The North Pacific Oxygen Uptake Rates over the Past Half Century

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ABSTRACT

The transport of dissolved oxygen (O₂) from the surface ocean into the interior is a critical process sustaining aerobic life in mesopelagic ecosystems, but its rates and sensitivity to climate variations are poorly understood. Using a circulation model constrained to historical variability by assimilation of observations, the study shows that the North Pacific thermocline effectively takes up O₂ primarily by expanding the area through which O₂-rich mixed layer water is detrained into the thermocline. The outcrop area during the critical winter season varies in concert with the Pacific decadal oscillation (PDO). When the central North Pacific Ocean is in a cold phase, the winter outcrop window for the central mode water class (CMW; a neutral density range of \( \gamma = 25.6-26.6 \)) expands southward, allowing more O₂-rich surface water to enter the ocean’s interior. An increase in volume flux of water to the CMW density class is partly compensated by a reduced supply to the shallower densities of subtropical mode water (\( \gamma = 24.0-25.5 \)). The thermocline has become better oxygenated since the 1980s partly because of strong O₂ uptake. Positive O₂ anomalies appear first near the outcrop and subsequently downstream in the subtropical gyre. In contrast to the O₂ variations within the ventilated thermocline, observed O₂ in intermediate water (density range of \( \gamma = 26.7-27.2 \)) shows a declining trend over the past half century, a trend not explained by the open ocean water mass formation rate.

1. Introduction

Marine heterotrophs require O₂ for respiration, and O₂ variability constrains the habitat of many species over large volumes of the deep ocean (Deutsch et al. 2015; Vaquer-Sunyer and Duarte 2008). To maintain a habitable deep ocean, O₂ must be supplied via the downward transport of surface water rich in the O₂ obtained from the atmosphere and photosynthesis. An imbalance between physical input and biological consumption rates, both highly sensitive to climate, can change O₂ levels over time. Such an imbalance has been widely predicted to occur with climate warming, driven largely by reduced O₂ supply (Keeling et al. 2010). Over the past 50 years, O₂ has declined within the subpolar gyre, while it has slightly increased within the subtropics (e.g., Emerson et al. 2004; Deutsch et al. 2005; Stramma...
crop expansion and contraction represents an important into the thermocline (Marshall 1997; Kwon et al. 2013). Improved understanding of the supply mechanism and its relationship to climate variations will help us better understand observed O\textsubscript{2} changes.

As a result of strong stratification within the North Pacific, convection is restricted to relatively shallow depths (usually <250 m; e.g., Suga et al. 2004), and the ventilation of deeper layers requires the transfer of water from the surface mixed layer into the thermocline. The transfer is primarily achieved through subduction (Sallée et al. 2012; also see below). The subduction of O\textsubscript{2}-rich surface water occurs through both wind-driven Ekman downwelling (Huang and Qiu 1994) and the injection of winter mixed layer water beneath the shoaling mixed layer base (Stommel 1979; Williams et al. 1995). The subduction mainly occurs at the boundary between subtropical and subpolar gyres, where seasonal air–sea heat exchange and boundary layer mixing are strong (Oka et al. 2011). A recent study by Kwon et al. (2013) highlighted the importance of the winter outcrop area, where thermocline waters are exposed at the sea surface during winter, for determining the annual mean subduction of surface water to the thermocline. The seasonal detrainment of mixed layer water mainly occurs during early spring stratification (Woods and Barkmann 1986) through a large outcropping area created during winter cooling. In contrast, seasonal entrainment of thermocline water back to the mixed layer occurs gradually over a contracted outcrop window. Repeating seasonal cycles can lead to the net downward transport of water masses into the thermocline (Marshall 1997; Kwon et al. 2013).

In this paper, we show that the seasonal cycle of outcrop expansion and contraction represents an important mechanism by which the North Pacific takes up O\textsubscript{2} (section 2). By diagnosing the change in the O\textsubscript{2} transfer rate over the past half century, we show that the North Pacific’s O\textsubscript{2} uptake rate has changed in concert with the Pacific decadal oscillation (PDO; Mantua et al. 1997), primarily through its influence on the interannual variability of the winter outcrop area (section 3). Analysis of historical hydrographic data (section 4) reveals a potential link between the multidecadal fluctuation in the estimated O\textsubscript{2} supply and observed O\textsubscript{2} variability within the North Pacific ventilated gyres. While we mainly focus on mode water density ranges that are directly ventilated from the open ocean’s surface, substantial O\textsubscript{2} decline has been reported for density layers that do not outcrop at the open ocean surface (e.g., Ono et al. 2001; Whitney et al. 2007; Mecking et al. 2008). Thus, we briefly discuss our results in the context of reported deoxygenation (section 4c).

### 2. Diagnosing O\textsubscript{2} exchange rate

#### a. Background

Processes contributing to the exchange of O\textsubscript{2} between the surface and the interior ocean include the subduction and mixing of O\textsubscript{2}-rich surface water across the base of the mixed layer. The area-integrated subduction rate, \( M \), of O\textsubscript{2}-rich mixed layer water is diagnosed by employing the kinematic approach of Cushman-Roisin (1987) as follows:

\[
M(\gamma, t) = \int_{A(\gamma_1 \leq \gamma < \gamma_2)} [O_2]_h \left( -\frac{\partial h}{\partial t} - U_h \nabla_h h - w_h \right) dA,
\]

where \( A(\gamma_1 \leq \gamma < \gamma_2) \) is the area over which waters of density \( \gamma_1 \leq \gamma < \gamma_2 \) outcrop at the base of the mixed layer; \( h \) is the mixed layer thickness determined using a density difference of 0.03 kg m\textsuperscript{-3} from the sea surface (de Boyer Montégut et al. 2004); \( U_h \) is the lateral velocity of water at the base of the mixed layer; \( w_h \) is the vertical velocity of water at the base of the mixed layer; and \( \nabla_h \) is the horizontal gradient operator. A positive \( M \) indicates downward transfer and is referred to as subduction. A negative \( M \) indicates upward transfer, also known as obduction.

Combined with mixing effects, the exchange rate becomes

\[
E(\gamma, t) = \int_{A(\gamma_1 \leq \gamma < \gamma_2)} [O_2]_h \left( -\frac{\partial h}{\partial t} - U_h \nabla_h h - w_h \right) - K_t \cdot \nabla_h [O_2]_h - K_v \cdot \nabla_v [O_2]_h dA,
\]

where \( \nabla_v \) is the vertical gradient operator; and \( K_t \) and \( K_v \) are the horizontal and vertical mixing coefficients, respectively, fixed at upper bounds of \( K_t = 10^4 \text{m}^2\text{s}^{-1} \) and \( K_v = 10^{-4} \text{m}^2\text{s}^{-1} \) (Sallée et al. 2012). A lateral mixing component that is perpendicular to the base of the mixed layer is required. Hence, the lateral mixing term (i.e., \( K_t \cdot \nabla_h [O_2]_h \cdot \nabla_h \)) includes \( \nabla_h \).

Since we are concerned with the interannual variability of the O\textsubscript{2} exchange rate across the moving base of the mixed layer, we employ the instantaneous subduction rate of Cushman-Roisin (1987) rather than the time-invariant subduction rate into the main thermocline of Marshall et al. (1993). When averaged over several annual cycles, the time-averaged subduction rate of O\textsubscript{2} across the time-varying mixed layer base is not necessarily equal to the subduction rate across the time-invariant winter...
mixed layer, because diapycnal mixing within the seasonal thermocline can modulate the O$_2$ transfer rate into the main thermocline (Nurser et al. 1999; Marshall et al. 1999; Kwon et al. 2013). Therefore, the O$_2$ subduction rate obtained using Eq. (1) includes not only O$_2$ fluxes entering the permanent thermocline, but also O$_2$ fluxes subject to alteration by diapycnal mixing and remineralization within the seasonal thermocline.

b. Attributing the mechanisms of subduction

To elucidate the mechanism by which O$_2$-rich mixed layer water subducts, we decompose the net subduction rate into its constituent mechanisms using Eq. (1). The contributions are from lateral induction beneath the temporarily varying mixed layer base (i.e., $Dh/Dt = \partial h/\partial t + Uh \nabla h$; Williams et al. 1995; Kwon et al. 2013) and the vertical transport of water ($w_h$) dominated by Ekman pumping/suction using the following equation:

$$M(y,t) = \int_A [O_2]_h \frac{Dh}{Dt} dA + \int_A [O_2]_h w_h dA. \quad (3)$$

The first term on the right-hand side of Eq. (3) provides the downward transport of O$_2$-rich surface water across the base of the mixed layer following a water parcel. Rapid shoaling of the mixed layer base during early spring can make a significant contribution to this term (Kwon et al. 2013). The second term arises from vertical movements of water mainly induced by wind stress curl (i.e., Ekman upwelling/downwelling).

To determine the contribution of each term in Eq. (1) to interannual variability, we calculate the mean seasonal cycles for the outcrop area $\overline{A}$, the time rate of change in the mixed layer depth $\partial h/\partial t$, and the terms involving ocean circulation $Uh \nabla h$ and $\overline{w_h}$. Then we replace the corresponding terms in Eq. (1) with the mean seasonal cycles. Each of these three cases can be described mathematically thusly:

Constant area case: $M_A(y,t)$

$$= \int_{A(y_t \leq y \leq y_f)} [O_2]_h \left( \frac{\partial h}{\partial t} - Uh \nabla h - w_h \right) dA \quad (4)$$

Constant mixed layer case: $M_h(y,t)$

$$= \int_{A(y_t \leq y \leq y_f)} [O_2]_h \left( -\frac{\partial h}{\partial t} - Uh \nabla h + w_h \right) dA \quad (5)$$

Constant circulation case: $M_c(y,t)$

$$= \int_{A(y_t \leq y \leq y_f)} [O_2]_h \left( -\frac{\partial h}{\partial t} - Uh \nabla h - \overline{w_h} \right) dA, \quad (6)$$

where the overbar represents the mean seasonal cycle of each term, averaged over the time period from 1958 to 2007. Correlations between the two subduction estimates obtained from the fully varying case and the idealized cases are then used to attribute contributions from the varying outcrop area, mixed layer depths, and circulation. The portion of variance attributed to each component is computed as the residual variance that is not explained by the time series when that component is held constant. Although these processes are not independent of one another, the method helps us understand their relative importance in determining the temporal variation of the subduction rate.

c. The model and data

We use the Simple Ocean Data Assimilation version 2.1.6 (SODA; Carton et al. 2000a,b; Carton and Giese 2008; available at http://www.atmos.umd.edu/~ocean/) to estimate the subduction rate of North Pacific upper-water masses. SODA uses an ocean general circulation model based on Parallel Ocean Program numerics (Smith et al. 1992). Vertical mixing of momentum, heat, and salt is represented using K-profile parameterization (Large et al. 1994). The model is run at a horizontal resolution of 0.25° × 0.4°, and model solutions of temperature, salinity, and velocity are remapped onto uniform 0.5° × 0.5° horizontal grid points (Carton and Giese 2008). The vertical resolution ranges from 10 m near the surface to ~250 m near the bottom, with a total of 40 vertical levels. The SODA hindcast simulation spans the time period from 1958 to 2007. An optimal interpolation method is used to assimilate temperature and salinity profiles and satellite sea surface temperature into the numerical model (Carton and Giese 2008). The model is known to reasonably capture observed large-scale upper-ocean physical structure and circulation (Carton et al. 2000b; Carton and Giese 2008).

In Figs. 1b, 1d, and 2, we compare some of the climatological mean features relevant to O$_2$ subduction between the model and the World Ocean Atlas 2009 (WOA09; Locarnini et al. 2010; Antonov et al. 2010; Garcia et al. 2010). The surface distribution of winter (January–March) mean saturated O$_2$ concentration, estimated using temperatures and salinity, compares well between SODA and the WOA09 (Figs. 1b and 1d). Both the model and the observations indicate that saturated O$_2$ concentrations increase from ~200 μmol kg$^{-1}$ in the southern flank of the subtropical gyre to ~350 μmol kg$^{-1}$ in the western subpolar gyre. The distribution of observed surface O$_2$ concentration is similar to that for saturated O$_2$ concentration (Fig. 1a), with a slight (<10%) supersaturation occurring within the subtropical gyre and an undersaturation occurring within the subpolar gyre (Fig. 1c).
density is also well reproduced in the model, especially in the open ocean (Figs. 1c and 1d).

The winter mixed layer depth, determined using a density difference of 0.03 kg m\(^{-3}\) from the sea surface (de Boyer Montégut et al. 2004), is reasonably reproduced in the model with some notable discrepancies (Fig. 2). In both the model and observations, deep mixed layers of \(\sim 150\) m form in the northwest Pacific at the transition area between the subtropical and subpolar gyres. The band of the mixed layer maxima extends from the northwest Pacific off the east coast of Japan toward the central North Pacific, coincident with mode water formation regions (Hanawa and Talley 2001; Yasuda 2003; Suga et al. 2004). Along the midlatitude northwest Pacific between 30\(^\circ\) and 40\(^\circ\)N, excess heat loss to the atmosphere (Fig. 3a) associated with strong winds triggers deep convection within the mixed layer during winter. Despite the qualitative agreements, the simulated winter mixed layer depths in the northeast Pacific between 40\(^\circ\) and 50\(^\circ\)N are shallower than the observations. Another notable discrepancy is the model’s inability to capture the observed separation of two deep mixed layer bands, one located at the Oyashio Front near 40\(^\circ\)N and the other located at the Kuroshio Front near 30\(^\circ\)N. Instead, in the model, deep winter mixed layers form in a broad region of the northwest Pacific as shown in Fig. 2b.

Since we focus on the subduction rate integrated over a broad range of density layers (see below), the model’s inability to reproduce the finescale structure of the mixed layer depth likely does not undermine our conclusions.

In addition to climatological mean features, the SODA simulation is able to reasonably reproduce multidecadal climate variability. We compute the empirical orthogonal function (EOF) for deseasonalized near-surface temperature, averaged over the top 10 m. The leading EOF pattern (Fig. 4a) and the associated principal component time series (Fig. 4b) roughly agree with those of Mantua et al. (1997), who used independent datasets of sea surface temperature to define the
PDO index. Such agreement provides confidence in the use of the SODA simulation for the study of large-scale upper-ocean processes and their decadal variability in the North Pacific.

To estimate the subduction of O$_2$ that is maximally consistent with both the observed distribution of O$_2$ and with the large-scale circulation, we combine the physical quantities from SODA with climatological O$_2$ fields. This approach avoids the relatively large biases present in model simulations of the O$_2$ cycle, which are in any case not part of the SODA model. To combine these datasets, we first obtain the monthly mean O$_2$ concentration at the base of the mixed layer using $1^\circ \times 1^\circ$ WOA09 monthly climatologies for temperature, salinity, and O$_2$ (Locarnini et al. 2010; Antonov et al. 2010; Garcia et al. 2010). The O$_2$ concentration interpolated to the base of the mixed layer is very close to the surface value (Fig. 1a). The vertically interpolated WOA09 O$_2$ data are then linearly interpolated to 0.5$^\circ$ SODA grids. We use the interpolated monthly

![Image](https://example.com/image1.png)

**Fig. 2.** Winter mean mixed layer depth determined using a density difference of 0.03 kg m$^{-3}$ from the sea surface. (a) Estimate from WOA09. (b) Estimate from SODA averaged over 1958–2007. Winter mean surface neutral densities are shown as black solid lines.

![Image](https://example.com/image2.png)

**Fig. 3.** Winter mean air–sea heat exchange, computed using version 2 of the Common Ocean Reference Experiment global air–sea flux dataset (Large and Yeager 2009). (a) Winter mean heat fluxes averaged over 1958–2006 (W m$^{-2}$). Positive values indicate heat gain by the ocean, and negative values indicate heat loss to the atmosphere. Black contour lines are the winter mean surface neutral densities averaged over 1958–2006, obtained from SODA. (b) The regression pattern of the annual mean PDO index on the annually averaged winter air–sea heat exchange (W m$^{-2}$ per one standard deviation of the PDO index). Values significant at 95% are shown. Positive values represent anomalously increased heat inputs to the ocean (or decreased heat losses to the atmosphere) during positive PDO years, while negative values represent anomalously decreased heat inputs to the ocean (or increased heat losses to the atmosphere). Black (green) contour lines represent the winter neutral densities composited for years with the PDO index greater than one standard deviation (less than minus one standard deviation). The PDO index is obtained from Mantua et al. (1997) (available at http://jisao.washington.edu/pdo/PDO.latest).
mean O$_2$ fields repeatedly for the product of O$_2$ and the water mass subduction rate across the base of the mixed layer [as expressed in Eq. (1)] over the simulation period from 1958 to 2007. This approach neglects interannual variations in surface O$_2$. To evaluate the potential importance of surface O$_2$ variability, we also derive subduction rates assuming that surface O$_2$ is at equilibrium with the atmosphere, allowing its interannual variations to be computed from the SODA temperature and salinity. The two approaches do not make a discernible difference for interannual variations of O$_2$ subduction rate, because surface O$_2$ concentration variation plays a minor role in the interannual variability of the O$_2$ subduction (see section 3).

The exchange rate of O$_2$ is then integrated over the entire North Pacific north of 10$^\circ$N in mode water density classes using a neutral density interval of $\Delta \gamma = 0.1$. For example, in our notation, a density bin of $\gamma = 25.6$ indicates a density range of 25.55 ≤ $\gamma$ < 25.65. Lighter density bins ($\gamma = 24.0$–25.5) are referred to as the subtropical mode water (STMW) density class and denser density bins ($\gamma = 25.6$–26.6) are referred to as the central mode water (CMW) density class. The mode water density classes defined in this study are different from the classical definitions of “mode water,” which is generally referred to as a water of potential vorticity less than 2 × 10$^{-10}$ m$^{-1}$s$^{-1}$ (Nakamura 1996; Suga et al. 1997). Thus, our STMW density class encompasses the classical “STMW” formed in the western and eastern North Pacific (Hanawa and Talley 2001). On the other hand, the CMW density class mainly forms in the central North Pacific (Fig. 5b).

### 3. Seasonal to decadal variability in the O$_2$ supply rate

The time mean transfer rate of O$_2$ has two peaks in two distinct density ranges, one corresponding to the STMW density class ($\gamma = 24.0$–25.5) and the other corresponding to the CMW density class ($\gamma = 25.6$–26.6) (Hanawa and Talley 2001; Fig. 6a). Annual mean O$_2$ subduction (Figs. 5b and 5f) mainly results from the seasonal cycle of the mixed layer depth that occurs along with horizontal migrations of surface outcrops. Shoaling of the deep winter mixed layer detrains O$_2$-rich surface water to the underlying thermocline during early spring (Cushman-Roisin 1987). Most seasonal subduction occurs within the winter outcrop, when its areal extent reaches a seasonal maximum (i.e., when the outcrop expands toward the equator during late winter; Figs. 5c and 5g). On the other hand, seasonal obduction occurs within outcrops whose areal extent is relatively contracted in high latitudes during fall (Figs. 5d and 5h). The imbalance between seasonal subduction rates and seasonal obduction rates leads to net annual mean O$_2$ subduction, which is focused within late winter (or early spring) outcrop regions (Figs. 5b and 5f; Stommel 1979). For the CMW density class, in particular, the effect of the seasonal cycle [i.e., lateral induction beneath the shoaling mixed layer base; $\int_A \left[ -[O_2]_b \frac{DH}{Dt} \right] \, dA$ in Eq. (3)] offsets the O$_2$ obduction induced by Ekman upwelling [$\int_A -[O_2]_b \vec{w}_b \, dA$ in Eq. (3); Fig. 6a]. Similar to Sallée et al. (2012), we also find that mixing effects are an order of magnitude smaller than advective subduction (Fig. 6b). Thus, our focus is O$_2$ transport due to the subduction of water masses. Once O$_2$ subducts into

![Figure 4](image-url)
FIG. 5. The climatology of $O_2$ distribution and subduction on two representative isopycnal surfaces of the CMW and STMW density classes. (a) The March climatology of the $O_2$ distribution ($\mu$mol kg$^{-1}$) on a $\gamma = 26.0$ (=1026.0 kg m$^{-3}$) isopycnal surface, linearly interpolated from WOA09. The depth (m) of the isopycnal is overlain with black solid lines. White indicates the area where the isopycnal does not exist in March. (b) An annually averaged map of the $O_2$ subduction rate over a seasonally migrating outcrop of $\gamma = 26.0$ (mol kg$^{-1}$ m$^{-2}$), an estimate from SODA and $O_2$ climatology data. The area integral of the subduction rate over the North Pacific is shown inside the panel using a unit of Tmol yr$^{-1}$ (1 Tmol = $10^{12}$ moles). Equation (1) is integrated over 50 years for each SODA grid point in order to produce the map. Positive values represent subduction. Black contour lines represent the March outcrop of $\gamma = 26.0 \pm 0.05$ averaged from 1958 to 2007. (c) As in (b), but with the exception that the average from February to July (the subduction period) is shown. (d) As in (c), but with the exception that the average from August to January (the obduction period) is shown. (e)–(h) As in (a)–(d), but with the exception that maps for a $\gamma = 25.0$ isopycnal are shown.
the thermocline, O$_2$-rich surface water spreads southward along isopycnals, and O$_2$ decreases as water moves away from the outcrop area (Fig. 5a and 5e).

Annual mean O$_2$ transfer rates integrated over the STMW and CMW density classes exhibit considerable fluctuations on interannual to decadal time scales (Fig. 7a). The time mean and standard deviations are (60 ± 23) Tmol O$_2$ yr$^{-1}$ for the STMW density class and (100 ± 23) Tmol O$_2$ yr$^{-1}$ for the CMW density class. Partly because of large interannual variations, we do not find any significant linear trends for the O$_2$ transfer rate over the past 50 years. For example, the estimated trend of -0.33 Tmol O$_2$ yr$^{-2}$ for the STMW density class is smaller than the 95% confidence interval of ±0.54 Tmol O$_2$ yr$^{-2}$. Likewise, the estimated trend of 0.36 Tmol O$_2$ yr$^{-2}$ for the CMW density class is also smaller than the 95% confidence interval of ±0.39 Tmol O$_2$ yr$^{-2}$. The lack of any significant trends is still valid even if we consider the effect of varying solubility for O$_2$ over the past half century (Figs. 7b and 7c). The correlation coefficients between the two estimates are above 0.98 (Figs. 7b and 7c). The strong correlations result from the fact that interannual variability in the O$_2$ transfer rate is dominated by variability in the water mass subduction rate rather than the variability of surface O$_2$ concentrations.

Multidecadal variations in O$_2$ subduction rates are related to the PDO, a dominant mode of climate variability in the North Pacific. Previous studies have suggested that winter mixed layer convection and stratification in the central North Pacific strongly respond to interannual variations in heat fluxes and winds at the sea surface (Qiu and Joyce 1992; Deser et al. 1996; Yasuda and...
Variations in winter mixed layer depth are, in turn, accompanied by changes in the formation and subduction rates of CMW (Ladd and Thompson 2002; Qu and Chen 2009) and hence the supply rate of \( O_2 \).

Figure 7a indicates that multidecadal variation in the \( O_2 \) subduction rate of the CMW density class is anticorrelated with the subduction rate of the STMW density class. Such an antiphased relationship can largely be explained through variations in outcrop areas of the CMW and STMW classes (Fig. 8b). When the central North Pacific Ocean is anomalously cold, the winter outcrop area of the CMW density water expands farther southward, resulting in a contraction of the outcrop area of the STMW density water (Schneider et al. 1999; Ladd and Thompson 2002; Oka et al. 2012; see also Figs. 3b and 8b). The expansion of the winter outcrop area of the CMW density class is associated with enhanced wintertime heat loss to the atmosphere in the central North Pacific (Fig. 3b). During positive PDO years when the CMW density outcrop is anomalously located farther south, more permanent subduction occurs because the water is unlikely to be reentrained back to the mixed layer during subsequent warm years, where the outcrop area is in its more poleward configuration. For this reason, long-term mean subduction of the CMW density class occurs south of the climatological mean outcrop locations (Fig. 5b).

To diagnose the relative importance of changes in the outcrop area, mixed layer depths, and circulation, we recompute the \( O_2 \) subduction rate after eliminating the interannual variability in each of the physical drivers, as expressed in Eqs. (4)–(6). Interannual variations in the outcrop area explain 48% of the total variance of the \( O_2 \) supply rate to the CMW density class (Fig. 9a). Interannual variations in the mixed layer depth and circulation

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**Fig. 8.** (a) The annual mean PDO index estimated from SODA, the principal component time series presented in Fig. 4b. (b) March outcrop areas integrated over the CMW and STMW classes, estimates from SODA. The correlation coefficient estimated between the PDO index in (a) and the March outcrop area in (b) is \( r = -0.66 \) for the STMW density class (gray solid line) and \( r = 0.62 \) for the CMW density class (black solid line).
individually contribute to total variance by 43% and 23%, respectively (Figs. 9b and 9c). Interannual variations in the outcrop area also account for the highest fraction (77%) of total variance for the O2 subduction rate to the STMW density class (Fig. 9d), followed by contributions from changes in the mixed layer depth (56%) and circulation (18%; Figs. 9e and 9f). Relative contributions from the outcrop area, the mixed layer depth, and circulation do not necessarily sum to 100% because the three factors identified are dependent on each other. For example, excess heat losses during winter tend to enhance winter mixed layer convection and at the same time expand the winter outcrop area. The combined effect leads to increased water mass formation and subduction.

The winter outcrop area correlates strongly with the PDO index \( r = 0.62 \) for the CMW density class and \( r = -0.66 \) for the STMW density class, \( p < 0.01 \) for both; Figs. 8a and 8b). Since the interannual variation of the outcrop area is a primary cause of the interannual variability of the O2 subduction, the subduction is related to the PDO. In general, a cold phase in the central North Pacific favors an anomalously high subduction rate for O2-rich surface water to the CMW density class. An expanded winter outcrop area, together with intensified mixed layer convection (Ladd and Thompson 2002; Qu and Chen 2009), leads to a greater amount of O2 transferred from the mixed layer to the thermocline during the early spring in positive PDO years. On the other hand, the autumn outcrop area, over which seasonally subducted O2 is entrained back into the mixed layer, remains fairly constant on interannual to decadal time scales. Therefore, more O2-rich surface water can annually enter the thermocline through broader winter outcrop windows of the CMW density range during positive phases of the PDO. Such multidecadal fluctuations of the O2 transfer rate to the CMW density class tend to be partly offset by changes in the O2 transfer rate to the STMW density class, because the total outcrop area for both water masses is nearly conserved.
We note that significant correlations ($p < 0.05$) between the simulated PDO index (Fig. 8a) and O$_2$ subduction rates (Fig. 7a) are only obtained when O$_2$ subduction rates are smoothed using a low-pass filter. Moreover, the correlation becomes weaker than those between the outcrop area and the PDO index, because the rate of change in the mixed layer depth and ocean circulations, both are weakly correlated with the PDO index, also plays an important role in determining the interannual variability of the subduction rate (see above). Nevertheless, the fact that a low-frequency component of the O$_2$ subduction variability is related to the PDO is important since the thermocline circulation tends to integrate the effect of O$_2$ subduction on decadal time scales (Ito and Deutsch 2010). The decadal fluctuation in the O$_2$ uptake rates can influence multi-decadal O$_2$ variability within the ventilated thermocline.

4. Linking subduction to observed O$_2$ variability

In this section, we explore the potential link between the O$_2$ uptake rate described in the previous section and observed O$_2$ variability. To this end, we use a historical dataset of O$_2$ concentration and discuss decadal changes in O$_2$ distributions on neutral density surfaces of the thermocline.

a. Mapping of O$_2$ data

We use quality-controlled standard-level bottle data from the World Ocean Database 2009 (WOD; available at http://www.whoi.edu/science/PO/hydrobase/php/index.php; Curry and Nobre 2013; Johnson et al. 2009; Garcia et al. 2010). After linearly interpolating discrete O$_2$ profiles onto neutral density surfaces (Jackett and McDougall 1997), O$_2$ data for each density surface are mapped onto $2^\circ$ horizontal grids using Gauss–Markov mapping (Thomson and Emery 2014). In the Gauss–Markov smoothing method, the best estimate for each grid point is a linear weighted sum of neighboring observations. Weights are determined so that the meansquare error of the estimates can be minimized (Thomson and Emery 2014). For each density surface, all available data below the winter mixed layer depth are taken from a North Pacific domain of $20^\circ$–$60^\circ$N, $120^\circ$–$230^\circ$E, and binned into 10-yr intervals. Thus, between 1955 and 2004, five decadal mean values provide temporal changes in O$_2$ concentration. The mapping procedure is repeated for density surfaces of $\gamma = 25.6$–27.2, with a density interval of $\Delta \gamma = 0.1$. These density surfaces constitute the upper thermocline ventilated by subduction (section 3), as well as the lower thermocline that does not directly outcrop at the open ocean’s surface. Note that O$_2$ variability over the STMW density class ($\gamma = 24.0$–25.5) is not considered here because decadal variability is small in the STMW density range.

Further analyses of the objectively mapped O$_2$ data are conducted as follows. We flag grid points where the squared mapping error exceeds 50% of observed O$_2$ variance within the North Pacific domain for each density surface and for each decade. For the flagged grid points, the estimation is poorly constrained due to a lack of neighboring observations or the poor quality of observed data. To reduce uncertainties in our analyses, we only consider grid points that are not flagged in any of the five decades at each density surface. This criterion excludes 20%–30% of the mapped values in the central North Pacific between $170^\circ$ and $200^\circ$E, north of $36^\circ$N, and south of $28^\circ$N. We compute an O$_2$ anomaly relative to its climatological mean value for each grid point at each density surface (e.g., Figs. 10a–10e).

To estimate decadal changes in saturated O$_2$ and apparent oxygen utilization (AOU; defined as saturated O$_2$ concentration minus observed O$_2$ concentration), we use the objectively analyzed temperature and salinity data, averaged over each of the five decades between 1955 and 2004, and provided by the World Ocean Atlas 2013 (WOA13; Locarnini et al. 2013; Zweng et al. 2013; available at https://www.nodc.noaa.gov/OC5/woa13/). We regrid the $1/4^\circ \times 1/4^\circ$ WOA13 data into $2^\circ \times 2^\circ$ grids and compute saturated O$_2$. The WOA13-derived saturated O$_2$ data are combined with our O$_2$ estimate in order to compute AOU. The WOA13 data are also used to calculate the thickness of neutral density surfaces (Figs. 11a and 11c), which are needed to compute a volume-averaged O$_2$ anomaly. The use of temperature and salinity obtained from the same WOD bottle data (where the O$_2$ data originates) does not make a discernible difference in our conclusions. Because of more extensive observations incorporated (Locarnini et al. 2013; Zweng et al. 2013), here we present results obtained using WOA13 temperature and salinity climatologies.

b. O$_2$ change within the ventilated thermocline

The O$_2$ anomaly distribution in the CMW density range exhibits strong spatial and temporal variability (e.g., Figs. 10a–10e), reflecting complex factors determining O$_2$ changes within the thermocline (Deutsch et al. 2006). Despite the strong spatial dependency of O$_2$ variability, some features can be related to the O$_2$ subduction change of the CMW density. Near the winter outcrop, the thermocline tends to be better oxygenated during the 1975–84 interval (Fig. 10c), when the PDO index shifts from negative to positive. Near the winter outcrop of $\gamma = 26.0$ in the central North Pacific, O$_2$ anomalies are positive and up to $\sim 15\mu$mol kg$^{-1}$. 
(approximately 7% of the climatological mean), lying above the mapping errors of the corresponding grid points.

Positive O$_2$ anomalies near the winter outcrop of $\gamma = 26.0$ are not caused by increasing O$_2$ solubility, as can be inferred from negative AOU anomalies occurring at the same time and locations. Instead, negative AOU anomalies (Fig. 10f) suggest enhanced O$_2$ subduction rates during the 1975–84 interval when the winter outcrop area of the CMW density class reaches its maximum (Fig. 8b). Because of the southward expansion of the winter outcrop during positive PDO years (Fig. 3b), a greater area of the thermocline can be exposed to O$_2$-rich surface water. This would lead to the increased area in which O$_2$ supply rates exceed remineralization rates. In the area near the outcrop of $\gamma = 26.0$ between 30°–60°N and 120°–210°E, the averaged O$_2$ anomaly peaks during the 1975–84 interval (Fig. 10f).

The period of maximum O$_2$ content near the winter outcrop of $\gamma = 26.0$ is followed by a period of maximum O$_2$ content in downstream regions of the subtropical gyre. Positive O$_2$ anomalies appear in most of regions south of 30°N during 1985–2004. Overall enrichments in O$_2$ during 1985–2004 relative to the 1975–84 period can be explained either by downstream effects of positive O$_2$ anomalies near the outcrop area or reduced O$_2$ consumption rates due to remineralization of organic matter. Considering that the ventilation time scale of the thermocline ($\sigma_0 = 25.5–26.6$) ranges from 10 to 27 years (Sonnerup et al. 1999; Huang and Qiu 1994), a delayed downstream effect is a plausible cause. Furthermore, previous studies have suggested that North Pacific gyre circulation intensifies due to the multidecadal strengthening of the westerlies during positive PDO years (Qiu and Joyce 1992; Yasuda and Hanawa 1997; Miller et al. 1998; Deser et al. 1999; Taguchi et al. 2007). Gyre intensification would allow less O$_2$ to be consumed by remineralization due to a decreasing transit time from the surface (Deutsch et al. 2005, 2006).

The multidecadal trend discussed for $\gamma = 26.0$ extends throughout the entire CMW density layers ($\gamma = 25.6–26.6$), although the multidecadal change tends to be damped toward lighter density layers. In general, the
volume-averaged $O_2$ anomaly over the North Pacific domain from $20^\circ$–$60^\circ$N, $120^\circ$–$230^\circ$E displays a decreasing trend from the 1960s to the 1980s, followed by an increasing trend from the 1980s to the 2000s (Fig. 11b). In particular, the rebound from the 1980s to the 2000s is dominated by increasing $O_2$ within the subtropical gyre south of $30^\circ$N and is partly offset by decreasing $O_2$ within the Alaska gyre. The declining $O_2$ trend observed within the eastern subpolar gyre north of $50^\circ$N (Figs. 10d and 10e) may reflect changes in the subpolar gyre, discussed in the next section.

Reduced amplitudes of $O_2$ variability toward lighter densities (Fig. 11b) could arise from the compensating role of biological $O_2$ consumption on physically driven $O_2$ changes near the subtropical surface. Lighter isopycnals (e.g., $\gamma = 25.6$) outcrop near the subtropical gyre, where perennial surface nutrient depletion limits the export of organic matter and associated respiration at depth. Because of the close relationship between nutrient and $O_2$ cycles, changes in circulation cause $O_2$ supply and demand to increase at similar rates, stabilizing the overall $O_2$ content on density surfaces that outcrop near nutrient-limited subtropical gyres (Deutsch et al. 2006).

c. $O_2$ change in intermediate water

$O_2$ variability within the CMW density class contrasts with persistently declining trends in $O_2$ over the past 50 years, as reported for the North Pacific subpolar gyre with a density range of $\gamma = 26.7$–27.2 (e.g., Ono et al. 2001; Watanabe et al. 2003; Whitney et al. 2007; Whitney et al. 2013). The decreasing trend has been attributed
to a cessation of the winter outcrop for the isopycnal of $\sigma_o = 26.6$ in the latter decades of the twentieth century (Emerson et al. 2004), decreased water mass formation due to sea ice loss in the Sea of Okhotsk (Watanabe et al. 2003; Nakanowatari et al. 2007), stratification in the North Pacific subpolar gyre (Deutsch et al. 2005; 2006), and/or changes in subpolar gyre circulation and mixing (Andreev and Baturina 2006).

We define the density layer of $\gamma = 26.7$–27.2 (i.e., a density layer that does not outcrop at the open ocean surface) as the North Pacific Intermediate Water (NPIW) class, and we explore $O_2$ variability within the NPIW density layer. Our results confirm the declining trends in $O_2$ anomalies for $\gamma = 27.0$–27.2 when averaged over the North Pacific domain 20°–60° N, 120°–230° E (Fig. 11d). The decadal $O_2$ decline is most pronounced within the subarctic–subtropical gyre boundary (e.g., Ono et al. 2001; Emerson et al. 2004; Mecking et al. 2008), where the NPIW density water dominates (Fig. 11c). The NPIW density water forms as a mixture of the Kuroshio, Oyashio, and Tsugaru warm currents (e.g., Talley 1993), perhaps independent of direct ventilation from the surface (e.g., Qiu and Chen 2011; Yagi et al. 2014). Andreev and Baturina (2006) suggested an important role for tidal mixing within the central Aleutian and northern Kuril regions in explaining $O_2$ decline in the Northwest Pacific Intermediate Water. Also, it is possible that less $O_2$ has been replenished from the surface in source regions of the Oyashio Current, either in the Sea of Okhotsk (Nakanowatari et al. 2007) or the Bering Sea (Andreev and Watanabe 2002). The distinct formation mechanism may have led to the distinct temporal evolution of $O_2$ anomalies within the NPIW density class.

5. Summary

The key process by which $O_2$-rich mixed layer waters are transferred to the North Pacific thermocline is the subduction resulting from repeated seasonal cycles of the mixed layer depth and outcrop area. The surface outcrop window expands when the ocean takes in $O_2$, and it contracts when the ocean takes $O_2$ out of the thermocline. When averaged over the annual cycle, only a portion of seasonally subducted water is entrained back into the mixed layer during the seasonal obduction period. As a result, annual mean subduction occurs through the late winter outcrop (e.g., Figs. 5b and 5f; Stommel 1979).

The $O_2$ uptake rate through the base of the mixed layer is found to be sensitive to climate variations and the associated rearrangement of surface density fields. The outcrop window of North Pacific thermocline water has varied in concert with the PDO. When the central North Pacific is in a cold phase, the winter outcrop area of the CMW density class expands farther southward in the central North Pacific Ocean, allowing more $O_2$-rich surface water to enter the thermocline. Consistent with the decadal variations in the $O_2$ uptake rate, positive $O_2$ anomalies appear near the winter outcrop of the CMW density class during a time period of 1975–84.

The enhanced $O_2$ supply rate during the 1980s, combined with intensified gyre circulations (Deutsch et al. 2005; 2006), may have led to an increase in $O_2$ content within the ventilated thermocline of the subtropical gyre during 1985–2004. While $O_2$ variability within the CMW density class exhibits multidecadal fluctuations linked to the PDO, $O_2$ has persistently declined over the past 50 years in density layers that do not directly outcrop at the open ocean’s surface (e.g., Ono et al. 2001; Emerson et al. 2004; Andreev and Baturina 2006; Whitney et al. 2007). Distinct temporal evolutions in $O_2$ content between the CMW density class ($\gamma = 25.6$–26.6) and the NPIW density class ($\gamma = 26.7$–27.2) suggest distinct mechanisms as the primary cause of $O_2$ changes.

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