Oxygen Seasonality, Utilization Rate, and Impacts of Vertical Mixing in the Eighteen Degree Water Region of the Sargasso Sea as Observed by Profiling Biogeochemical Floats

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Abstract  Seasonal oxygen structure and utilization in the Sargasso Sea are characterized using nine profiling floats with oxygen 2021 sensors (years 2005–2008), deployed in an Eighteen Degree Water (EDW) experiment (CLIMODE). During autumn-winter when the mixed layer is deepening, oxygen increases from the surface to the base of the EDW at 400 m. During spring-summer, oxygen decreases except between the seasonal pycnocline and compensation depth, creating the seasonal shallow oxygen maximum layer (SOMax) with oxygen production of 0.04 μmol kg⁻¹ day⁻¹. In the underlying seasonal oxygen minimum (SOMin), the oxygen utilization rate (OUR) is 0.10 μmol kg⁻¹ day⁻¹, decreasing with depth to 0.04 μmol kg⁻¹ day⁻¹ in the EDW. Remineralization in May to August is double that of August to November. The Sargasso Sea is a net carbon producer; estimated annual export production from the top 100–250 m is 2.9 mol C m⁻² and from the top 400 m is 4.2 mol C m⁻². Below the EDW, oxygen decreases seasonally at the same time as in the EDW, indicating remineralization down to 700 m. However, on isopycnals in this deeper layer, oxygen increases during May to September, likely due to lateral advection from nonlocal surface outcrops. Summer shoaling of these isopycnals creates this paradox. The complex vertical oxygen structure in the upper 200 m enables important vertical diffusive flux that modifies the OUR calculated from oxygen change. Ignoring mixing underestimates maximum remineralization by 19% and underestimates maximum net production by 88%. However, vertical mixing is negligible in the deeper layers, so the associated total integrated remineralization error is 5%–9%.

1. Introduction

Oxygen in the ocean interior is sourced by air-sea gas exchange during ventilation and photosynthesis near the surface and is drawn down by aerobic respiration. It follows that, on a sufficient time scale, one can estimate the oxygen utilization rate (OUR) below the euphotic zone by physically tracing a previously ventilated water parcel while measuring its oxygen concentration (Riley, 1951). OURs in the ocean interior indicate the strength of the biological pump, which draws carbon from the atmosphere and sequesters it into the deep ocean. One can determine the net export of fixed carbon at a given depth by integrating the rate of the annual oxygen consumption below, inferring the amount of organic carbon required to sustain remineralization (Bushinsky & Emerson, 2015; Jenkins, 1982; Martz et al., 2008). This respiration is also correlated with the rate of regeneration of the nutrients that support photosynthesis. Thus, understanding the variability of remineralization in the ocean interior, in particular the rate, depth dependence, and seasonality, is fundamental when using oxygen concentrations to diagnose the carbon cycle. This variability is complicated by the modification of oxygen concentration by physical transport, including advection and mixing.

Upper ocean variability in the Sargasso Sea of the western North Atlantic is marked by the thick layer of subtropical mode water, “Eighteen Degree Water (EDW),” its erosion and regeneration, and the cycle of the seasonal pycnocline above the EDW (Billheimer & Talley, 2016a; Talley & Raymer, 1982; Worthington, 1959). The seasonal pycnocline is identified by the strongly stratified layer in the upper 200 m of the water column. The underlying EDW layer is identified for this work by the portion of the ςθ = 26.2–26.7 kg m⁻³ density layer with potential vorticity (PV) less than 1 × 10⁻¹⁰ m⁻¹ s⁻¹ (Billheimer & Talley, 2016a), where PV is defined, neglecting relative vorticity on large scales.
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**Figure 1.** Profile locations of the nine Argo-equivalent CLIMODE floats with oxygen sensors, November 2005 to April 2008. Yellow: profiles in February to March, when the deepest mixed layers occur in the region. Blue: profiles from other months. Green (red) circles: first (last) profile of a given float. Floats are labeled at last position. Gray dots: profiles not included in the Eighteen Degree Water (EDW) annual average in Figure 3. Cyan “x”: location of Bermuda and the BATS and Station S time series. The mean Gulf Stream location during the sampling period is shown by the thick black curve, calculated as the location at which the temperature at 200 m equals 15°C (Fuglister, 1963) using the Roemmich-Gilson Argo Climatology (Roemmich & Gilson, 2009).

\[
P V = -\frac{f \frac{\partial \rho}{\partial z}}{\rho},
\]

where \( f \) is the Coriolis parameter and \( \rho \) is locally referenced density.

EDW has two sources: (1) winter convection driven by cold, dry air outbreaks, which creates deep mixed layers in the northern Sargasso Sea, destroying the seasonal pycnocline, injecting low PV and high, near-saturation oxygen concentration into the EDW layer; and (2) PV dynamics in the eastern portion of the Gulf Stream where this layer outcrops, between 55°W and the Newfoundland Ridge (Joyce et al., 2013). For the latter, a portion of the Gulf Stream “exits” and recirculates back to the southwest into the Sargasso Sea, carrying ventilated water in the EDW and denser layers. Winter convection is the dominant source of low PV and high oxygen in the EDW, as is apparent from their immediate onset after winter convection and rapid decay in years when convection is weak, even while Gulf Stream-sourced EDW continues (Billheimer & Talley, 2016a). Both processes affect the seasonal oxygen cycle in the Sargasso Sea.

Beginning roughly in May, solar insolation halts the winter deep convection process, allowing quick formation of the seasonal pycnocline through both vertical restratification and horizontal slumping (Mahadevan et al., 2012; Johnson et al., 2016), isolating the EDW layer from the surface. Over the course of summer and fall, EDW PV increases, due to mixing with surrounding water masses, while EDW oxygen decreases, driven by both mixing and aerobic remineralization. The uniform oxygen of the thick winter EDW layer is an ideal initial condition for observing OURs throughout the remainder of the year when EDW is isolated from the atmosphere.

Injection of high oxygen, low nutrient surface waters during EDW renewal also decreases nutrient concentration in the thick EDW layer, which separates high nutrients below the main nutricline underlying the EDW from the euphotic zone where they can be utilized (Falter et al., 2005). The low nutrient/high oxygen EDW layer is apparent on vertical sections in the Sargasso Sea in atlases such as Koltermann et al. (2011) (e.g., WOCE Hydrographic Program sections A20, A22, A03). This leads to a major question about how observed normal levels of productivity are sustained in the nutrient-poor central Sargasso Sea. Fawcett et al. (2018) hypothesize this is associated with export of nutrient-poor organic material by, for example, transparent exopolymer particles.

The Sargasso Sea has a rich history of biogeochemical observations thanks to longstanding repeat hydrographic profiles at Station S and Bermuda Atlantic Time-series Study (BATS) (Figure 1), which lies south of the most intense EDW (Talley & Raymer, 1982). Many studies have estimated OUR at BATS, primarily
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using two distinct methods. The first observes changes in oxygen concentration and models the oxygen budget, estimating the physical transport of oxygen by evaluating upper ocean advection and diffusion (Ono et al., 2001; Riley, 1951), mixed layer/thermocline models (Jenkins, 1987), or inverse models (Rintoul & Wunsch, 1991; Sarmiento et al., 1990). For instance, Ono et al. (2001), focusing on the seasonal pycnocline and 150 m layer below the compensation depth, observe a seasonal cycle in OUR that decreases with depth, with peak remineralization seasonally mirroring the development of the shallow oxygen maximum (SOMax). Vertical mixing and lateral advection were shown to have significant impact on calculated OURs.

The second method evaluates the time rate of change of oxygen concentration, or “age equation” using various chemical tracers to determine the ventilation date (Jenkins, 1982; Jenkins & Goldman, 1985; Sarmiento et al., 1990). Assuming near-100% oxygen saturation during ventilation, one can take a single observation of oxygen saturation at a later time and determine the OUR using the chemical tracer age. Jenkins and Goldman (1985) estimate EDW OUR using this approach, using \(^3\)He/\(^3\)H ratios, and compare that estimate with seasonal changes in oxygen concentration at Station S. They find that OUR obtained from the age equation is consistent with observations of oxygen rate of change seasonally. However, the oxygen rate of change likely underestimates biologically driven OUR. As noted in Ono et al. (2001), there are two major problems, both physical:

1. Not accounting for mixing. Seasonally, oxygen saturation develops a mirror image structure across the compensation depth, \(Z_c\), defined as the depth at which net community production (NCP) equals zero. Immediately after winter, ventilation ceases and the surface warms, creating a shallow seasonal thermocline. Photosynthesis then drives development of a SOMax, in which the increasing oxygen trapped beneath the seasonal thermocline, but still above \(Z_c\), becomes supersaturated. Below \(Z_c\), the SOMax is mirrored by a layer of oxygen undersaturation (shallow oxygen minimum or SOMin), driven by the remineralization of organic matter, producing a clear seasonal cycle in OUR. Turbulence in the vicinity of this large vertical gradient in oxygen concentration should drive a substantial vertical flux of oxygen across \(Z_c\), rendering an underestimate in OUR below \(Z_c\).

2. Not accounting for advection. We observe a time lag of arrival of the seasonal maximum oxygen concentration at depths below the locally outcropping EDW (Section 3.1.2). This could arise from isopycnal advection and mixing from denser winter outcrops northeast of the EDW region (Joyce et al., 2013). This arrival of higher oxygen during the local nonventilation season leads to an underestimate of local OUR beneath the EDW. On the other hand, in the EDW density range, the Gulf Stream advects lower oxygen waters from the south (Palter & Lozier, 2008) (Supplementary Figure S11), which mix laterally into the EDW and could lead to an overestimate of local OUR.

Development of stable oxygen optode sensors (Körtzinger et al., 2005) has allowed high quality, high resolution, year-round Lagrangian observations of the annual oxygen cycle by profiling floats (Bushinsky & Emerson, 2015; Fawcett et al., 2018; Hennon et al., 2016; Martz et al., 2008; Riser & Johnson, 2008). These floats provide a new platform for observing the annual oxygen budget and estimating OURs, providing greater spatial coverage and reducing the effects of advection on estimates at fixed stations (Hennon et al., 2016).

With profiling floats equipped with oxygen sensors that span the northern Sargasso Sea, deployed in the CLIVAR Mode Water Dynamics Experiment (CLIMODE) (Marshall et al., 2009) from 2005 to 2008 (Figure 1), we assess remineralization rates, depth dependence, and seasonality of oxygen concentration throughout the EDW formation region. We diagnose the effect of vertical turbulent flux on the seasonal evolution of oxygen using profiling float-based estimates of vertical eddy diffusivity from Billheimer and Talley (2016a).

We show that turbulent mixing across the compensation depth, \(Z_c\), has a large influence on the oxygen rate of change just above and below \(Z_c\) (~40%) (~25%) (Sections 3.2 and 3.3). The net result is a substantial underestimation of OUR if oxygen rate of change is used alone, without considering oceanic transport of oxygen concentration, agreeing with previous inferences by Jenkins and Goldman (1985) and Ono et al. (2001). Lateral transport of high oxygen from nonlocal sources affects oxygen change in deeper isopycnal layers in the permanent pycnocline.
2. Data and Methods

Nine APEX-SBE profiling floats were deployed in the Sargasso Sea region in November 2005 as part of CLIMODE (http://sam.ucsd.edu/climode_apexfloats/) (Figure 1). Calibrated data are available at Talley et al. (2020). Three additional identical floats were deployed to observe Subantarctic Mode Water formation in the southeast Pacific during the same timeframe, as part of a separate CLIVAR Experiment “SAMFLOC”; their data were also calibrated and included in the same archive. The profilers were equipped with Seabird 41 conductivity-temperature-depth (CTD) instruments and Aanderaa Optode 3830 oxygen sensors, parked at 500 dbar, and alternately profiled from 500 to 1,800 dbar to the surface every 5 days. Profiling float lifetimes ranged from 5.5 to 28 months. Except that the parking depth was 500 m rather than 1,000 m, the floats were Argo-equivalent with double the profiling frequency. The APEX floats were assembled and operated by Webb Research Corporation. Acquisition and initial processing of the Systeme Argos message files were carried out by the U. Washington Argo laboratory. We carried out the delayed mode calibration as described in this section. Float deployment and delayed mode calibration information are given in Table 1. Data from an individual CLIMODE float, “2721,” are presented in Figure 2. Similar time series for all nine floats are included in Supplementary Figures S1 through S9.

EDW is defined as in Billheimer and Talley (2016a): water in the potential density range of $\sigma_\theta = 26.2$–$26.7 \text{ kg m}^{-3}$, and with PV < $1 \times 10^{-10} \text{ m}^{-1} \text{ s}^{-1}$.

2.1. Float Sensor Delayed Mode Calibration

Float conductivity was corrected for thermal lag following Johnson et al. (2007), and corrections for salinity drift and offset were made using the methodology of Owens and Wong (2009), using salinity data from alternating deep profiles. Delayed mode pressure corrections were made for all floats following protocol from Argo Data Management (2013) (updated in Wong et al. [2021]).

Delayed mode corrections to oxygen optode sensors were made for all floats. The data were first corrected for pressure and salinity according to the optode manual (November 4, 2005), using the raw pressure and raw salinity. Next, a quality control process was performed by taking each float individually, partitioning the data into four depth bins (chosen relative to the sampling frequency 0–200, 200–500, 500–1,000, 1,000–1,850 dbar), and removing any data whose fluctuation from the mean was greater than three standard deviations, where the mean and standard deviation are relative to each depth bin for each float. This resulted in removing ~0.1% of the data. Then the pressure was re-corrected, reversing the suggested correction of the 2005 manual and using the recommended pressure compensation of 3.2% per 1,000 dbar of Uchida et al. (2008), this time using the corrected pressure and salinity.

A final oxygen concentration dependent correction was made using bottle oxygen data from the CLIMODE hydrography campaign (http://sam.ucsd.edu/ltalley/climode_ctdhydro/index.html and https://cchdo.ucsd.edu/search?q=CLIMODE) (Table S1; Talley et al., 2020). The SAMFLOC oxygen corrections were based on bottle oxygen data from the AAIW05 campaign (https://cchdo.ucsd.edu/search?q=AAIW05). These floats were built and operated a decade before the finding that collection of an air oxygen value on every profile can remove most of the uncertainty in the optode calibration (Johnson et al., 2015); therefore calibration using in situ oxygen data was necessary. The sensors had very little drift, but significant offsets in general. Float oxygen was therefore corrected by comparing the closest (in time and space) profiles of float optode oxygen to the Winkler-derived oxygen measured on the coincident hydrographic cruises (Tables 1 and S1). For each float, the closest float-to-cast profiles were found within a specified time window of 23 days. Bottle data above 100 m depth were excluded for use in the correction. Float oxygen sensors, in the absence of air oxygen measurements, are now usually calibrated using the World Ocean Atlas (WOA) collection of hydrographic data (Drucker & Riser, 2016; Martz et al., 2008; Takeshita et al., 2013). Our calibrations using nearly coincident shipboard profiles are likely as robust given the high quality of the shipboard oxygen data and the negligible optode sensor drift, and are preferable given the spatial averaging in the WOA climatology that introduces uncertainty given the large isopycnal depth and oxygen gradients in this Gulf Stream region.

The comparison of float to bottle oxygen was considered on pressure or density levels, depending on the correlation of the bottle to float oxygen concentration (Table 1). For example, float-to-bottle profile pairs
near or within the Gulf Stream or far away from each other in space and/or time were especially more highly correlated when compared on density levels. A prominent example is float 2713, which was just on the north side of the Gulf Stream, which caused isopycnal depths to differ between the float and the calibration station, yielding a low correlation coefficient in pressure, but a high coefficient in density space. Comparisons on pressure levels were more highly correlated where large oxygen gradients in density space existed. Seven out of nine floats were compared on density levels. The result of the comparison was an oxygen-dependent correction, derived from Model II regression of the float-to-bottle oxygen difference, and which was applied to the rest of the profiles of a given float.

Float oxygen records were checked for sensor drift by comparing float oxygen time series on isopycnal levels with the WOCE Global Hydrographic Climatology (WGHC) (Gouretski & Koltermann, 2004). For each float, the isopycnal level was chosen to be that with the least oxygen variation in the climatology. No significant sensor drift was found with the exception of an abrupt upward shift by 15–20 μmol kg\(^{-1}\) on float 2724 after profile 102. Optodes have been shown to drift less than 0.3% per year (Bushinsky et al., 2016), which is less than the calibration uncertainty over the 2-year lifetime of the CLIMODE floats.

The quality of the oxygen-dependent correction relative to the climatology was assessed for each float by comparing all float observations to WGHC on depth levels (Table S1 and Figure S10). For seven floats, the root mean square error relative to WGHC was reduced after the correction. Float 2722 had a large remaining error, but had the shortest record (with only 17 1,800-m depth profiles) and spent its entire life in and out of the Gulf Stream, a region with large error for the climatology. Error for float 2724, with the mid-lifetime oxygen shift, also remained large.

### 2.2. OUR and Float Profile Time Series

Remineralization rates were obtained by evaluating the equation for oxygen rate of change:

\[
\frac{\partial O_2}{\partial t} + \mathbf{u} \cdot \nabla O_2 = \nabla \cdot \left( K \cdot \nabla O_2 \right) + \text{OUR}
\]

where \(\mathbf{u}\) is the velocity vector, \(K\) is a diffusivity tensor, and OUR is the remineralization-driven OUR. OUR < 0 indicates oxygen consumption. Following the quasi-Lagrangian trajectories of profiling floats, advective terms are neglected. The OUR is thus:

\[
\text{OUR} = \frac{\partial O_2}{\partial t} - \nabla_H \cdot \left( \kappa_H \nabla O_2 \right) = \frac{\partial}{\partial z} \left( \kappa_z \frac{\partial O_2}{\partial z} \right)
\]

where \(\kappa_z\) is vertical eddy diffusivity, which can be depth dependent, and \(\kappa_H\) is lateral eddy diffusivity, which can be spatially dependent. Both vary seasonally.

Seasonal estimates of upper ocean vertical diffusivity \(\kappa_z\) in the Sargasso Sea are taken from our PV budget analysis using these floats (Billheimer & Talley, 2016a). An upper bound estimate just below the near-surface seasonal thermocline was obtained assuming vertical mixing only, hence parameterizing all processes that lead to apparent vertical mixing of PV. The estimated \(\kappa_z\) ranged from \(\sim 3 \times 10^{-5}\) to \(\sim 17 \times 10^{-5}\) m\(^2\)s\(^{-1}\), low in summer and high in winter. We apply this seasonally varying \(\kappa_z\) at 55 m, linearly fit in depth to a constant \(1 \times 10^{-5}\) m\(^2\)s\(^{-1}\) at 720 m, in the permanent thermocline. This diffusivity profile is consistent with previous measurements of diapycnal diffusivity in the permanent pycnocline of the North Atlantic subtropical gyre.

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**Figure 2.** Float “2721.” Thin black curves are isopycnal contours, ranging from \(\sigma_\theta = 25.8–27.1\) kg m\(^{-3}\) on 0.1 kg m\(^{-3}\) intervals. (a) Float profile locations, with the same features shown in Figure 1. (b) Log potential vorticity. (c) Oxygen concentration.
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(Ledwell et al., 1993) as well as estimates above the permanent pycnocline that decrease with depth, from the 250–500 m depth interval to the 500–1,000 m range (Whalen et al., 2012). The seasonal enhancement is due to higher wind energy in winter that drives a more vigorous internal wavefield, and a more vigorous winter-time mesoscale eddy field, which is also correlated with higher diffusivity (Whalen et al., 2018).

Lateral mixing of PV was also addressed in Billheimer and Talley (2016a) using the CLIMODE isopycnal-following “bobber” floats (Fratantoni et al., 2013) to estimate lateral diffusivity $\kappa_H$, which ranged from $\sim 0.5$ to $3 \times 10^4$ m$^2$·s$^{-1}$. $\kappa_H$ was lowest in the southern Sargasso Sea and highest in the Gulf Stream extension, and maximum in winter. We concluded that 30% of the total post-winter PV decay (increase) in the EDW layer arose from mixing between of high PV from the Gulf Stream PV into the low EDW PV.

Data from the eight full annual cycles sampled by six of the floats in the Sargasso Sea were averaged to represent the EDW region’s seasonal cycle (Figure 3). Float profiles were interpolated onto 5 m intervals and then averaged over the six floats in 15 day, 10 m bins. Averaging reduced abrupt changes in float oxygen concentrations on depth and density levels, when floats were occasionally caught in eddies or transported into different regimes in a non-Lagrangian fashion (an example of such behavior is evident for float 2721 from August to November (Figure 2)). Profiles east of 48°W were not included because they were not in the Sargasso Sea’s core EDW region (gray in Figure 1). Three floats (2713, 2719, and 2722) were omitted from the averaging as they were deployed in the Gulf Stream and were not representative of the EDW region. Float 2713 crossed the Gulf Stream multiple times with repeated forays into the northern and eastern Sargasso Sea before being rapidly swept eastward by the Azores Current. Float 2719 was eventually transported eastward into the North Atlantic Current. Float 2722 remained in the Gulf Stream for its short life. These three floats do provide valuable information on the lateral structure of oxygen, showing locations of low and high oxygen to the north (Gulf Stream) and east of the EDW region (Figures S11 and S12).

Oxygen rate of change was calculated on both depth and density levels (Figures 4a and 4b), using a sliding 6-month time scale over the seasonally averaged float oxygen concentration. That is, for every 15 day time step, a line was fit to the oxygen concentrations spanning 90 days before and after the time step (on a looping annual record), the slope of which yielded $\partial O_2/\partial t$. Depth levels were 10 m intervals from 0 to 800 m. Density levels were $\sigma_0 = 25.7$–26.3 kg m$^{-3}$ on 0.2 kg m$^{-3}$ intervals and $\sigma_0 = 26.4$–27.1 kg m$^{-3}$ on 0.05 kg

| UW ID | Deploy date | Last date | $n_p$ | $\Delta t$ (days) | $\Delta d$ (km) | $r_\sigma$ | $r_p$ | $\Delta o_{corr}$ (μmol kg$^{-1}$) |
|-------|-------------|-----------|------|------------------|----------------|-----------|------|--------------------------|
| CLIMODE (EDW) | | | | | | | | |
| 2713 | November 10, 2005 | October 31, 2007 | 143 | 4 | 38 | 0.9976 | 0.6789 | <5 |
| 2719 | November 13, 2005 | October 13, 2007 | 139 | 6 | 137 | 0.9971 | 0.9628 | 5–8 |
| 2720 | November 17, 2005 | October 7, 2006 | 65 | 15 | 74 | 0.9761 | 0.9528 | 10–25 |
| 2721 | November 17, 2005 | September 3, 2007 | 130 | −13 | 42 | 0.9625 | 0.9719 | <5 |
| 2722 | November 14, 2005 | April 27, 2006 | 34 | −6 | 11 | 0.9637 | 0.9810 | 10–25 |
| 2723 | November 18, 2005 | May 22, 2007 | 109 | 14 | 67 | 0.9876 | 0.9782 | 22–30 |
| 2724 | November 20, 2005 | April 8, 2008 | 174 | −14 | 44 | 0.9940 | 0.9826 | 22–30 |
| 2725 | November 23, 2005 | July 21, 2006 | 48 | 23 | 88 | 0.8857 | 0.8575 | 15 |
| 2726 | November 15, 2005 | January 18, 2007 | 86 | 15 | 68 | 0.9944 | 0.9920 | 10–15 |
| SAMFLOC (AAIW) | | | | | | | | |
| 2827 | February 22, 2006 | May 21, 2008 | 165 | 5 | 84 | 0.9995 | 0.9993 | 17–20 |
| 2848 | February 22, 2006 | September 25, 2007 | 117 | 5 | 20 | 0.9997 | 0.9997 | −2 to −6 |
| 2849 | March 1, 2006 | May 4, 2007 | 87 | 4 | 5 | 0.9992 | 0.9996 | −2 to −6 |

Note. U. Washington (UW) ID number, date, number of 500 dbar profiles ($n_p$), time and distance between the closest float profile and bottle cast ($\Delta t$, $\Delta d$), correlation coefficients between matches on density and pressure levels ($r_\sigma$, $r_p$; bold indicates oxygen calibration in density or pressure), and range of oxygen-dependent correction ($\Delta o_{corr}$). $\Delta t < 0$ indicates that the nearest float profile to a bottle cast occurred before the bottle cast. See also Table S1.
Figure 3. Averaged time series of profiles for the six floats indicated by blue and yellow dots in Figure 1. Mean mixed layer depth for the six floats (black dashed curve). The mixed layer depth of a given profile is defined by the depth at which the difference in density exceeds a threshold of $\Delta \sigma = 0.03$ kg m$^{-3}$ from the density at 10 m depth (de Boyer Montégut et al., 2004). Thin black curves contour the same isopycnals as in Figure 2. (a) Potential temperature, $\theta$. (b) Log potential vorticity (PV). (c) Oxygen concentration. (d) Oxygen concentration in isopycnal coordinates.

Figure 4. Oxygen rates of change with mean mixed layer depth (black dashed curve). (a) Oxygen rate of change, calculated along isobars from the values of Figure 3c. (b) Oxygen rate of change, calculated along isopycnals from the values of Figure 3d, retranslated to depth coordinates using mean isopycnal depth. Depth range is cut off above 73 m because that is, the mean depth level of $\sigma = 25.8$ kg m$^{-3}$. (c) Oxygen rate of change, composed of the upper 250 m from Figure 4a and below 250 m from Figure 4b. Black contours in (a and c): isopycnals. Green contour: zero oxygen change. Colorbar is the same for all three panels. (d) Average profiles of oxygen rate of change (nonwinter) based on (c). The solid black dashed curve is the mean profile from May to November, the red curve is the mean profile from May to August, and the blue curve is the mean profile from August to November.
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m⁻³ intervals, since a coarser range of isopycnals was needed to represent even depth spacing within the seasonal pycnocline.

The vertical gradient of oxygen was calculated uniformly on a 50 m vertical scale, while the second derivative was calculated on a 50 m scale above 200 m depth and on a 100 m vertical scale below 200 m depth.

Changes in oxygen concentration due to aerobic remineralization are often described by

\[ \Delta [O_2]_{\text{remin}} = [O_2]_{\text{observed}} - [O_2]_{\text{preformed}} \]  

where \([O_2]_{\text{preformed}}\) is the oxygen concentration at the time of ventilation. The OUR is

\[ \text{OUR} = \frac{\Delta [O_2]_{\text{remin}}}{\Delta t} \]  

where \(\Delta t\) is the time since ventilation. With the high temporal resolution of float profile oxygen, we calculate OUR directly from the changes in oxygen between profiles, corrected for vertical mixing (Equation 3). We use Equation 5 instead of the usual apparent oxygen utilization (AOU), for which preformed oxygen is replaced by oxygen at 100% saturation:

\[ \text{AOU} \approx \frac{[O_2]_{\text{observed}} - [O_2]_{\text{sat}}}{\Delta t} \]  

The AOU approach (Equation 6) is useful when an observation of oxygen at one point in time is paired with a tracer providing \(\Delta t\) since ventilation. Errors in this OUR estimate arise from assumptions that temperature is constant after ventilation and that \([O_2]_{\text{sat}} = [O_2]_{\text{preformed}}\) (e.g., Ito et al. 2004).

3. Results

3.1. Seasonal Cycles

The average seasonal cycle of potential temperature, PV, and oxygen concentration (Figure 3) is described here in five layers: the seasonal pycnocline, SOMax layer, SOMin layer, EDW layer, and permanent pycnocline. Unaveraged data from float 2721 (Figure 2) illustrate the onset and extremes of the annual cycle more graphically. Time series plots for all nine floats are provided in the supplement (Figures S1–S9).

3.1.1. Mixed Layer and Seasonal Pycnocline

The largest seasonal variations in oxygen, temperature, and stratification in the northern Sargasso Sea occur in the upper 150 m, dominated by the formation and erosion of the seasonal pycnocline. The seasonal pycnocline, marked by high PV, forms in May, and warms and intensifies until September (Figures 3a and 3b); the average mixed layer depth during this period is shallower than 25 m. During this time oxygen content above the seasonal pycnocline decreases (Figures 3c and 2c), consistent with decreasing oxygen solubility due to warming and resulting outgassing.

Conversely, at the onset of surface cooling in mid-September, the mixed layer and the center of the seasonal pycnocline start to deepen and their oxygen concentration begins to increase, consistent with enhanced solubility and ingassing. The seasonal pycnocline weakens until it is obliterated by the deep mixed layers that connect all the way through to the underlying EDW layer in late winter. At this point, high oxygen content reaches down to the bottom of each float's deepest winter mixed layer (Figure 2c), which is deeper than the averaged mixed layer depth (Figure 3c). This ventilates the EDW, replenishing its source properties (oxygen and PV shown here). The highest surface and seasonal pycnocline oxygen are achieved after the maximum mixed layer deepening that reduces the oxygen saturation through entrainment of lower oxygen underlying waters; when convective entrainment ceases, oxygen in the still cold mixed layer is free to increase to saturation.

3.1.2. SOMax

Just below the seasonal pycnocline, a “SOMax” develops from spring to fall. This is a manifestation of the Shulenberger-Reid effect (Shulenberger & Reid, 1981), due to photosynthesis in the euphotic zone that extends beneath the developing springtime thermocline. The overlying seasonal thermocline acts as a barrier for oxygen outgassing, trapping photosynthetically produced oxygen below. This results in increasing
oxygen concentration and oxygen supersaturation. Supersaturation can also result from warming of the trapped layer if the warming rate exceeds consumption, but because we plot oxygen concentration rather than saturation, the effect shown is due to photosynthetic oxygen production.

Riser and Johnson (2008) observed the SOMax in the subtropical North and South Pacific using profiling floats equipped with oxygen sensors. They showed that the subtropical euphotic zone is an annual net producer of fixed carbon, contrary to previous in situ bottle experiments, without having to invoke episodic bloom events. For the Sargasso Sea, Jenkins and Goldman (1985) confirmed the photosynthetic origin of the SOMax in the euphotic zone, but showed that it is complicated by the loss of oxygen to the atmosphere and downward mixing as the seasonal thermocline breaks down. Recent post-CLIMODE profiling floats with both oxygen and nitrate sensors, deployed near Bermuda from late 2009 to 2014, have been used to explore the conundrum of robust NCP in the Sargasso Sea as estimated from oxygen and carbon budgets, but with low nutrient availability due to the thick, low-nutrient buffer of the EDW layer and a non-Redfieldian relationship of nitrate to oxygen and carbon, as mentioned in Section 1 (Fawcett et al., 2018).

Figure 3, from our CLIMODE floats, shows that the Sargasso Sea SOMax reflects the strength and depth of the seasonal pycnocline. Supersaturation of oxygen begins to develop in early May. Oxygen concentration in this layer continues to increase until September, when surface heating changes sign and the seasonal pycnocline begins to weaken. As the surface mixed layer gradually deepens, isopycnals are successively stripped from the seasonal pycnocline and outcrop, deepening the extremum of the seasonal pycnocline. As this occurs, the SOMax begins to equilibrate from the top, reducing its magnitude and revealing a new, deepening maximum that tracks the seasonal pycnocline downward until both the pycnocline and SOMax are destroyed by the winter convection that produces and ventilates the EDW.

Note that temperature in the seasonal thermocline (Figure 3a) increases during the development of the SOMax, penetrating to the depth of the SOMin layer. Below the mixed layer, this does not change the oxygen concentration, but does affect AOU, leading to differing rates of change of oxygen concentration versus AOU. As said above, we therefore focus on time rate of change of oxygen concentration as the most informative indicator of OUR.

3.1.3. SOMin

Just below the euphotic zone and below the SOMax, at ~100 m, an oxygen minimum layer (“SOMin”) appears (Figure 3c). (The minimum is more pronounced in unaveraged float data, e.g., Figure 2c.) The existence of this layer is consistent with previously estimated OUR depth profiles, which peak just below the compensation depth, $Z_c$, and exponentially decrease with depth (Jenkins, 1980, 1987; Martin et al., 1987; Martz et al., 2008; Riley, 1951; Sarmiento et al., 1990). In late winter, the mixed layers that regenerate EDW are much deeper than $Z_c$, introducing a nearly uniform, saturated oxygen concentration. When the deep EDW layer is capped by seasonal warming, high oxygen consumption just beneath the compensation depth draws down the oxygen concentration quickly compared with greater depths within the EDW, resulting in the SOMin within the evolving EDW. In non-EDW regions, winter mixed layers do not exceed $Z_c$ (or only exceed $Z_c$ by a small amount), which produces a vertical oxygen gradient that already decreases with depth before the exponentially decreasing profile of OUR is allowed to act to produce a SOMin.

3.1.4. EDW Layer

The EDW layer is characterized by nearly vertically uniform temperature, with a low PV (low stratification) core between the $\sigma_\theta = 26.4$ and $\sigma_\theta = 26.5$ isopycnals. EDW is renewed in late winter, when convective mixing injects high oxygen concentration into deep mixed layers. While the late winter mixed layer depth averaged over the six CLIMODE floats is ~250 m (Figure 3), winter mixed layer depths regularly exceed this level in the most intense EDW formation region in the northern Sargasso Sea near the Gulf Stream extension (Worthington, 1972b), reaching deeper than 450 m in some cases (Figure 2). These events are critical for introduction of high oxygen concentration below the average mixed layer depth. This is apparent in Figure 3c, as oxygen concentration increases in the 250–450 m layer during late winter.

After development of the seasonal pycnocline in May, isolating the EDW layer from the atmosphere, the oxygen concentration decreases at seasonally variable rates. From May through July, oxygen is more rapidly depleted, while from August through December the rate slows.
Part of the EDW’s summer oxygen depletion could be due to lateral, along-isopycnal mixing with Gulf Stream waters (Palter & Lozier, 2008). The Gulf Stream west of 55°W carries low oxygen waters, as apparent in the EDW isopycnal layers sampled by the CLIMODE floats, including floats 2013, 2019, and 2022 (Figures S11a and S11b). This would mirror Billheimer and Talley (2016a)’s finding, that low PV in the EDW is partially removed (increased) by lateral mixing with high PV from the Gulf Stream, with 30% of the PV destruction due to lateral mixing and 70% due to vertical mixing with overlying, more stratified water. (The Gulf Stream has a zonal dipolar structure in oxygen and PV: low oxygen, high PV at EDW densities west of 55°W, and high oxygen, low PV east of 55°W, which is the second source of renewed EDW discussed by Joyce et al. [2013].)

The CLIMODE float years (2005–2008) were characterized by strong convective EDW formation, which maintained the high oxygen, low nutrient core of EDW. When EDW formation nearly completely ceased in winter 2011–2012, oxygen within the EDW decreased and nutrients increased, indicating continued remineralization without vigorous ventilation (Billheimer & Talley, 2013; Fawcett et al., 2018). Weak EDW formation retreated to its Gulf Stream “exit” source east of 55°W (Joyce et al., 2013), where CLIMODE floats also show high oxygen (Figures S11a and S11b). Weak EDW formation has continued to characterize more recent years as well (Billheimer & Talley, 2016b; Stevens et al., 2020).

### 3.1.5. Permanent Pycnocline

The Sargasso Sea’s permanent pycnocline, beneath the EDW layer, deepens from winter to early summer and shoals from summer to autumn (Worthington, 1972b). This annual cycle is evident in our averaged observations of the depth of the $\sigma_\theta = 26.5$ to $\sigma_\theta = 27.0$ kg m$^{-3}$ isopycnals (Figure 3). Worthington (1977) conjectured that the seasonal deepening by 100–150 m is due to intense late winter convective mixing. Of our 10 complete winter EDW formation events followed by summer/autumn restratification, nine exhibited pycnocline shoaling during restratification (Figures S1–S9). However, the magnitude and timing of the individual rebounds differ. Some especially strong rebounds, such as on float “2721” of Figure 2, could be amplified by local cyclonic eddies. The rebound on float “2720” could be due to migration of the float southward during summer to the permanent pycnocline ridge at about 36°N.

The importance of isopycnal heave in physically shifting the oxygen distribution is apparent when oxygen is plotted as a function of density instead of depth (Figure 3d). Oxygen in the permanent pycnocline, at $\sigma_\theta > 26.5$ kg m$^{-3}$, is much more uniform in time compared with oxygen in depth coordinates. Isopycnals in the permanent pycnocline do not outcrop at the sea surface, hence are not ventilated, in the EDW region of the Sargasso Sea. High oxygen indicating ventilation is found to the northeast in the Gulf Stream “exit” region east of 55°W where Gulf Stream-sourced EDW also originates (Joyce et al., 2013), and farther east (Figures S11c and S11d).

### 3.2. Oxygen Rate of Change

Calculating the rate of oxygen change is the first step toward calculating OUR, which also includes an adjustment for vertical mixing (Section 3.3. Eq. 3). We calculate the rate of change in both depth and density layers (Figures 4a and 4b based on Figures 3c and 3d). Each time step represents the middle of a 6-month differentiation time scale. For the upper 250 m, depth is our coordinate choice since properties are affected by physical proximity to the surface fluxes, euphotic zone, and mixed layer depth. Below 250 m, the oxygen rate of change is dominated by isopycnal heave (Section 3.1.5). Thus, in order to best represent the deeper oxygen processes, we differentiate oxygen concentration along isopycnal levels and project it back to depth coordinates using mean isopycnal depth. We then merge the two at 250 m with linear interpolation (Figure 4c). This is the approximate depth of $\sigma_\theta = 26.4$ kg m$^{-3}$, which is the top of the EDW, does not cross the compensation depth, and is not influenced by fluctuations in the permanent pycnocline depth. The oxygen change profiles are then averaged for the warm seasons when there is a seasonal pycnocline: over summer (May to August) and autumn (August to November), and over the full time period (Figure 4d).

Within the surface mixed layer, using depth coordinates (Figures 4a and 4c), the late winter oxygen increase results from ingassing and introduction of oxygen via mixed layer development. When the seasonal pycnocline develops beginning in May, the near-surface layer warms, driving oxygen outgassing above the mixed layer depth. When surface heat flux changes sign in September, the oxygen rate of change in this surface
layer reverses, indicating oxygen ingassing as a result of enhanced solubility. These are reflected in the mean seasonal profiles (Figure 4d), with strong negative oxygen change in May to August, and the beginnings of the winter oxygen increase apparent in the August to November average. We have not estimated air-sea oxygen flux directly from the floats; such estimates are best if air is sampled by the optodes and used for calibration, as introduced in recent years (Bushinsky & Emerson, 2015, 2018; Bushinsky et al., 2017).

Below the seasonal pycnocline, a very thin layer with $\partial O_2/\partial t < 0$ drives the development of the SOMax from May to mid-August, descending from ~40 to ~75 m depth (Figure 4a). A tongue of $\partial O_2/\partial t < 0$ follows, descending over the same depth range from September to December. This is an artifact of isobaric differentiation across the SOMax that descends from summer into autumn, following the seasonal pycnocline that is evident in PV (Figure 3b). The descent of the positive oxygen change peak is also apparent in the mean profiles (Figures 4a and 4c). Since the peak in $\partial O_2/\partial t$ changes depth seasonally, averages across depth levels mask its magnitude, disguising the persistent tongue of positive $\partial O_2/\partial t$ from May to September (visible in Figure 4d) that produces the SOMax. The oxygen rate of change (ORC) is slightly positive in May-August ($0.005 \mu\text{mol kg}^{-1}\cdot\text{day}^{-1}$), and, although negative, is a positive peak relative to the ORC profile for the other months shown. This and the underlying SOMin are the oxygen change features that are most affected by vertical mixing (Section 3.3).

At 100–150 m, a layer of maximum oxygen decrease, $\partial O_2/\partial t < 0$, is observed, peaking in magnitude from May to August (Figures 4a and 4c), creating the SOMin. This layer has seasonality similar to the oxygen changes in the SOMax and seasonal pycnocline, descending while weakening from July to December, where the maximum decrease reaches a depth of 250 m. In the averaged profiles (Figure 4d), negative peaks in the 100–200 m range mark the maximum consumption layer that produces the SOMin. $\partial O_2/\partial t$ is strongly negative from May to August (~0.13 $\mu\text{mol kg}^{-1}\cdot\text{day}^{-1}$), while weaker and deeper from August to November, with maximum net oxygen decrease mirroring the maximum net oxygen increase in the overlying layer.

The EDW oxygen changes are best described from the hybrid depth-density coordinates (Figure 4c). The annual cycle is dominated by oxygen increase during ventilation season and decay the rest of the year. The oxygen increase between 300 and 450 m, which is calculated in isopycnal coordinates, lags the EDW signal above 300 m. The lag is likely due to the time it takes for the signal of the deepest EDW convection to spread throughout the EDW region. While we estimate the mean mixed layer depth of the Sargasso Sea to penetrate as deep as 250 m, individual float mixed layer depths in the most intense EDW formation regions can reach 500 m (e.g., Figure 2 and mixed layer depth maps from these floats and others in Billheimer and Talley [2013]). We associate the oxygen decrease from May to December with remineralization. The rate of change slows in late summer, associated with the greatest shoaling of the permanent pycnocline. We speculate that this slowing of oxygen decline is associated with arrival of higher oxygen from the Gulf Stream “exit,” associated with adjustment of the gyre circulation. In the averaged profiles (Figure 4d), oxygen depletion decreases approximately linearly with depth through the EDW layer (from ~0.07 $\mu\text{mol kg}^{-1}\cdot\text{day}^{-1}$ at 200 m to 0.01 $\mu\text{mol kg}^{-1}\cdot\text{day}^{-1}$ at 400 m depth), and is positive below that, likely due to lateral advection.

Beneath the EDW, the oxycline and permanent pycnocline are coincident (depth > 400 m and $\sigma_\theta > 26.55$ kg m$^{-3}$) (Figure 3c). Hence when density changes along an isobar there is a corresponding change in oxygen concentration. In depth coordinates (Figure 4a), as isopycnals below 300 m descend from January to May, oxygen concentration uniformly increases throughout the water column below 300 m. From May to September, the reverse occurs. As isopycnals descend over a short time period beginning in September, oxygen follows the isopycnals, producing a positive oxygen change in depth coordinates. Because it is tightly tied to the pycnocline depth, this oxygen structure tells us little about oxygen processes, including remineralization and arrival of high oxygen from remote surface sources. Thus isopycnal coordinates are essential (Figures 4b and 4c).

Even in isopycnal coordinates, oxygen rates of change in the permanent pycnocline have a simple but initially puzzling structure, with oxygen increase arriving in spring-summer. This cannot reflect local ventilation since the water column is highly stratified at this time. The late arrival must result from along-isopycnal transport of oxygen from the denser winter surface outcrops outside the EDW region. Our small set of floats suggests two surface sources of high oxygen concentration (Figures S11c and S11d): the Gulf Stream between 55°W and 50°W (Newfoundland Ridge), and the Newfoundland Basin east of the Newfoundland
Ridge. The Gulf Stream location is the “exit” region for low PV into the EDW layer from the southern Gulf Stream, and produces a weak mode water regardless of the strength of winter convection and mixed layer depths in the Gulf Stream recirculation (Billheimer & Talley, 2013; Joyce et al., 2013). This “exit” region also directs high surface oxygen into the permanent pycnocline beneath the EDW (Joyce et al., 2013). The eastern source regions are connected back toward the Sargasso Sea through westward recirculation (Figure 9.1 in Talley et al. [2011]), which is also evident from visual inspection of many Argo trajectories (not shown). The simultaneous arrival of high oxygen in all isopycnal layers below 450 m could also suggest a dynamical influx of waters carrying high oxygen during the shoaling of the permanent pycnocline due to cessation of deep winter convection within the EDW (e.g., Worthington, 1972b).

Importantly, the mean profiles in Figure 4d are oxygen rates of change and not OUR. Where there is a significant contribution from turbulent mixing (next section) and lateral transport, equating oxygen rate of change with OUR could lead to large error (Equation 3). Subsurface oxygen and PV in the subducted EDW layer are correlated (Figure 5). If both quantities evolved linearly, one would expect a strong correlation, but the evolution of each could be driven by separate mechanisms. Neither PV nor oxygen concentration evolves linearly with time (Billheimer, 2016) (Figure 4d), so the fact that they are strongly correlated ($r = -0.63$, 95% significance level of 0.015, zero lag on 5-day resolution) suggests a shared evolution mechanism that is physical in nature.

3.3. OUR Including Vertical Mixing and Lateral Transports

In this section we calculate OUR from the oxygen rate of change (Section 3.2), corrected for vertical mixing. We noted in the previous section that lateral transport processes, including mixing, cannot be ignored in the deeper layers, that is, the EDW and permanent pycnocline, and we address this briefly at the end of this section.

3.3.1. Vertical Mixing

Vertical mixing depends on the vertical diffusivity, $\kappa_z$, and the vertical gradient of oxygen. The diffusivity $\kappa_z$ from Billheimer and Talley (2016a), is depth and time-dependent (Section 2.2). The vertical gradient of oxygen has a rich structure that reflects the seasonality of the SOMax, SOMin, and EDW layers (Figure 6a). At the bottom of the EDW layer at 500 m, a band of $\partial O_2 / \partial z > 0$ persists throughout the year, as oxygen increases upward from the underlying permanent pycnocline to the oxygen-rich EDW layer. At the top of the EDW layer at ~200 m, the sign of the gradient reverses to $\partial O_2 / \partial z < 0$, ascending from the EDW layer to the SOMin. Transitioning from the SOMin to the SOMax at ~100 m, the oxygen gradient reverses sign once again, becoming strongly positive. There is one final sign reversal, crossing from the SOMax into the mixed layer, which has a lower oxygen concentration. The upper 300 m bands of positive and negative oxygen vertical gradient seasonally track the progression of the SOMax and SOMin, deepening throughout the subduction season. The second vertical derivative of oxygen (Figure 6b) is weakly positive just below the EDW layer, weakly negative within the EDW layer, strongly positive in the SOMin layer, and strongly negative in the SOMax.

The vertical mixing term of Equation 3 is plotted in Figure 6c, using the time-evolving $\kappa_z$ profile and the vertical oxygen gradient of Figure 6a and taking the vertical derivative of the product. Below the mixed layer, the result is weakly seasonal, with its dominant non-winter features summarized by the May to November mean profile (Figure 6d).

Turbulent mixing through the SOMax layer introduces low oxygen water from the seasonal pycnocline above and from the oxygen minimum layer below. Therefore, ignoring vertical mixing in this oxygen maximum layer overestimates respiration and hence underestimates NCP. For May-August, vertical mixing reduces the oxygen concentration by 0.045 μmol kg$^{-1}$·day$^{-1}$. In the mirror image SOMin layer, vertical mixing...
with high oxygen water from both above and below also has a significant impact, increasing oxygen by up to 0.028 μmol kg\(^{-1}\)·day\(^{-1}\); ignoring it leads to an underestimate of OUR/remineralization.

Vertical mixing has a much weaker effect on EDW and pycnocline oxygen concentration because the vertical gradients are smaller. Low oxygen water from the SOMin layer above and low oxygen water from the permanent thermocline below are transported via turbulent flux into the oxygen-rich EDW layer, at a rate of 0.005 μmol kg\(^{-1}\)·day\(^{-1}\). Thus ignoring vertical mixing results in a small overestimate of remineralization rate in the EDW. There is a small increase in oxygen in the pycnocline just below the EDW due to downward mixing of oxygen, which results in a small underestimate of remineralization if vertical mixing is ignored.

### 3.3.2. OUR and Compensation Depth

Our estimated OUR (Figure 7a) is the oxygen rate of change (Figure 4c) minus the turbulent mixing rate of oxygen (Figure 6c), assuming negligible lateral mixing. Compared with an estimate based only on oxygen rate of change, this OUR has enhanced NCP in the SOMax and remineralization in the SOMin. It features a seasonally continuous oxygen production layer, with positive NCP throughout the time of year when the seasonal pycnocline is present. Note that the OUR estimate is limited to the water column below 55 m since that is the shallowest level at which we could generate estimates for second vertical derivatives using reasonable vertical length scales. Additionally, OUR is not estimated within the mixed layer since our \(\kappa_z\) estimates are not valid there.

To quantify the effect of vertical mixing on OUR, hence remineralization rates, we look at the oxygen rate of change, \(\partial\text{O}_2/\partial t\), and OUR for August (Figure 7b). This allows comparison with previous profiling float studies that produce a single seasonal estimate of OUR (or NCP) by fitting a line to oxygen concentrations over a set period of time, usually a six to eight month range, during the season when the mixed layer does not influence interior oxygen concentration (Hennon et al., 2016; Martz et al., 2008; Riser & Johnson, 2008). Due to our 6-month differentiation time scale, an individual August profile represents a rate of change using information from May to November, excluding mixed layer data. This yields our best estimate for the overall annual magnitude of OUR.

We see that the effect of vertical mixing on the annual magnitude of OUR is significant in the upper 200 m in the SOMax and SOMin (Figure 7b). Our estimated OUR in the SOMax is 0.04 μmol kg\(^{-1}\)·day\(^{-1}\) (production).
based on the August oxygen rate of change of 0.005 μmol kg⁻¹·day⁻¹ (Section 3.2). This is consistent with Riser and Johnson’s (2008) float-based estimates of ∼0.05 μmol kg⁻¹·day⁻¹ in the subtropical North Pacific and ∼0.025 μmol kg⁻¹·day⁻¹ in the subtropical South Pacific. All are lower than Jenkins and Goldman’s (1985) Sargasso Sea oxygen rate of change between March and July, of 0.20 ml l⁻¹, which equates to ∼0.071 μmol kg⁻¹·day⁻¹ when vertical mixing is included.

In the SOMin layer, the seasonal cycle of respiration from the compensation depth down to 200 m mirrors the cycle of net production above (Figures 7a and 7c). This is consistent with the paradigm that seasonal remineralization is driven by sinking particulate organic carbon (POC), maximized during the spring bloom (Ono et al., 2001), although remineralization of dissolved organic carbon (DOC) is also enhanced during this time of year (Carlson et al., 1994). From the mean August profile, we estimate a remineralization rate of 0.099 μmol kg⁻¹·day⁻¹ based on an oxygen rate of change of –0.08 μmol kg⁻¹·day⁻¹. (If we use the May-August time frame, the remineralization rate is 0.16 μmol kg⁻¹·day⁻¹ given the oxygen rate of change of –0.13 μmol kg⁻¹·day⁻¹.) With effects of mixing removed (Figure 7b), this is consistent with the maximum remineralization rate of 0.077 ± 0.031 μmol kg⁻¹·day⁻¹ from Hennon et al. (2016), who averaged several profiles of oxygen rate of change from profiling floats in the Sargasso Sea. These estimates are higher than the range of maxima summarized by Sarmiento et al. (1990) of 0.013–0.063 μmol kg⁻¹·day⁻¹, but more conservative than the maximum rate of 0.14 μmol kg⁻¹·day⁻¹ found by Ono et al. (2001).

Hence, not including vertical mixing in a remineralization rate estimate in the SOMin yields a 19% error (0.099 or 0.081 μmol kg⁻¹·day⁻¹ ignoring mixing). Not including vertical mixing in an estimate of NCP in the SOMax produces a more severe error of 88% (0.042 or 0.005 μmol kg⁻¹·day⁻¹ ignoring mixing).

The compensation depth $Z_c$, where OUR is 0, varies seasonally, consistent with Ono et al. (2001), ranging from 55 m in May to 95 m in October (Figure 7a). The August compensation depth, representing the May to
November seasonal value, is 78 m (Figure 7b). This is negligibly modified compared with that from oxygen rate of change alone, because mixing across the compensation depth affects the oxygen maximum above and oxygen minimum below about equally, leaving the zero crossing at about the same depth. Within the limitations of our chosen 10 m vertical resolution, extrema in the upper 200 m are slightly shallower when considering OUR, with peak NCP occurring at 65 m instead of 75 m, and peak consumption occurring at 115 m rather than at 125 m.

In the low PV, high oxygen EDW core centered at 300 m, we estimate a remineralization rate of 0.038 μmol kg$^{-1}$·day$^{-1}$ which is nearly the same as the oxygen rate of change because the vertical mixing contribution of 0.005 μmol kg$^{-1}$·day$^{-1}$ is so weak. This falls between previous estimates by Worthington (1959) and Palter et al. (2005), who used independent different methodologies. Worthington (1959) reported oxygen observations from Rakestraw and Carritt (1948), who repeatedly occupied a Montauk-Bermuda section during 1937–1938. In the EDW formation region in the Sargasso Sea, biological consumption of oxygen in the EDW core was reported at 0.060 μmol kg$^{-1}$·day$^{-1}$ over the course of nine months. Palter et al. (2005) utilized World Ocean Circulation Experiment (WOCE) sections through the EDW region in the Sargasso Sea in 1997 to estimate an EDW OUR of 0.015 ± 0.004 μmol kg$^{-1}$·day$^{-1}$. This estimate was produced by computing −AOU/Δt (Equation 6), where Δt, the time since ventilation, is given by CFC age. We suspect that Worthington (1959)'s estimate is too large because it does not account for the low oxygen introduced to the EDW layer through vertical mixing with the SOMin above, and potentially important along-isopycnal mixing with the low oxygen core of the Gulf Stream (Figure S11; Palter & Lozier, 2008). It is also possible that the three studies differ due to interannual variability: remineralization of sinking POC mirrors NCP, and net annual primary productivity is known to vary interannually depending on the severity of winter mixing in the region.

Next we compare our results with previous studies of the impact of vertical mixing on OUR estimates. In the 100–400 m layer, which includes EDW, Jenkins and Goldman (1985) used the nitrate budget at Station S (Figure 1) to estimate that the downward turbulent flux of photosynthetically produced oxygen in the region produces a 24% error in OUR (5.4 M m$^{-2}$ yr$^{-1}$ versus 4.1 M m$^{-2}$ yr$^{-1}$ ignoring mixing). In contrast, when we integrate from the compensation depth to 400 m using our profiling float-based estimate of OUR in the northern Sargasso Sea, the underestimate is much smaller, introducing only 5% error (6.2 M/m²/yr ignoring mixing). The large difference may stem from our use of the physics-based PV budget to obtain mixing rates versus evaluating the nitrate budget to determine the flux of correlated oxygen.

Ono et al. (2001) apply a constant vertical diffusivity of 1 × 10$^{-4}$ m$^2$·s$^{-1}$ to seasonally averaged oxygen concentrations from six years of BATS sampling, and conclude that vertical mixing of photosynthetically produced oxygen introduces an error of 23% in remineralization estimates, integrated from 100 to 250 m over 240 days (2.08 ± 0.38 M m$^{-2}$, with a contribution of 0.48 ± 0.12 M m$^{-2}$ by removing effects of vertical mixing). In the present study, vertical mixing produces only 9% error in OUR integrated to 250 m depth (2.84 M m$^{-2}$, with a contribution of 0.25 M m$^{-2}$ by removing effects of vertical mixing). The enhanced mixing contribution of Ono et al. (2001) likely results from the application of a uniform vertical eddy diffusivity of 1 × 10$^{-4}$ m$^2$·s$^{-1}$, compared to the seasonally varying vertical diffusivity in this study that tapers with depth.

### 3.3.3. Oxygen and Lateral Processes

Lateral advection and mixing can affect all depths. Because the near surface oxygen cycle is large, and is restricted to densities that outcrop locally in the Sargasso Sea, we assume that lateral processes are far less important than vertical processes in the SOMax and SOMin. However, in the EDW and permanent pycnocline/oxycline layers where oxygen has much smaller vertical gradients, vertical mixing has little impact on OUR (Figure 7b). These deeper layers, which connect to the Gulf Stream and remote winter surface outcrops, are relatively more impacted by lateral processes.

Seasonal oxygen increase in isopycnal layers in the permanent pycnocline beneath the EDW ($\sigma_0 > 26.6$ kg m$^{-3}$, below 450 m) occurs 4–6 months after deep ventilation in the EDW region (Figures 4c and 7a). “OUR” calculated from oxygen rate of change with adjustment for vertical mixing is positive through this layer (Figure 7d), which cannot be due to biological processes. We have already explained that it cannot be due to local ventilation (Section 3.2). Hennon et al. (2016) also observed an increase in oxygen below 500 m during this time of year in subtropical mode water regions. In this EDW region, it could result from westward along-isopycnal advection and mixing from denser winter surface water farther to the northeast and east.
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(Joyce et al., 2013). Hence we suggest that the annual cycle on isopycnals is dominated by lateral processes rather than local remineralization. As our study was focused on the averaged EDW within its confined formation region in the northern Sargasso Sea, and we do not have sufficient oxygen-equipped Argo floats to resolve these pathways, we did not further investigate or quantify these lateral effects, but it would be a useful future study.

3.4. Net Community Production

The oxygen cycle in the upper ocean of the Sargasso Sea is marked by strong seasonality in both photosynthetic production of oxygen and remineralization (Figure 7a). For quantitative comparison of NCP with other studies, as in Section 3.3 we use the mean August profile (Figure 7b) as our best estimate of the overall magnitude of annual OUR. Due to the 6-month differentiation time scale, interpretation of the OUR magnitudes in Figure 7a should be approached with caution when this temporal window overlaps with the mixed layer. The method allows for a sliding and adjustable scale for computing OUR, yielding a qualitative picture of its seasonality, but error is associated with estimates a few months before and after oxygen is introduced via deep winter mixed layers.

Because we do not have a closed oxygen budget, we do not construct the annual NCP integrated from the mixed layer through the seasonal pycnocline as this layer is affected by air-sea gas exchange and respiration as well as photosynthesis. Nevertheless, we can remark that the OUR time series (Figure 7a) is consistent with the Sargasso Sea as a net producer of fixed carbon. The OUR time series shows persistent positive production when the SOMax is capped by the seasonal pycnocline, suggesting positive NCP throughout the year. Mass balance estimates at BATS agree (Emerson, 2014); profiling floats equipped with oxygen sensors show the same in the subtropical North and South Pacific (Riser & Johnson, 2008). The positive NCP layer is strongest and shallowest from May to August, weakening and deepening from August to November (Figure 7c). Maximum magnitude is attributed to the spring bloom (Michaels & Knap, 1996). We attribute the apparent deepening of the maximum production layer to erosion of the seasonal pycnocline and consequent outgassing of supersaturated oxygen that slowly occurs from the top.

Integrating OUR from the compensation depth to the zero crossing at 425 m (Figure 7b) and multiplying by density, we obtain a total remineralization estimate of 4.09 mol O₂ m⁻² over 240 days, which we extrapolate to 6.14 mol O₂ m⁻² over 1 year. The latter corresponds to 4.2 mol C m⁻² export production using 1.45 O₂: 1 Corg (Anderson & Sarmiento, 1994). This integrated estimate is consistent with a mass balance estimate of annual NCP at BATS of 3.8 ± 1.2 mol C m⁻² yr⁻¹ (Emerson, 2014), with the range of estimates at BATS compiled by Fawcett et al. (2018), and with global subtropical NCP estimates, also summarized by Emerson, 2014).

For further comparison with other studies, we integrate from 100 to 250 m to obtain a shallow remineralization estimate of 2.86 mol O₂ m⁻². This compares with estimates by Ono et al. (2001) of 2.08 ± 0.38 mol O₂ m⁻² over 240 days (April to December) and Jenkins and Goldman (1985) 1.7–3.0 mol O₂ m⁻² (April to November) and falls within the range of Sargasso Sea estimates summarized by Sarmiento et al. (1990) of 0.6–3.3 mol O₂ m⁻² annually.

4. Conclusions

In this study we provide a comprehensive survey of northern Sargasso Sea seasonal oxygen structure and utilization using profiling floats, in the region of deep winter mixing that produces the thick EDW. It is the first such survey, to our knowledge. The results are summarized as follows:

1. The seasonality and depth dependence of OUR are consistent with the seasonal maximum production and sinking of POC. Maximum late spring production in the Sargasso Sea above the compensation depth is mirrored by maximum remineralization in the layer just below, with a profile of OUR that decays with depth.

2. A layer of positive NCP near the bottom of the euphotic zone exists from May to November, and possibly year-round, indicating that the Sargasso Sea is a net producer of fixed carbon. The estimated
oxygen production rate, adjusted for vertical mixing, is 0.04 μmol kg\(^{-1}\)·day\(^{-1}\). This supports previous studies at Bermuda (BATS) and other subtropical regions of the global ocean (Emerson, 2014; Riser & Johnson, 2008)

3. Estimated annual remineralization in the SOM\(_\text{in}\) layer is 0.10 μmol kg\(^{-1}\)·day\(^{-1}\), after accounting for vertical mixing, and tapers to 0.04 μmol kg\(^{-1}\)·day\(^{-1}\) in the EDW layer below. Remineralization in May to August is double the rate of August to November

4. Export production over the top 100–250 m is 2.9 mol C m\(^{-2}\), and 4.2 mol C m\(^{-2}\) over the top 425 m. The former is comparable to previous studies near Bermuda while the latter might be an overestimate, although it is consistent with an OUR-based estimate extending down to 2,800 m depth (Jenkins & Doney, 2003). Fawcett et al. (2018) emphasize that OUR-based estimates of carbon production are valid in this region, but that nitrate-based estimates are not, due to the low nutrient levels here, and nonRedfield behavior of nitrate. They hypothesize that the NCP is much more reliant on carbon here than nutrients. The low nutrient concentrations through the upper 400–500 m are directly due to the formation of EDW by deep mixing in late winter (Palter et al., 2005). During a year when such ventilation was greatly weakened, nutrients within the EDW layer did begin to build up (Billheimer & Talley, 2013; Fawcett et al., 2018)

5. The contribution of vertical mixing to estimates of OUR near the compensation depth is substantial: production in the SOM\(_\text{ax}\) layer is larger and remineralization in the SOM\(_\text{in}\) layer is larger than estimates based on oxygen rate of change alone. Vertical mixing between the SOM\(_\text{in}\) and underlying EDW results in a small correction to OUR in the EDW layer, reducing estimates of remineralization. Considerations of vertical mixing also contribute to the seasonal structure of OUR, enhancing production estimates in the SOM\(_\text{ax}\) in late summer. The finding that vertical mixing plays a substantial role in modifying oxygen rates of change may be applicable to other parts of the world upper ocean, particularly regions with large vertical oxygen gradients such as the oxygen minimum zone in the tropics (e.g., Llanillo et al., 2018)

6. We observe a seasonal fluctuation in depth and strength of the SOM\(_\text{ax}\) that coincides with the depth and strength of the seasonal pycnocline, which indicates the prominent role that stratification plays in the seasonality and depth structure of the Sargasso Sea oxygen distribution

7. Beneath 450 m, in the permanent pycnocline underlying the EDW, a summer oxygen increase that coincides with seasonal shoaling of the permanent pycnocline suggests the importance of lateral along-isopycnal oxygen mixing and transport from dense surface outcrops in the eastern Gulf Stream. Vertical mixing has much smaller impact on the permanent pycnocline oxygen balance due to the small vertical oxygen gradient and low vertical diffusivity

8. We show that the oxygen rate of change more accurately represents the oxygen evolution when calculated in depth coordinates shallower than the maximum mixed layer depth, and when calculated in isopycnal coordinates for deeper levels

9. We observe magnitudes of OUR throughout the water column and depth integrated remineralization that fall within the range of previous Sargasso Sea studies that use a wide range of methods

This study does not attempt to close the upper ocean oxygen budget. Our estimates of NCP and remineralization are not balanced, as one would expect them to be on an annual time scale (Brix et al., 2006). We have not considered the influences of mixed layer development and gas exchange that play large roles in the oxygen evolution in the upper 200 m of this region. We also anticipate error in OUR estimates associated with unquantified lateral transport processes.

With expected future sustained deployments of profiling floats equipped with oxygen sensors in this region, this study could be extended to examine the influence of interannual variability of EDW formation on remineralization rate and structure. The large interannual variability of EDW formation has a significant impact on annual net production in the region (Palter et al., 2005), and therefore may influence the remineralization of POC and overall export production on interannual time scales.

### Data Availability Statement

The calibrated CLIMODE profiling float data with oxygen are publicly available through the UCSD digital library (Talley et al., 2020), [https://doi.org/10.6075/J08S4NF2](https://doi.org/10.6075/J08S4NF2). Cruise reports and hydrographic data are also publicly available through the same UCSD digital library resource, and at the Carbon and Climate
Hydrographic Data Office (CCHDO) (https://cchdo.ucsd.edu/search?q=CLIMODE). The Roemmich-Gilson Argo Climatology (Roemmich & Gilson, 2009) is available at http://sio-argodatamgt.org/ARGO_Climatology.html.

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