an indication that the AOU signal has started to reverse from an AOU increase to a AOU decrease in recent years. This change in sign resembles the bidecadal cycles in AOU observed in longer-term time series data in the subarctic North Pacific [Andreev and Kusakabe, 2001; Ono et al., 2001; Kumamoto et al., 2004; Whitney et al., 2007] (see Figure 1), although the forcing mechanisms for these cycles still need to be explored further.

2. Data and Methods

[6] Two Repeat Hydrography cruises, P2 and P16N (data available at http://cchdo.ucsd.edu/pacific.htm), were conducted in the North Pacific as part of the U.S. CLIVAR/CO$_2$ program (Figure 1). CLIVAR/CO$_2$ P2 (15 June to 27 August 2004) repeated the World Ocean Circulation Experiment (WOCE) section P2 along a nominal latitude of 30°N, which is at about the center of the subtropical gyre as indicated by climatological streamlines (blue contour lines in Figure 1). WOCE P2 consisted of four different cruises conducted by Japanese scientists of which WOCE P2T (7 January to 10 February 1994) was designed to obtain nutrient and chemical data in conjunction with conductivity-temperature-depth data at 59 stations covering the entire width of the Pacific Ocean. The WOCE P2T data are the ones used here for comparison with CLIVAR/CO$_2$ P2. Since for WOCE P2T only the measured sum of nitrate and nitrite concentrations is reported to the CLIVAR and Carbon Hydrographic Data Office (CCHDO), nitrate concentrations in the subsurface ocean are approximated by the sum of nitrate and nitrite for this cruise. This is justified by the fact that nitrite concentrations, away from denitrification regions, are usually negligible below a nitrite spike at 75–125 m [Vaccaro, 1965].

[7] The northern leg of CLIVAR/CO$_2$ P16N (10–30 March 2006), which we focus on in this paper, followed a nominal longitude of 152°W in the eastern North Pacific between Hawaii (17°–23°N) and Kodiak (56°N) and crossed through the North Pacific subtropical and subpolar gyres (Figure 1). The streamlines on $\sigma_0 = 26.6$ kg m$^{-3}$ in Figure 1 indicate that at 152°W the boundary region between the subtropical and subpolar gyres extends from 40° to 50°N with its center at about 45°N. This boundary...
area, as shown in Figure 1, also corresponds to a meridional minimum in AOU because water with low AOU concentrations is carried from the isopycnal outcrops into the ocean interior along the gyre boundary and AOU increases as the water moves farther along streamlines into the ocean gyres. Data from four other cruises along the 152°W meridian (Fiona, 11–29 August 1980 (nominally 155°W); Marathon II, 5 May to 7 June 1984; WOCE P16N, 7 March to 8 April 1991; and STUD97, 1–21 November 1997) will be used for comparison with CLIVAR/CO2 P16N to the north of Hawaii, whereas Fiona and STUD97 only covered the subtropical portion of the section to the south of 44° and 45°N, respectively. AOU differences between these four earlier sections have been analyzed in the past [Emerson et al., 2001, 2004; Mecking et al., 2006]. Comparison on isopycnal surfaces at depths of ~2000 m (the deepest depth occupied on STUD97) showed that systematic offsets in oxygen concentrations between these cruises are at most 1–2 μmol kg⁻¹ [Emerson et al., 2001]. A similar comparison of the CLIVAR/CO2 P2 and P16N sections with the corresponding WOCE sections at depths ≥2000 m indicates that the deep oxygen data between the two programs agree within 1 μmol kg⁻¹ or less.

In order to compare AOU data between cruises, the AOU data are first objectively mapped on vertical sections for each cruise following procedures by Roemmich [1983]. To eliminate biases from the up and down movement of isopycnals, the mapping is done in density space using vertical correlation scales of 0.4 kg m⁻³ for σθ ≤ 26.0 kg m⁻³ and 0.3 kg m⁻³ for σθ > 26.0 kg m⁻³. Horizontal correlation scales are set to 2° of latitude for the meridional sections along 152°W. Larger horizontal correlation scales (4° of longitude) are used for the mapping of the zonal sections along 30°N because zonal gradients in AOU in the North Pacific Ocean (on isopycnals) are much smaller than meridional AOU gradients which implies a larger horizontal correlation. AOU differences are calculated by subtracting the maps for the earlier cruises from the maps for the CLIVAR/CO2 sections. The resulting differences are then projected onto the depth of the isopycnals (also objectively mapped and averaged between the two cruises involved) so that the AOU difference maps can be viewed in depth space (Figures 2 and 3) as illustrated, for instance, by Mecking et al. [2006, Figures 5 and 9]. Areas where there are insufficient data for either cruise used for the differentiation are marked by gray shading. Most prominently, there is a data gap for WOCE P16N between 48°20′N and 52°30′N that was caused by bad weather during the cruise. This produces the gray vertical bar around 50°N in the CLIVAR/CO2 P16N minus WOCE P16N difference map (Figure 2c). In addition, because difference maps are calculated in density space, large gray areas occur at depths shallower than the climatological winter mixed layer (dashed green lines in Figures 2 and 3) if an isopycnal exists during one cruise but not the other because of the occupation of the cruises in different seasons of the year.

Differences are also not calculated where cruise tracks deviate more than 3° from the CLIVAR/CO2 sections. Such a deviation mainly occurs because the 152°W sections vary in their approach to Hawaii so that the AOU comparisons with CLIVAR/CO2 P16N are terminated at 24°N for Fiona and Marathon II, at 21°N for WOCE P16N, and at 23.5°N for STUD97 (Figure 2). Also, while the nominal longitude of the Fiona cruise is technically to the west of the other 152°W cruises (by 3°), zonal AOU gradients within the broad region of the subtropical gyre are small and much less than temporal variations observed in AOU [Emerson et al., 2001]. Hence, we treat the Fiona cruise for the purpose of this paper as being one of the “152°W sections.” For WOCE P2T, the bottle oxygen data reported to the CCHDO are flagged as questionable or bad to the west of 147.5°W.

3. Observed AOU Changes

The AOU difference maps (calculated relative to CLIVAR/CO2) show that significant changes in AOU have occurred for both the 152°W and the 30°N sections. As in the earlier North Pacific repeat section analyses [Watanabe et al., 2001; Emerson et al., 2001, 2004], the differences are the most prominent at or near σθ = 26.6 kg m⁻³ (purple...
contours in Figures 2 and 3). From 1980 (Fiona) to 2006 (CLIVAR/CO2 P16N), the AOU change along 152°W corresponds mostly to an increase with maximum values \(>35 \, \mu\text{mol kg}^{-1}\) just above \(\sigma_\theta = 26.6 \, \text{kg m}^{-3}\) at \(39^\circ\)N (Figure 2a). Differences within and at the base of the climatological winter mixed layer (green dashed line) can be even larger, but since they are likely due to seasonal effects, they are not considered here.

[11] From 1984 (Marathon II) to 2006 (CLIVAR/CO2 P16N), AOU along 152°W has also mostly increased (Figure 2b) except for the vertical band of AOU decrease at \(25^\circ\)N that probably results from shifts within the zonal current system to the north of Hawaii [Emerson et al., 2001] and some reductions of AOU above the base of the winter mixed layer. Similar to the 1984 to 1997 comparisons by Emerson et al. [2001], the AOU increase relative to Marathon II (from 1984 to 2006) extends over a wider density range than for the Fiona (1980) section (Figure 2a) and also the WOCE P16N (1991) section (Figure 2c). This indicates that shallower parts of the permanent thermocline may also experience significant interannual AOU variations. Nevertheless, there is also a clear AOU difference between 1984 and 2006 that is associated with \(\sigma_\theta = 26.6 \, \text{kg m}^{-3}\) (Figure 2b). At \(43^\circ-47^\circ\)N, which is at the boundary between the subtropical and subpolar gyres, an increase in AOU that exceeds \(20 \, \mu\text{mol kg}^{-1}\) occurs at \(\sigma_\theta = 26.6 \, \text{kg m}^{-3}\) and slightly below. This signal extends (with interruptions) to the south and to the north into the subtropical and subpolar gyres, respectively.

[12] From 1991 (WOCE P16N) to 2006 (CLIVAR/CO2 P16N), changes in AOU along 152°W are still mostly positive (i.e., an AOU increase with time) and still exceed values of \(20 \, \mu\text{mol kg}^{-1}\) (Figure 2c), meaning that they are of similar magnitude as in the Fiona and Marathon II comparisons with CLIVAR/CO2 P16N. The AOU increase is clearly centered around \(\sigma_\theta = 26.6 \, \text{kg m}^{-3}\), is the largest between \(37^\circ\) and \(47^\circ\)N, and extends from there southward into the subtropical regions where AOU differences are smaller than for Fiona and Marathon II. A hint of AOU increase also continues to be found in the subpolar gyre to the north of the WOCE P16N data gap.

[13] A reversal of the AOU signal is finally apparent from 1997 (STUD97) to 2006 (CLIVAR/CO2 P16N). Between \(40^\circ\) and \(45^\circ\)N, the northern end of the STUD97 section, AOU on \(\sigma_\theta = 26.6 \, \text{kg m}^{-3}\) has decreased by as much as \(>10 \, \mu\text{mol kg}^{-1}\). Farther to the south, there is still an increase in AOU on \(\sigma_\theta = 26.6 \, \text{kg m}^{-3}\) that is of the same order of magnitude, whereas on shallower isopycnals south of \(35^\circ\)N, AOU has also decreased by a small amount. As discussed in section 2, the boundary between the subpolar and subtropical gyres at 152°W is centered at about \(45^\circ\)N. The onset of the AOU decrease on \(\sigma_\theta = 26.6 \, \text{kg m}^{-3}\) within a few degrees of the latitude of this location is consistent with circulation patterns since the gyre boundary region is directly connected to the isopycnal outcrop (green line in Figure 1) through strong eastward flow within the North Pacific Current (see blue streamline contours in Figure 1). Any ventilation anomaly carried in from the isopycnal outcrop region would hence first arrive at 152°W near the gyre boundary. A time lag is then expected for the anomaly to reach the interior of the subpolar and the subtropical gyres.

[14] At 30°N, AOU differences between 1994 (WOCE P2) and 2004 (CLIVAR/CO2 P2) show a maximum increase on \(\sigma_\theta = 26.6 \, \text{kg m}^{-3}\) to the east of \(\sim 160^\circ\)W and are more variable in the western part of the section (Figure 3). Decreases in AOU occur at shallower densities and depths (100–400 m), similar to the ones observed south of \(35^\circ\)N at 152°W between 1997 and 2006 (Figure 2d). In contrast to the increase in AOU in the eastern part of the 30°N section (Figure 3), no significant changes in AOU were found on \(\sigma_\theta = 26.6 \, \text{kg m}^{-3}\) in a zonal repeat section comparison that was done farther south at 24°N [Mecking et al., 2006] (see Figure 1 for location of this section). Together, the 30° and 24°N sections suggest that the AOU signal on \(\sigma_\theta = 26.6 \, \text{kg m}^{-3}\) that is carried in along the subtropical-subpolar gyre boundary, as seen in the 152°W sections (see above), is confined to the eastern part of the subtropical gyre and reaches 30°N but not 24°N. The meridional AOU difference sections along 152°W support this conclusion since the AOU anomalies on \(\sigma_\theta = 26.6 \, \text{kg m}^{-3}\) coming from the north do not extend southward of 25°–30°N (Figure 2). This distribution pattern of the AOU anomalies is also found in modeling studies [Deutsch et al., 2005, 2006] which show
two pathways of oxygen variability in the North Pacific: one that originates from decadal variability in ventilation rates in the northwestern Pacific and extends eastward into the northern portion of the subtropical gyre and another that results from decadal variability in the position of the southeastern boundary of the subtropical gyre with the separation between the pathways occurring at 25°–30°N [see also Sabine and Gruber, 2006]. The former, northern one of these pathways is the one that we are mostly concerned with in this paper.

4. Comparison With Salinity, Nitrate, and Other Biogeochemical Properties

Comparison of AOU with salinity on $\sigma_\theta = 26.6$ kg m$^{-3}$ shows that the large AOU increases observed along 152°W and 30°N between the 1980s/early 1990s and the late 1990s/2000s are not compensated by a change in salinity (Figures 4a and 5a). Along 152°W, this is indicated by AOU values being larger, at a fixed salinity, during STUD97 (1997) and CLIVAR/CO$$_2$$ P16N (2006) compared to the 1980s/early 1990s cruises for $S < 34.04$ (Figure 4a), which corresponds to the region to the north of ~25°N. Along 30°N, AOU values during CLIVAR/CO$$$_2$$ P2 (2004) are larger than those in the early 1990s (WOCE P2) at almost all salinities (Figure 5a). Along this section there have also been substantial changes in salinity [Kouketsu et al., 2007], as shown by plots of salinity versus longitude on $\sigma_\theta = 26.6$ kg m$^{-3}$ (Figure 6a). A freshening occurred from 1994 to 2004 to the west of about 170°W, whereas an increase in salinity from 1994 to 2004 is evident within the lateral salinity minimum at 130°–140°W (Figure 6a). The corresponding AOU changes, on the other hand, are most pronounced between these two regions at 140°–170°W (Figure 6b) where salinity has remained approximately constant (Figure 6a). This decoupling of the AOU variations from variability in salinity in the 30°N sections as well as in the 152°W sections (>25°N) reconfirms that the AOU changes in the northern portion of the subtropical gyre are larger than could be explained by a simple shift in gyre position [Emerson et al., 2001; Mecking et al., 2006].

Plots of AOU versus nitrate on $\sigma_\theta = 26.6$ kg m$^{-3}$ at 152°W (Figure 4b) and 30°N (Figure 5b) show, in contrast to the AOU versus salinity plots (Figures 4a and 5a), that the approximately linear relationship between AOU and nitrate has remained constant over time. This suggests, as observed by Emerson et al. [2001], that nitrate changes are linked to AOU changes in stoichiometric proportions [Redfield et al., 1963; Anderson and Sarmiento, 1994]. On the basis of comparison of deep values, measurements of other nutrients, namely, phosphate and silicate, appear to be less consistent (within their signal-to-noise ratios) among cruises than the oxygen and also the nitrate measurements. But temporal variations in phosphate and silicate concentrations (not shown), as for nitrate, also occur concurrently with the AOU changes on $\sigma_\theta = 26.6$ kg m$^{-3}$ (i.e., high/low AOU concentrations are correlated with high/low nutrients). In the case of phosphate and nitrate, a correlation with AOU is expected because these nutrients are produced and oxygen is consumed during the remineralization of organic matter and waters with high AOU (low oxygen) concentrations usually have high nitrate and phosphate concentrations. In the case of silicate, a connection to AOU is less obvious because the dissolution of silicon-containing shells exported from the surface ocean does not involve the consumption of dissolved oxygen. However, since silicate is regenerated in the subsurface ocean just as the other nutrients (nitrate, phosphate) and the North Pacific Ocean contains very active silicate cycling [Sarmiento et al., 2004], it is plausible that changes in ventilation processes that are the likely cause of the temporal variability in AOU [Deutsch et al., 2005, 2006] (see also section 5), nitrate, and phosphate observed in the North Pacific Ocean may cause similar variations in silicate.

Dissolved inorganic carbon (DIC) and pH are also linked to AOU during organic matter remineralization, but these properties are also affected by the input of anthropogenic carbon into the ocean. Observed changes in DIC and pH (not shown) between the WOCE and CLIVAR/CO$$$_2$$ P2 and P16N sections (the other 152°W data sets do not contain these carbon parameters) exceed those that can be explained by the AOU changes alone. In this case, the AOU changes can be used to divide temporal variations in DIC.
and pH into an AOU-related ventilation component and into an anthropogenic component [Sabine et al., 2008; R. Byrne et al., manuscript in preparation, 2008].

5. Discussion

While interannual AOU variability also occurs on shallower density surfaces, \( \sigma = 26.6 \text{ kg m}^{-3} \) stands out as the isopycnal with the most persistent AOU changes which occur on decadal time scales and which have been measured along 152°W over a period of 26 years (1980–2006). This isopycnal lies between the potential vorticity minimum associated with North Pacific Central Mode Water (\( \sigma = 26.0–26.5 \text{ kg m}^{-3} \)) and the North Pacific Intermediate Water salinity minimum (\( \sigma = 26.8 \text{ kg m}^{-3} \)) but does not correspond to a distinct thermocline water mass in itself. A time series of AOU anomaly on \( \sigma = 26.6 \text{ kg m}^{-3} \), constructed by averaging the gridded AOU data for each 152°W cruise between 40° and 45°N (44°N for the Fiona cruise) and then subtracting the mean of the cruises, summarizes the evolution of AOU in the subtropical-subpolar gyre boundary region, as discussed in section 3: lowest AOU values in the 1980s and early 1990s, highest AOU values in 1997 followed by above average AOU values in 2006 (Figure 7a, red symbols).

5.1. Comparison With Ocean Station P

Multidecadal time series of AOU in the subpolar regions suggest that the AOU variations may occur on a \( \sim 20 \)-year cycle that is superimposed on a small increasing trend in AOU [Andreev and Kusakabe, 2001; Ono et al., 2001; Kumamoto et al., 2004; Whitney et al., 2007]. Of these time series, Ocean Station P at 50°N, 145°W, where oxygen data have been collected since 1956 [Whitney et al., 2007], lies in close proximity to our 152°W study region.
5.3. Changes in Biological Pump and Water Mass Distributions as Potential Forcing Mechanisms

Other possible mechanisms for causing AOU changes that have been discussed in the literature [e.g., Emerson et al., 2001, 2004] include changes in the biological pump and changes in isopycnal depths as well as water mass variability caused, e.g., by changes in the positions of the subtropical and subpolar gyres. The latter mechanism was ruled out as a dominant mechanism because there have been substantial changes in the AOU versus salinity relationships (Figures 4a and 5a) that would not be expected in the case of a simple shift in gyre positions (see section 4).

In the case of the biological pump, one would expect an increase in AOU if carbon export from the surface ocean and, subsequently, remineralization of organic matter in the subsurface ocean were to be enhanced or if the stoichiometric ratios of oxygen consumption to nutrient and inorganic carbon production during remineralization [Anderson and Sarmiento, 1994] were to increase. However, since remineralization occurs throughout the water column (with an approximate exponential decrease in remineralization rates with depth) [Martin et al., 1987], it seems unlikely that changes in the biological pump, like changes in wind forcing, would cause AOU changes that clearly follow one isopycnal surface \( \sigma_\theta = 26.6 \text{ kg m}^{-3} \) as observed in the North Pacific. Also, the constancy of the AOU to nitrate ratios (Figures 4b and 5b) suggests that the stoichiometric ratios have not substantially changed. Additional support for the biological pump not being the main source of the observed AOU variability stems from changes in CFC ages that have been found to occur concurrently with AOU changes during earlier repeat section analyses [Watanabe et al., 2001; Mecking et al., 2006] and from modeling studies [Deutsch et al., 2005, 2006], all of which point toward the dominance of physical processes. For the new CLIVAR/CO2 data, the calculation of ventilation ages from CFCs has become more difficult because of the slowing/halting of atmospheric CFC increases since the 1990s. Hence, the comparison of ventilation ages from CLIVAR/CO2 P16N and P2 with earlier data will be subject of more extensive future analyses of CFC ages.

5.4 Decadal Oscillation as Potential Forcing Mechanism

A potential forcing mechanism for the observed decadal-scale AOU variations is the Pacific Decadal Oscillation (PDO) which is defined as the leading component of North Pacific sea surface temperature (SST) variability [Mantua et al., 1997] and is also related to changes in the wind forcing of the oceanic gyre circulation [Deser et al., 1999]. However, the peak-to-peak period of the (filtered) PDO index (Figure 7b) which is on the order of 40–60 years is much greater than that of the bidecadal AOU cycles at Ocean Station P (Figure 7a). A connection between our AOU anomaly time series at 40°–45°N, 152°W (also marked as red symbols in Figure 7b) and the PDO is also questionable. Statistically, a significant correlation \((r > 0.9)\) exists at a lag of \(\sim 15\) years. But this would imply that the low AOU values at 40°–45°N, 152°W in the 1980s/early 1990s are the ocean’s lagged response to the negative PDO values that occur before the well-documented 1976–1977 climate shift [e.g., Mantua et al., 1997] and that the high AOU values in the late 1990s/2000s correspond to the peak in the AOU anomaly at Ocean Station P data in 2000–2001. But this would imply that the PDO is positive as well. Hence, an inverse correlation between the PDO and the AOU anomalies is not conclusive either. It is also unclear why changes in wind forcing would affect one isopycnal, \( \sigma_\theta = 26.6 \text{ kg m}^{-3} \), more than any other.
because the isopycnal surface is moved into an area of larger organic matter degradation [Emerson et al., 2004]. Examination of the depth of $\sigma_\theta = 26.6 \text{ kg m}^{-3}$ in the 152°W and 30°N sections (Figure 8) provides an inconclusive picture: At 152°W between 40° and 45°N, where the largest AOU variations on $\sigma_\theta = 26.6 \text{ kg m}^{-3}$ are observed (Figures 2 and 7), the isopycnal is the deepest in 2006 (CLIVAR/CO2 P16N) and the shallowest in 1984 (Marathon II), as shown in Figure 8a. AOU values, on the other hand, are the largest in 1997 and at low levels for all the 1980s/early 1990s cruises. Farther to the south, between 24° and 27°N, a distinct difference in isopycnal depth is also present with the depth of $\sigma_\theta = 26.6 \text{ kg m}^{-3}$ being about 50–75 m deeper in both 1984 and 2006 than during the other cruises. In contrast, the largest difference in AOU at this location is between the 1984 (Marathon II) and 2006 (CLIVAR/CO2 P16N) cruises (Figure 2b). At 30°N, $\sigma_\theta = 26.6 \text{ kg m}^{-3}$ is deeper in 2004 (CLIVAR/CO2 P2) than in 1994 (WOCE P2) between 150° and 170°W by as much as 50 m (Figure 8b). This region overlaps with the area of largest AOU increase (Figures 3 and 6b) but does not cover the entire area.

5.4. Changes in Ventilation as Potential Forcing Mechanism

[25] Given the shortcomings of the mechanisms discussed in sections 5.2 and 5.3, we return to the suggestion that changes in ventilation are responsible for the observed decadal-scale AOU variations (as implicitly assumed in sections 3 and 4). More specifically, it has been proposed that periodic changes in the outcrop position of $\sigma_\theta = 26.6 \text{ kg m}^{-3}$ including a complete cessation of the outcropping of this isopycnal (since it is the densest one to outcrop in the open North Pacific in climatological data) may play a more important role than circulation changes for producing AOU variations in the North Pacific Ocean [Emerson et al., 2004; Mecking et al., 2006]. The consistency with which AOU anomalies in the North Pacific have been found on $\sigma_\theta = 26.6 \text{ kg m}^{-3}$, as pointed out by Emerson et al. [2004] and reconfirmed by the CLIVAR/CO2 data presented in this paper, indicate that there must be something distinctly different about this isopycnal compared to others. The fact that $\sigma_\theta = 26.6 \text{ kg m}^{-3}$ is the densest isopycnal to outcrop in the open North Pacific in late winter (see Figure 1) provides an explanation why this isopycnal would be affected the most by the periodic ventilation mechanism.

[26] The idea is that oxygen concentrations are reset to their equilibrium concentrations when the 26.6 kg m$^{-3}$ isopycnal outcrops in the northwestern Pacific. In contrast, if $\sigma_\theta = 26.6 \text{ kg m}^{-3}$ stops outcropping, perhaps for several years in a row, old waters with high AOU values (and high CFC ages) [Mecking et al., 2006] are recirculated within the subtropical and subpolar gyres without this resetting of the boundary condition. As a result, large temporal AOU differences occur in the northwestern Pacific that are then carried eastward with the North Pacific Current and southward with the subtropical gyre circulation toward the 152°W and 30°N sections discussed in this paper. During this part of the process, the influence of recirculated waters is obviously increased, consistent with the notion that a greater influence of older Alaskan gyre waters may be the reason for the periodic increases in AOU concentrations observed in the western subarctic gyre [Andreev and Kusakabe, 2001] (see Figure 1 for Andreev and Kusakabe’s study location). As indicated by the constant AOU to nitrate relationships (Figures 4b and 5b), nutrients and other biogeochemical properties (see section 4) appear to be affected similarly to AOU by the periodic ventilation of the 26.6 kg m$^{-3}$ isopycnal.

[27] Most recently, Andreev and Baturina [2006] and Whitney et al. [2007] suggested that the 18.6-year nodal tidal cycle due to lunar orbital fluctuations that may substantially alter mixing in the Kuril Straits and modulate the PDO [Yasuda et al., 2006] could be the source of the oxygen (and AOU) cycles observed at Ocean Station P because it has a similar period. Future oxygen observations in the North Pacific will be needed to confirm the connection between ventilation events and the tidal forcing cycle [Whitney et al., 2007]. Since variations in the tidal mixing in the Kuril Straits affects sea surface density in the northwestern Pacific [Yasuda et al., 2006], this mechanism would be consistent with the idea that variations in the outcrop of $\sigma_\theta = 26.6 \text{ kg m}^{-3}$ (see above) are causing the AOU variability observed in the ocean interior.

[28] Unfortunately, measurements of sea surface density in the northwestern North Pacific remain sparse, particularly in winter. Ono et al. [2001] constructed a time series of
wintertime sea surface density in the Oyashio region off Japan which shows decadal variations in density, but data farther to the north and northeast would be needed to fully explore interannual variability in the $\sigma_\theta = 26.6$ kg m$^{-3}$ isopycnal as well as its forcing mechanisms. Figure 1 location of the outcrop and of the study region by Ono et al.). Argo floats do not cover the north of 42°N in analyses of mixed layer depth and density from Argo data [Ohno et al., 2004; Whitney et al., 2007]. However, measurement of ocean salinity through the launch of the Aquarius satellite within the next few years [Lagerloef, 2004] in combination with data from already orbiting SST satellites will provide the opportunity to better monitor sea surface density. This will be particularly useful in remote locations such as the northwestern North Pacific and will help to better identify the sources of AOU variability in the ocean interior.

6. Conclusions

[29] The CLIVAR/CO$_2$ measurements in the North Pacific reveal AOU changes that remain predominant on $\sigma_\theta = 26.6$ kg m$^{-3}$ and include the onset of a reversal in the AOU signal between 1997 and 2006 from AOU increase to decrease. Different from other oceans where decadal variations in AOU have been observed within different kinds of mode waters [McDonagh et al., 2005; Johnson and Gruber, 2007], the changes in the North Pacific are not confined to a particular water mass. Instead, they occur on or close to the isopycnal that is the densest to outcrop in the open North Pacific according to climatological data ($\sigma_\theta = 26.6$ kg m$^{-3}$). It is suggested that periodic changes in the outcrop location of this isopycnal (including complete cessation of outcropping) may be the reason why the decadal-scale AOU changes are so closely confined to this isopycnal. Long-term monitoring of wintertime sea surface density in the northwestern North Pacific will be required to fully explore interannual and decadal variability in the ventilation of this isopycnal as well as its forcing mechanisms.

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