Hydrographic and dissolved oxygen variability in a seasonal Pacific Northwest estuary

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Abstract

Hypoxia is an issue of growing concern for coastal communities. In the California Current System, a prototypical eastern boundary current, attention has been focused on explaining the trend of increasing shelf hypoxia. Despite the regional focus on hypoxia in eastern boundary regions, relatively few studies have examined smaller estuarine systems. Here, we present results from an observational study in Coos Bay, a small estuary on the southern Oregon coast, subject to seasonal upwelling/downwelling winds, strong tides, and wide fluctuations in freshwater input. Coos Bay exhibits characteristics of a salt-wedge type estuary under high river discharge conditions ($>150 \text{ m}^3 \text{ s}^{-1}$), a well-mixed estuary under low discharge conditions ($0\text{-}30 \text{ m}^3 \text{ s}^{-1}$), and partially-mixed estuary during times of moderate discharge ($30\text{-}150 \text{ m}^3 \text{ s}^{-1}$). The observed vertical stratification and along-estuary salinity gradients correlate significantly with river discharge and tidal forcing. Despite a strong coupling with coastal waters where hypoxia has been present, we do not find evidence for pervasive hypoxia in Coos Bay. We find that upwelling on the shelf advects low dissolved oxygen water into the estuary on synoptic timescales. Early in the upwelling season (April and May), dissolved oxygen minima are found at the estuary mouth, while later in the summer (September), dissolved oxygen minima are found at the riverine end, presumably due to decreased discharge and increased productivity. However, in a given year, the overall strength of the upwelling season is not a good predictor of low dissolved oxygen levels in the estuary.

Keywords dissolved oxygen; hypoxia; estuarine dynamics; upwelling; Pacific Northwest; Coos Bay estuary

1. Introduction

In the past decade, there has been growing concern about the increase in hypoxia on the mid- and inner continental shelf of the California Current System (CCS) (Grantham et al., 2004; Bograd et al., 2008; Chan et al., 2008; Connolly et al. 2010). Hypoxia occurs when waters become undersaturated in dissolved oxygen (DO), causing organisms to suffer adverse and potentially lethal effects (Rabalais et al., 2010). Depending on the effects and the organisms being assessed, thresholds for hypoxia vary widely in the literature (Vaquer-Sunyer and Duarte, 2008; Rabalais et al., 2010). For example, $6.5 \text{ mg O}_2 \text{ L}^{-1}$ is a threshold for oxygen stress in cool water fish (e.g., salmonids) commonly used by regulatory agencies (Brown and Nelson, 2015). For hypoxia, a threshold of $2 \text{ mg O}_2 \text{ L}^{-1}$ is frequently cited, which is what we will use here in defining hypoxic waters.

In the CCS during the summer, equatorward winds drive upwelling of deep, nutrient-rich and oxygen-poor waters onto the outer shelf (Huyer, 1983). Coupled physical and biological processes regulate the DO concentrations of these waters (Monteiro et al., 2006; Bograd et al., 2008; Adams et al., 2013). While outer shelf hypoxia is natural, the recent development of mid- and inner shelf hypoxia is linked to changes in basin-scale atmospheric and oceanic processes that have led to decreases in the oxygen content of upwelled water (Bograd et al., 2008; Chan et al., 2008; Pierce et al., 2012; Peterson et al., 2013), an increase in upwelling-favorable wind stress (Bakun 1990; Snyder et al. 2003), and productivity-driven increases in respiration (Thomas et al., 2003; Grantham et al., 2004). Understanding changes in the conditions of CCS waters has wide implications for other eastern boundary current systems around the world that experience similar dynamics (Epifanio et al., 1983; Chavez and Messié, 2009).

Despite the overall increase in hypoxic area in the CCS, there is significant along-coast variability in the observed incidence of hypoxia (Peterson et al., 2013). On a regional scale, this spatial variability is attributed to wider shelf regions facilitating longer residence times and more organic matter input, thus elevating the potential for the development of hypoxia (Barth et al., 2005; Peterson et al., 2013). Distinct from this regional shelf-scale variability, a latitudinal gradient exists along the CCS, with O'Higgins and Rumrill (2008) reporting that warmer water estuaries in the southern

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4 portion of the CCS (e.g., Elkhorn Slough or the Tijuana River) experience lower DO
5 levels, i.e. more oxygen stress, a higher percentage of the time compared to their
6 northerly counterparts (e.g., South Slough or Padilla Bay).
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10 Considerable attention has been directed towards understanding the drivers of
11 spatiotemporal variability in hypoxia on the shelf (Connolly et al., 2010; Pierce et al.,
12 2012; Adams et al., 2013; Peterson et al., 2013) and in larger estuarine environments like
13 the Columbia River (Roegner et al., 2011) and Hood Canal (Newton et al., 2007) on the
14 US West Coast or the Gulf of Mexico (Rabalais et al., 2002) and Chesapeake Bay (Hagy
15 et al., 2004; Scully, 2013) on the US East Coast. In the Columbia River estuary, low-DO
16 waters entered the estuary on flood tides after upwelling wind events, associated with the
17 advection of higher salinity, nutrient rich oceanic water into the estuary (Roegner et al.,
18 2011). A similar correlation of upwelling winds and low-DO was found in the South
19 Slough, a small arm of the Coos Bay estuary in southern Oregon, although in that case,
20 the evidence suggested that it was high respiration following increased productivity
21 fueled by the nutrient input that caused the DO decrease (O'Higgins and Rumrill, 2008).
22 In general, though, much less attention has been given to the vulnerability of smaller
23 coastal estuarine environments in the CCS (Roegner et al., 2002; O'Higgins and Rumrill,
24 2008; Brown and Power, 2011).
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28 Given the observed along coast variability in DO levels, one might ask which
29 estuaries are most susceptible to direct intrusions of low-DO water from the shelf versus
30 local drawdown of DO driven by nutrient input and the increased productivity and
31 respiration that follows. At a specific location inside an estuary, both processes are likely
32 to occur at different temporal and spatial scales, depending on the time of year, tidal
33 stage, variations in wind and river forcing, and water depth.
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37 Here, we focus on understanding what drives DO variability in Coos Bay, a
38 relatively small estuary on the southern Oregon coast, which is subject to the highly
39 seasonal conditions common throughout the coastal Pacific Northwest. Since no seasonal
40 description of the water properties along the Coos Bay estuary exists, we first focus on
41 identifying the dominant dynamics through a monthly along-estuary hydrographic
42 surveying program coupled with several longer-term time series of water properties. We
43 next investigate variations in DO levels and compare them with the observed
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hydrography over various timescales, including the seasonal, interannual, tidal, and synoptic (weather-driven) range. Finally, we compare our dataset to a historic DO dataset that extends back to the late 1950s in Coos Bay.

2. Study Location

The Coos Bay estuary is mesotidal with mixed semidiurnal tides ranging from 2.3 m at the mouth to 2.2 m at the city of Coos Bay (Rumrill, 2006). It is located south of Heceta Bank, adjacent to a relatively narrow continental shelf (Fig. 1). High sedimentation rates and tidal fluctuations result in large intertidal areas that make up approximately half of the estuary's 54 km² surface area (Hickey and Banas, 2003; Rumrill, 2006). These extensive flats, in conjunction with a deep, dredged navigation channel, produce an ebb-dominant system where flood tides are dampened by friction with the flats and ebb tides rush out the channel (Hyde, 2007). Tidal currents average 1 m s⁻¹ with maximum-recorded currents at 1.7 m s⁻¹ (Baptista, 1989). Based on these tidal speeds, a mean tidal excursion is 14 km, about half the distance from the mouth to the Coos River opening (Fig. 1). However, compared to other small PNW estuaries, such as Yaquina Bay, Willapa Bay, and Grays Harbor, where the tidal prism is ~50% of the estuarine volume, the tidal prism in Coos Bay is only ~30%, suggesting that both tidal and exchange processes might be important to estuarine circulation (Hickey and Banas, 2003).

The estuary has one opening to the Pacific Ocean at its southern end, near the town of Charleston (Fig. 1). The Coos River is the primary source of freshwater input to the system. The majority of previous work has focused on the South Slough, the smaller southern arm (Rumrill, 2006; O'Higgins and Rumrill, 2008) and the site of the South Slough National Estuarine Research Reserve (SSNERR).

3. Data and Methods

3.1. Monthly CTD Transects

To describe the seasonal changes in hydrography and DO levels along the estuary, we conducted monthly sampling over a roughly two-year period. During each sampling cruise we obtained along-channel hydrographic sections of salinity and temperature using

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4 a conductivity/temperature/depth (CTD) sensor. CTD profiles were collected from a 20-
5 foot aluminum boat, the *R/V Pugettia*, of the Oregon Institute of Marine Biology
6 (OIMB). Data used in this study span sampling cruises starting in Nov-2012 and
7 continuing through Jul-2014 (Table 1). We targeted flood tide and high water times in
8 order to sample the expected maximum extent of the salt intrusion into the estuary. Only
9 2 of the 25 sections were taken on ebb tides (Fig. S1). On three cruise dates, we occupied
10 an along-estuary transect from the mouth to the up-estuary end and then back to the
11 mouth (Table 1), allowing us to examine the influence of tidal stage on the observed
12 hydrographic and DO transects. A single transect took on average 1.5 hours to sample
13 with a mean station spacing of 1.5 km (Table 1).
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16 For the majority of the fieldwork we used a RBR Titanium XR-620 profiling
17 CTD with three cable-mounted sensors (Rinko DO, Seapoint Turbidity and Seapoint
18 Fluorometer). The instrument was factory calibrated each year. We sampled at 6 Hz and
19 use downcast data pressure averaged into 1 dbar bins. On a few occasions, a different
20 sensor set up was used (Table 1), including a SeaBird 19plus CTD, and a RBR Concerto
21 CTD with DO (bulkhead-mounted Oxyguard). The SeaBird lacked a DO sensor, so for
22 these cruises no DO data were collected along estuary. For gridding and visualization
23 purposes, we created distance versus depth sections of T , salinity (S), and DO, by linearly
24 interpolating between stations and in depth.
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27 28 29 30 31 32 33 34 35 36 37 38 39 40 **3.2. Water Quality Loggers**

41 The monthly sampling cruises are adequate to resolve large seasonal differences
42 in water properties, but do not give sufficient temporal resolution to resolve tidally-driven
43 variability or synoptic, weather-driven variability. To put our monthly cruises in context,
44 we obtained records from three YSI model 6600 data loggers that measured temperature,
45 salinity, DO, turbidity, and pH. These loggers were deployed 0.5 m off the bottom at 3.1,
46 6.9, and 8.4 km from the mouth and provide time-series of water quality along the estuary
47 (Fig. 1). Two of the loggers are maintained by the Confederated Tribes of the Coos,
48 Lower Umpqua, and Siuslaw (CTCLUSI) water quality-monitoring program, which has
49 been sampling continuously since Oct-2011. One is located at the Empire Docks at
50 43.3942°N, 124.2804°W (EMP; $x = 6.9$ km) in water depth of 6 m, while the other is at
51 the Bureau of Land Management boat ramp at 43.4139°N, 124.2789°W (BLM; $x = 8.1$
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1 km) on the North Spit of Coos Bay in 5 m water depth. The loggers take measurements
2 every 15 minutes and are quality controlled and processed by the water quality managers
3 for each tribe. Here we use data spanning the 2012 and 2013 water years that cover Oct-
4 2012 to Sept-2013 and Oct-2013 to Sept-2014, respectively. The third water quality
5 logger (YSI model 6600) is by the Charleston Bridge at 43.3380°N, 124.3210°W (Fig. 1)
6 and has been maintained by the SSNERR since Apr-2002 in a similar setup to the other
7 YSI loggers. The SSNERR logger is in 2.5 m water depth and at $x = 3.1$ km.
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9 We filter all of the logger data with a one-hour median filter, then subsample to
10 hourly intervals. Gaps in the data are left, i.e., no interpolation is done to fill time periods
11 with missing data.
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13 **3.3. Department of Environmental Quality Data**

14 To put the CTD and recent time series data into historical context, we obtained a
15 fifty-year record (1957-2007) of temperature, salinity, and oxygen in Coos Bay from the
16 Oregon Department of Environmental Quality (DEQ) LASAR database (Brown and
17 Power, 2011). Latitude, longitude, station, date, time, and sampling matrix (surface water
18 or bay/estuary/ocean) were recorded along with DO, salinity, temperature, and percent
19 saturation DO. Unfortunately, however, the depths of each observation were not
20 recorded. DO measurements preceding 1989 were made through Winkler titrations, while
21 post-1989 rapid-pulse polarographic oxygen sensors were used (*pers. comm.*, Larry
22 Caton, OR DEQ).
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24 **3.4. Environmental Conditions**

25 Three main forcing mechanisms, river discharge, tides, and along-shelf winds,
26 control estuarine water properties in small PNW estuaries like Coos Bay (Hickey and
27 Banas 2003). A significant shift in wind conditions along the Oregon coast occurs during
28 the spring transition, when along-shelf winds turn predominantly southward and drive
29 upwelling (e.g., Barth et al. 2007). In the fall, a similarly sharp transition occurs back to
30 the stormy winter months where northward, downwelling favorable wind conditions are
31 common. Although variations in upwelling wind strength do exist across the entire West
32 Coast, observations suggest that conditions over the Oregon coastline are coherent. Thus,
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we use wind conditions taken from a Coastal-Marine Automated Network (C-MAN) station at Newport, OR (Fig. 1), as it represents the most complete time series available. We compared these wind data to those measured offshore at two NOAA buoys: 46015 (located 15 nm west of Port Orford and approximately 75 km southwest of Coos Bay) and 46050 (located 20 nm west of Newport), and found significant correlations at zero time lag in each case ($R^2 > 0.7$). Local wind data from the SSNERR meteorological station (Campbell CR-10X), located in Charleston, would be ideal but these time series have significant gaps during 2012-2013 and do not consistently show northward wind events as the instrument tower is blocked by topography (Rumrill, 2006). The Newport C-MAN station data is low-pass filtered with a 25-hr moving average to remove diurnal variability, subsampled into a daily time series, and rotated into a coordinate system aligned with the local coastline (Barth et al., 2007).

Discharge data were extracted from the Coos Watershed Association (CWA) river gauge network spanning the water years 2002-2013. From these data, total freshwater input to the estuary was estimated by extrapolating discharge data and drainage areas from the major instrumented tributaries (South Fork Coos River, East Fork Millicoma River; West Fork Millicoma River; and Marlow Creek) to the total drainage area ($\sim 1542 \text{ km}^2$) of the Coos watershed. No USGS gauges exist for the time period of this study.

Sea surface height records were obtained from a NOAA tide gauge located in Charleston, OR (station 9432780). We use the verified hourly water levels in the analysis below. Tides in Coos Bay are mixed semi-diurnal, commonly having a lower low water and a higher high water each day (Fig. S1) that results in one strong ebb current and one strong flood current each day.

3.5 Upwelling wind events

The upwelling season on the PNW coast in the CCS is often defined using various combinations of wind conditions and sea surface heights (Logerwell et al., 2003; Van Holmes, 2007), as well as biological conditions (e.g., Hooft and Peterson, 2006). Here we rely on these previous estimates and calculate mean 2012-2013 spring and fall transition dates based on the full range reported at <http://www.cbr.washington.edu/status/trans>. Within each upwelling season, we identify distinct upwelling events based on several

criteria in the Newport C-MAN wind data: (1) the upwelling wind stress must exceed -0.15 N m^{-2} , (2) each identified wind stress peak must be preceded by a relaxation event, where the wind stress $\geq 0 \text{ N m}^{-2}$, and (3) each event must be longer than 1 day and occur within the mean upwelling season dates.

During each upwelling event, we calculate time series of T and S anomalies at each sensor location by subtracting out the time mean T and S over the upwelling event time period. For DO, we use the full value in the following upwelling and tidal variability analysis, in order to examine if hypoxic levels are reached.

4. Results

Temperature, salinity, and dissolved oxygen levels in Coos Bay show remarkable variations on timescales from tidal to interannual (Fig. 2), and are related to wind forcing, river discharge, and tides. River discharge and winds over the water years 2012-2013 (a water year, WY, starts October 1, so that water year 2012 goes from 1-Oct-2011 to 30-Sept-2012) are related to the prevailing atmospheric conditions (Fig. 2). River discharge in WY 2012 ranged from $2\text{-}800 \text{ m}^3 \text{ s}^{-1}$, relatively higher than WY 2013 and with several large springtime peaks in March and April (Fig. 2a). The peak discharges are associated with storm events that have large northward wind stresses (Fig. 2b).

The mean date for the 2012 spring transition was 29-Apr-2012, relatively late compared to other years, including 2013 when the spring transition occurred on 5-Apr-2013. However, the 2012 upwelling season ended later as well, with a mean date for the fall transition of 17-Oct-2012, almost a month later than in 2013 (Fig. 2b). The 2013 upwelling season was marked by two strong relaxation events where the wind stress, $\tau > 0 \text{ N m}^{-2}$ for several consecutive days, reducing the cumulative upwelling strength over that year's season. There were 11 distinct upwelling events in 2012, compared with only 9 in 2013 (Fig. 2b), although the average magnitude and duration of individual events in each year were similar.

4.1. Seasonal hydrography and dissolved oxygen variability

Salinity profiles from the CTD transects show the remarkable seasonality in Coos Bay (Fig. 3), and are representative of conditions during each season. The variability in

the along-estuary salinity gradient and stratification reveal a system shifting seasonally through different estuarine classifications. In the fall (Fig. 3a), the estuary is partially mixed—isohalines are slightly tilted, and considerable freshwater is present, mixing along the length of the estuary. Wintertime conditions in Coos Bay cause the estuary to become a salt wedge. Fig. 3b shows the much fresher, and much more stratified salinity section. Isohalines are nearly horizontal as large inputs of freshwater drive the salt intrusion down-estuary. Springtime conditions are largely variable (Fig. 3c), dependent on the freshwater input and wind conditions. The April 2013 section shown was taken following an upwelling-favorable wind event on the shelf. The depth-averaged along-estuary salinity gradient is large, although the isohalines are nearly vertical and a plug of saltier water was observed in the lower layer at the mouth. Despite the strong along-estuary gradient, stratification was weak, and the estuary was well-mixed. Summertime conditions in Coos Bay are well-mixed, especially by late summer in September (Fig 3d). The salt intrusion extended to the limit of our observations, with the freshest water having $S > 30$ and vertical isohalines.

The CTD data also show a large seasonal shift in observed DO levels along the Coos Bay estuary (Fig. 3). In fall and winter months, DO levels are relatively high (>8 mg L⁻¹), with small along estuary gradients (Fig. 3a-b). During spring and summer, however, DO levels decrease everywhere in the estuary (Fig. 3c-d), falling close to 5 mg L⁻¹. In the spring (Fig. 3c), the lowest DO levels are found at the mouth, associated with colder, saltier ocean water, while in September at the end of the dry, upwelling season, the lowest DO waters are found at the riverine end (Fig. 3d).

Time series data from the loggers located within the estuary support the seasonality revealed by the CTD sections (Fig. 2c-e). During the wet fall and winter months, salinity varies widely from 3-30 psu, in tandem with high discharge events. Salinity gradually increases from spring through the summer, reaching up to 32 psu by Sept-Oct with much lower variability (Fig. 2d). In contrast, temperatures show relatively little variability during the fall and winter months, when both ocean and river end-members are cold (Fig. 2e). As summer progresses, temperatures increase as does the variability, most likely due to the interplay between warmer riverine-influenced waters and colder, upwelled waters coming from the mouth. The BLM logger recorded the

lowest salinities, as it is located farthest up-estuary (Fig. 1), although all three locations showed similar seasonal variability in T and S . Higher frequency variability on tidal and synoptic timescales is discussed below in more detail.

DO levels also exhibit a seasonal cycle that supports the representative CTD sections shown in Fig. 3. Highest overall levels of DO were observed in Dec-Jan of each year (Fig. 2c), although intermittent maxima during the upwelling season also occurred. The upwelling season was marked by the lowest recorded DO levels at each location (Fig. 2c), although the variability of DO levels increased during the dry season. In 2012-2013, we found no record of hypoxia in Coos Bay, either in the time series data (Fig. 2c) or CTD data (Fig. 3). However, $\text{DO} < 6.5 \text{ mg L}^{-1}$ throughout the upwelling season, and in 2013, $\text{DO} < 5 \text{ mg L}^{-1}$ at the end of the dry season in September (Fig. 2c). The EMP and SS loggers recorded the lowest DO values, although the seasonal variability was similar, much like the observed T and S .

Fig. 4 illustrates the along-estuary variation in salinity and stratification over the seasons using all the monthly CTD data. The depth-averaged, along-estuary salinities support the general picture obtained by examining the full S transects with stronger gradients in the winter months coinciding with the freshest waters observed (Fig. 4a). Previous studies suggest functional dependencies of both the vertical stratification and the along-estuary salinity gradient with the river discharge (Monismith et al., 2002; Ralston et al., 2008; MacCready and Geyer, 2010). There is a slight dependence of the along estuary salinity gradient ($\partial S / \partial x$) on Q_r , the river discharge (Fig. 4b), where we have taken the section mean for each monthly transect to get a section averaged value of $\partial S / \partial x$. Q_r is calculated as the mean discharge over the 4 days preceding the CTD transect. We find that $\partial S / \partial x \sim Q_r^{0.19}$ ($R^2 = 0.90$, $p < 0.01$), which is slightly higher than the power of $1/7$ (0.14) estimated previously (Monismith et al., 2002; Ralston et al., 2008), but much lower than that predicted by classic estuarine theory (MacCready and Geyer, 2010). The correlation does not change significantly if $Q_r^{1/7}$ is used. However, at the highest Q_r , there is a hint of saturation in $\partial S / \partial x$, leveling off near 1 psu km^{-1} .

The along-estuary vertical stratification, $\partial S / \partial z$, also changes seasonally: during summer and fall months, the stratification is low throughout the estuary, increasing into the winter and spring. There is no persistent spatial trend in $\partial S / \partial z$, although it does

increase slightly up estuary in certain winter and spring months. However, this bias might reflect the calculation method of $\partial S/\partial z$ as the total depth (Δz) decreases as one approaches the Coos River and exits the dredged channel (Fig. 1). Here, we define the stratification as $\Delta S/\Delta z$, where ΔS is the difference between S over the upper 2 m and the lower 2 m of the water column (Fig. 4c). For vertical stratification, $\partial S/\partial z$, theory predicts it will increase as $\sim Q_r^{2/3}$ (MacCready and Geyer, 2010). We find that the change in stratification is well explained by a power law dependence on river discharge, such that the measured $\partial S/\partial x \sim Q_r^m$ with $m = 0.67$ ($R^2 = 0.84$, $p < 0.01$).

4.2. Interannual variability in hydrography and DO

Although a similar seasonal cycle appears in both 2012 and 2013, the T - S data reveal clear variation between the two years (Fig. 2). In 2012, salinity ranged from 2.5-33.9 and temperature ranged from 5.9-17.6°C. The influence of freshwater did not start to dominate until February, corresponding to the later onset of large discharge events to the estuary that year (Fig. 3). In 2013, conditions were warmer and saltier overall at the same logger location, with the minimum $S = 10.8$, the maximum $S = 34.6$, and temperatures from 6.9-20.4°C. In contrast to 2012, the coldest waters in 2013 were observed in January instead of during an upwelling event like in April 2012. January 2013 experienced a larger range in salinity than in 2012 due to some large, early season discharge events. However, for all other months, the salinity range in 2013 was much reduced from that of 2012. Diminished discharge in the mid to late wet season of 2013 might account for the disparity in the observed salinity range.

The weaker freshwater inflow to the estuary in 2013 likely caused a reduction in the estuarine residual flow and an increase in the extent of the salt intrusion, affecting the local T - S properties at the EMP and BLM locations. Coupled with the overall weaker summer upwelling season in 2013 (Fig. 2a), the water temperatures were higher in Coos Bay that year. The 2013 upwelling season started and ended earlier than 2012 based on every available estimate of transition dates on the OR coast, and easily seen in the wind data (Fig. 2a). No matter the upwelling strength, a smaller residual flow would prevent the coldest upwelled summer waters from reaching the logger, and also would facilitate longer residence times in the estuary.

The seasonal occurrence of lower DO waters in an estuary, such as Coos Bay, is generally due to a combination of physical (advection and residence time) and biologic (*in situ* respiration; Rumrill, 2006) processes. However, similar to the observed hydrographic variability shown here for 2012 and 2013, and the interannual variability observed previously in shelf DO (Grantham et al., 2004; Chan et al., 2008), we find interannual changes in estuarine DO levels in Coos Bay. In 2012, waters were well oxygenated, with few values below the 4.6 mg L⁻¹ threshold found to maintain 90% of all species (Vaquer and Duarte, 2008) and none below the 2 mg L⁻¹ threshold (Fig. 2c). Late in the dry season of 2013, there were more measurements of waters below 4.6 mg L⁻¹, but still none below the 2 mg L⁻¹ threshold. The lower DO values in Sept-2013 correspond to an overall decreasing linear trend over the 2013 upwelling season, compared to no trend in 2012. Despite a stronger, longer, and more consistent upwelling season in 2012 (Fig. 2) that incorporated cold, salty shelf waters into the estuary, there were fewer instances of lower DO. The combined effect of upwelling wind events and tides is discussed below.

4.3. Synoptic variability

The advection of shelf waters into Coos Bay can cause *T*, *S*, and DO variability on synoptic, or weather, timescales typically on the order of 2-7 days. In the high discharge season, shelf waters can enter the estuary both through tidal dispersion processes and an advective pathway driven by the estuarine exchange flow (MacCready and Geyer, 2010). When discharge goes towards zero in the dry summer months, however, tides become the main mechanism for shelf waters to enter Coos Bay (Hickey and Banas, 2003).

Fig. 5 shows an example from April 2013 of how upwelling favorable winds on the shelf allowed an inflow of relatively low DO into the estuary. The onset of the wind event on April 22nd (*t* = 0 days) resulted in minimum DO two days later. During this period, DO dropped by >2 mg L⁻¹ at both the South Slough and BLM stations (Fig. 5a). *T* and *S* changed correspondingly as well, with *S* increasing and *T* decreasing over the event, as colder, saltier waters from the shelf entered the estuary (Fig. 5d). We show all time series of DO, if available, but only the *T* and *S* anomalies from the BLM location as the other sensor locations showed very similar variability for those water properties. Unfortunately, the EMP DO sensor was malfunctioning during this particular event.

These time series capture the sequence of the movement of upwelled water into Coos Bay, likely explaining the strong along-estuary gradient in DO observed in the CTD transect in Apr-2013 (Fig 3). In the Apr-2013 event, T decreased and S increased in concert with the swing in DO, with maximum S and minimum T occurring on flood tides (Fig. 5a,d). With the large spatial DO gradients, tidal currents can induce swings of 3-4 mg L⁻¹ in DO levels at a single location (Fig. 5a). Thus, we expect that both tides and the residual circulation play a role in explaining the observed T , S , and DO trends during this springtime event. Tidal variability is discussed further in the next section.

An upwelling wind event does not always result in a slow decrease in DO, however. In late Aug-2012, the DO levels recorded at all locations show variability on tidal timescales (Fig. 5c), but no longer term decrease as observed in Apr-2013. In general, upwelling events earlier in the season (i.e., April and May) show the largest relative changes in DO during and a few days after the onset of the upwelling winds (e.g., Fig. 3a). We identify these events as Type 1, accounting for 63% of the events in 2012 (7 out of 11 events) and 67% of the events in 2013 (6 out of 9 events). Type 3 events typically occur later in the summer, and show a similar variability on tidal time scales, where minimum DO occurs at the end of flood tides (discussed below), but without the longer time period drawdown (Fig. 5c). No significant trend in T or S over the upwelling event time period was observed during Type 3 events either (Fig. 5f). Only one event in each year was classified as Type 3. In contrast, Type 2 events showed a longer term drawdown of DO, similar to Type 1, but had the lowest DO values at the end of ebb tides, i.e. at low tide (Fig. 5b). A Sept-9-2013 event illustrates the Type 2 signal, with DO decreasing by 2-3 mg L⁻¹ after the onset of upwelling winds, but with the lowest DO values found at low tide. Type 2 events occurred two times in both 2012 and 2013, during the latter part of the season. Interestingly, in Type 2 events, the upwelling signal was still apparent over the two day time period, with a small decreasing trend in T and increasing trend in S observed. Thus, the difference between Type 1 and Type 2 is where the lowest DO waters are sourced, apparently coming from up-estuary in Type 2 events and from the shelf during Type 1 events.

4.4. Tidal variability

When large horizontal gradients exist in T , S , and DO, tidal currents can induce large variability at a fixed location inside the estuary. The sign of the T , S , or DO gradient determines the correlation between tidal stage and the observed value, with the magnitude of the change also affected by the tidal range (spring vs. neap) and the magnitude of the gradient itself. In Coos Bay, we find that the M_2 semi-diurnal constituent dominates tidal variability in T , S , and DO at all logger locations (not shown), with significant diurnal variations present in the DO time series.

The three examples used to show synoptic variability during the upwelling season also demonstrate the range of tidal variations. In all cases, flood tides bring relatively salty and cold water from the mouth into the estuary, with ebb tides taking that water back out. This results in a positive relationship between S and sea surface height, H , at each logger location, and an inverse relationship between T and H (Fig. 5d-f).

For DO, the tidal variability is more complicated, as biological processes can also play a role in modulating DO levels in both the shelf and estuarine systems. During Type 1 and Type 3 events, we find that flood tides and high water are associated with lower DO values, suggesting that oceanic waters are relatively less saturated in DO (Fig. 5a,c). For Type 2 events, we find that the lowest DO values are recorded during ebb tides and low water (Fig. 5b), suggesting that the up-estuary waters are relatively less saturated.

Fig. 6 shows the effect of tidal advection on the along-estuary transects, from the three sampling cruises where two transects were taken consecutively (Fig. S1). During the late flood tide of 27-Apr-2013, we observed the advection of low DO and high S waters into the estuary (Fig. 6a,d). Minimum DO values stayed roughly constant at 5 mg L^{-1} and were associated with maxima in $S \sim 34$. Over the 1.4 hours between transects, the mean S and minimum DO along-estuary profiles moved roughly 4.5 km landward, equivalent to a speed of 0.9 $m s^{-1}$. This speed is a reasonable estimate of tidal current speeds in Coos Bay based on previous data (Hickey and Banas, 2003; Hyde, 2007). A similar advection of water properties from the shelf into the estuary was observed on 26-Mar-2013 during two transects taken during the middle and end of a flood tide (Fig. 6b,e). The gradients in these transects were relatively smaller than during Apr-2013, but we estimate a similar tidal speed from the data. Finally, during winter (Fig. 6c,f), the DO horizontal gradient was observed to be very small, so tidal advection had little effect on

the observed profiles of minimum DO, although slightly saltier water was advected in on the latter part of the flood tide (Fig. 6c). These CTD transect data support the tidal variability observed in the time series at BLM, EMP, and SS, which show how the tidal advection of gradients is imprinted on top of seasonal and interannual variability.

5. Discussion

5.1. Hydrographic variability and estuarine dynamics

Physical processes set the ambient DO conditions in the estuary, while biological processes may modulate DO within the estuary. Rapid decreases of DO in Coos Bay can be linked to upwelling-favorable wind events on the shelf during early summer (Figs. 2, 5). The inundation of the estuary with predominantly salty, oceanic water through the dry season (Figs 2, 5) confirmed that shelf waters are key in setting the conditions for the estuary, especially when freshwater flow is virtually nonexistent. Coos Bay is a seasonal estuary, transitioning from a strongly stratified, salt-wedge type regime in winter during storms, to a well-mixed, oceanic-influenced type in the summer. These observations are consistent with previous studies that focused on the South Slough alone (Rumrill, 2006; O'Higgins and Rumrill, 2008), which can be thought of as a smaller analogy to the Coos Bay system. However, one large difference between the South Slough and the greater Coos Bay estuary is in the greater role that estuarine exchange flow plays in Coos Bay. With a deep, dredged channel, Coos Bay has a relatively small tidal prism compared to other PNW estuaries (e.g., Grays Harbor, Willapa Bay, Yaquina Bay) outside of the Columbia River (Hickey and Banas, 2003), as well as a larger freshwater source. Thus, although we do observe strong tidal variations at sensor locations located within a tidal excursion of the mouth, we also see strong variability linked to discharge in the wetter months. In the South Slough, the overall shallower depths and lower freshwater input favor tidal dispersion processes over an advective exchange flow (MacCready and Geyer, 2010), for the purpose of flushing the estuary.

The implication of this dependence on estuarine exchange flow, as well as tidal flushing, for Coos Bay is a stronger dependence on discharge than other relatively small PNW estuaries. The difference in water properties observed in 2012 and 2013 support this notion; a wetter 2012 led to a fresher, and potentially more connected estuarine/shelf

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4 system than in 2013, which was relatively dry. In 2013, the smaller freshwater input
5 would lead to a smaller exchange flow by decreasing the along-estuary salinity gradient,
6 and thus, relatively less flushing of the estuary that would partially explain the warmer
7 temperatures observed that year.
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11 We further corroborate the observational classification of Coos Bay from the
12 CTD transects by quantifying the Simpson number, Si , also referred to as the horizontal
13 Richardson number. Si is the ratio of tidal mixing strength to the estuarine circulation
14 strength (Stacey et al., 2001). Here, $Si = (g\beta h_0^2 \partial S / \partial x) / (C_D u_t^2)$, where $\partial S / \partial x$ is the depth-
15 averaged, along-estuary salinity gradient, β is the coefficient of expansivity for salinity
16 ($7.7 \times 10^{-4} \text{ psu}^{-1}$), g is the gravitational constant 9.8 m s^{-2} , h_0 is water depth, u_t is a
17 representative tidal velocity, and $C_D = 2 \times 10^{-3}$ is a bottom drag coefficient (MacCready
18 and Geyer, 2010). The CTD data provided $\partial S / \partial x$ (Fig. 4), and tidal velocities and water
19 depth were approximated at 1 m s^{-1} and 10 m, respectively.
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28 We find that $Si = 0.25\text{-}0.36$ for the fall months (Oct. to Dec.), $Si = 0.36\text{-}0.53$ for
29 the winter months (Jan. to Mar.), $Si = 0.36\text{-}0.61$ for the spring months (Apr. to Jun.), and
30 $Si = 0.09\text{-}0.30$ for the summer (Jul. to Sep.). Higher Si indicates stronger stratification
31 that inhibits mixing, and the resultant increased current shear facilitates stronger
32 baroclinic flow (Stacey et al., 2001). Low values of Si , which occurred more often in the
33 summer in Coos Bay, indicate that tidal mixing dominates. This supports the well-mixed
34 conditions observed in summer, when the baroclinic forcing is weak and tidal mixing
35 occurs more easily. The Si value for September was the lowest, falling under the
36 threshold where well-mixed conditions dominate, $Si < 0.1$ (Simpson et al., 1990;
37 MacCready and Geyer, 2010). Higher values of Si in the winter and spring months
38 indicated the potential for stronger baroclinic flow.
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50 **5.2. DO variability and link to estuarine dynamics**

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52 Dissolved oxygen levels were lower overall in the dry season than in the wet
53 season (Fig. 7), a result that indicates either 1) water properties in Coos Bay were
54 strongly coupled to shelf processes where a similar temporal signal is observed, or 2)
55 biological processes drew down ambient DO levels during summer. The coupling
56 between shelf waters and the estuarine waters has been previously documented for Coos
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1 Bay (Czielsa, 1999; Roegner and Shanks, 2001), and other, similar small Pacific
2 Northwest estuaries (Roegner et al., 2002; Brown and Ozretich, 2009; Brown and Power,
3 2011). However, the higher incidence of lower DO levels late in the dry season coincides
4 with longer residence times as the dry season progresses, potentially allowing biological
5 processes to decrease DO. Quite likely, both processes are important; though quantifying
6 the biological component of this DO drawdown is beyond the scope of this study. Fig. 7
7 shows that in general, the lowest DO waters observed in the CTD transects along the
8 main channel coincide with the warmest waters, e.g., in September (Fig. 7b). These
9 conditions are favorable to create the Type II upwelling response seen in the logger data
10 discussed above. The relatively low DO waters generally are found at higher salinities,
11 although the minimum DO levels are not found at maximum *S* (Fig. 7c). Instead, there
12 appears to be DO minima at intermediate salinities, both in the CTD and DEQ data.

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26 The exception to the DO-*T* relationship occurs in April of both years (Fig. 7),
27 where low DO waters are observed near the mouth, with a strong along-estuary gradient.
28 The minimum in DO in these April transects coincides with the observed occurrence of
29 colder and saltier upwelled water, the Type I upwelling response discussed above, in
30 contrast to the other transects. Thus, relatively “younger” upwelled water, i.e., from early
31 in the season, may be the oceanic source of DO minima. As the upwelling season
32 progresses, the “older” upwelled water can increase in DO due to photosynthesis and can
33 be blocked from entering the estuary unmixed due to a decreased estuarine exchange
34 flow.

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42 Lower DO waters were also observed as the dry season progressed (Table 2; Fig.
43 7). The location of the minima in DO levels along the estuary migrated from the mouth in
44 the spring to the head of the estuary in the summer (Fig. 7). This spatiotemporal shift is
45 due to upwelled shelf water spending longer in the estuary. Increased residence times in
46 Coos Bay in the dry season were facilitated by weakened residual flow resulting from
47 diminished discharge lowering the buoyancy forcing to the estuary. Decreased estuarine
48 exchange flow in late-summer inhibits the advection of low DO shelf waters into the
49 estuary; upwelled waters are observed early in the summer after the spring transition, but
50 not in late summer. Thus, the low DO observed in Coos Bay in late summer must occur
51 due to local processes—when waters spend more time in the estuary, they are subject to

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4 increased biologic respiration that draws down DO levels. Further research on the
5 spatiotemporal variability in net ecosystem metabolism in Coos Bay would serve to
6 corroborate this finding.
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10 To give historical context to the observed occurrence of low DO in Coos Bay, we
11 examined the DEQ dataset that spans 1957-2007 and covers a large portion of the estuary
12 (Fig. S2, *inset*). Limiting the DEQ data to only the dry season months, we find only one
13 measurement of DO $< 2 \text{ mg L}^{-1}$ (Fig. 7), with the majority of all observations $> 5 \text{ mg L}^{-1}$.
14 The one low DO measurement appears cold and salty relative to all the other data, and is
15 found near the mouth of the estuary at $x = 3 \text{ km}$, suggesting it is derived from upwelled
16 waters, much like the April CTD data (Fig. 7). The general trend in DEQ DO levels
17 shows that the lowest DO levels are found at the warmest temperatures and at $S \sim 24\text{-}28$.
18 These $T\text{-}S$ properties only occur in the riverine end of the estuary and in the summer
19 months, when surface heat fluxes and relatively warm river water contribute to the high
20 temperatures and relatively lower salinities (compared to oceanic values where $S > 30$).
21 Unfortunately, while the DEQ data are temporally robust, they lack any information
22 about sample depths, making it difficult to draw conclusions about the historic
23 occurrence of hypoxia in the estuary with these data alone. In addition, sampling was
24 widely inconsistent between years (Fig. S2); some years had less than a dozen samples,
25 while others had hundreds. Also note that the majority of the samples were located far
26 from the mouth, with no samples taken at along-estuary distances $x < 3 \text{ km}$ from the
27 ocean. Nonetheless, combined with the more recent water quality data discussed above,
28 these data suggest that pervasive hypoxia has not occurred in the main channel of Coos
29 Bay over the last 50 years.
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46 The two-year data record from the CTCLUSI and the eleven-year record from the
47 SSNERR show that although near bottom DO levels diminished as the dry season
48 progressed, the majority of the time ($\sim 90\%$ in the dry season), the waters are not hypoxic
49 (Table 2). September had the maximum occurrence of low DO levels at all locations,
50 although the EMP logger was significantly higher than the other two. It is likely lower
51 than the BLM logger because of its closer proximity to the mouth. Thus it might receive a
52 greater proportion of low DO upwelled water (Fig. 7).
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4 Interannually, it is likely that the DO values $<4.6 \text{ mg L}^{-1}$ observed in 2013 can be
5 attributed to the diminished residual flow in that year facilitating longer residence times
6 and allowing *in situ* biologic DO drawdown. Additionally, the estuarine water
7 temperatures in the summer of 2013 were higher than in 2012 (Fig. 2), facilitating
8 increased biologic oxygen demand and decreased solubility of oxygen across the air-
9 water interface.
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12 The data in this analysis show that hypoxia is not observed in the main channel of
13 Coos Bay, and has not been for the past decade and possibly longer. DO conditions prior
14 to 1957 need to be assessed further, possibly through the use of paleo-oxygenation
15 proxies (Goody et al., 2009).
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18 The absence of hypoxic waters in the main channel of Coos Bay is in part due to
19 the well-mixed conditions in summer (Fig. 2). Unlike other Pacific Northwest estuaries
20 that experience intermittent hypoxia—the Columbia River (Roegner et al., 2011) which
21 has year-round moderate to high discharge causing strong stratification, and Hood Canal
22 (Newton et al. 2007), which is a deep fjord-like estuary—Coos Bay does not have a
23 strong enough summer freshwater inflow nor is it deep enough to cause stratification that
24 prevents mixing of aerated surface waters with deeper, DO-depleted waters. The physical
25 processes affecting hypoxia in estuaries that face eutrophication issues, such as in the
26 Gulf of Mexico (Rabalais et al., 2002) or even many of Denmark's small estuarine
27 systems (Conley et al., 2012), are different than those found here. For example, strong
28 stratification often inhibits vertical mixing and enables DO levels to decrease further,
29 with re-oxygenation only occurring during strong wind events that mix the water column
30 (Scully, 2013). The well-mixed conditions and weak exchange flow in Coos Bay in late
31 summer implies that, even if hypoxic shelf waters existed offshore, they could only enter
32 the estuary through tidal dispersion processes. Indeed, previous work in Yaquina Bay, a
33 slightly smaller estuary north of Coos Bay, shows similar flood-tide propagation of low-
34 DO waters into the estuary (Brown and Power, 2011).
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37 However, the waters observed entering Coos Bay from the shelf were not
38 hypoxic. Along-shelf variability in the occurrence of inner-shelf hypoxia could explain
39 this finding (Peterson et al., 2011; Send and Nam, 2012). Regions in the CCS with
40 narrow continental shelves experience less hypoxia than areas with wider shelves
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(Peterson et al., 2011). Upwelling on a wide shelf occurs in a relatively deeper depth range than over a shallow shelf, given the same wind-forcing and shelf bathymetry. The deeper depth range limits the re-oxygenation, due to photosynthesis or wind and wave driven mixing, of the naturally low-DO upwelled waters (Send and Nam, 2012). On a narrow shelf, such as near Coos Bay, upwelled water reaches shallower depths where water parcels would benefit from oxygen introduced by photosynthetic organisms and wind-driven mixing processes.

While hypoxia is not yet a problem in the main channel of Coos Bay and the South Slough, monitoring should continue as the effects of climate change advance, as many eastern boundary current systems will see conditions move to being more favorable for increased inner-shelf hypoxia (Chavez and Messié, 2009). Increased surface-ocean warming, projected under the Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report AR5 (Rhein et al., 2013), will enhance stratification and decrease mixing. Upwelling winds in the CCS are also intensifying, causing a greater volume of water to be upwelled onto the shelf (Bakun, 1990). The DO content of these waters is decreasing as the OMZ shoals in the CCS (Bograd et al., 2008; Pierce et al., 2012).

In tandem with the coastal effects of climate change on hypoxia, alterations to Coos Bay's hydrography should be expected under future climate change. An analysis of climate models from the IPCC showed the PNW experiencing little change in mean precipitation in the coming century, however, the winters are projected to be wetter and the summers, drier (Mote and Salathé, 2010). Less precipitation later in the wet season and into the dry season could alter the estuarine circulation by further reducing the along-estuary salinity gradient, analogous to the differences discussed above between the 2012 (drier, lower DO) and 2013 (wetter, higher DO) water years. This will cause longer residence times, and presumably, higher susceptibility to biologically-driven hypoxia.

6. Conclusions

Coos Bay is a strongly seasonal system as evidenced by the *T-S* properties of water in the main channel of the estuary. Most of the seasonality can be linked to variable freshwater inflow to the estuary and coastal shelf processes. The seasonality of Coos Bay results in large swings in its estuarine parameters and circulation, so much so that Coos

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4 Bay can be classified into a seasonally-shifting pattern of estuarine regimes, from
5 strongly stratified during episodic wintertime storms to well-mixed during the
6 summertime dry months. During summer, tidal processes dominate exchange with the
7 ocean, while during other months tides induce large variations in local water properties,
8 e.g., 3-4 mg L⁻¹ swings in DO levels.
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11 The seasonal variability has implications for dissolved oxygen levels and the
12 health of the estuary. While stratification is strong under high discharge conditions in the
13 winter, colder, well-oxygenated river waters, and downwelling conditions on the shelf
14 promote high DO levels. In the summer, discharge diminishes and residual flows
15 stagnate, causing long residence times in the estuary. Upwelled shelf waters in the
16 estuary are relatively low in DO, yet stronger upwelling over a season did not result in
17 lower DO levels overall in Coos Bay. Despite recent alarm about inner-shelf hypoxia on
18 the Oregon coast, there appears to be no hypoxia in the main channel of Coos Bay
19 currently, and there is little evidence of past hypoxia in the estuary.
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Table 1. Information on along-estuary transects, including date of observations, CTD instrument used, average tidal stage during transect^a (and average # of hours past LLW in parentheses), number of casts, whether dissolved oxygen was measured, and what part of the spring/neap cycle (S/N) was occurring.

#	Date	Instrument	Tidal stage (hours ^a)	# Casts	Oxygen	Tides
1	11-3-12	RBR XR-620	flood (13.0)	22	Y	S
2	1-19-13a	RBR XR-620	flood (1.5)	16	Y	N
3	1-19-13b	RBR XR-620	flood (3.1)	17	Y	N
4	2-21-13	RBR XR-620	low slack (0.5)	9	Y	S
5	3-12-13	RBR XR-620	ebb (7.6)	11	Y	S
6	3-26-13a	RBR XR-620	flood (4.8)	16	Y	S
7	3-26-13b	RBR XR-620	high slack (6.2)	16	Y	S
8	4-27-13a	RBR XR-620	flood (3.9)	16	Y	S
9	4-27-13b	RBR XR-620	flood (5.3)	16	Y	S
10	5-19-13	RBR XR-620	ebb (22.1)	19	Y	N
11	7-11-13	SBE 19plus	ebb (23.4)	16	N	S
12	8-16-13	YSI CastAway	low slack (0.9)	17	N	S
13	9-7-13	YSI CastAway	flood (13.9)	16	N	S
14	9-17-13	RBR XR-620	flood (3.5)	11	Y	S
15	9-21-13	RBR XR-620	flood (13.9)	13	Y	S
16	10-16-13	SBE 19plus	flood (5.6)	8	N	S
17	10-20-13	RBR XR-620	high slack (17.9)	14	Y	S
18	11-30-13	RBR Concerto	high slack (17.8)	14	Y	S
19	1-25-14	RBR Concerto	low slack (0.8)	17	Y	N
20	2-22-14	RBR XR-620	low slack (0.8)	21	Y	N
21	3-18-14	RBR XR-620	flood (5.7)	11	Y	S
22	4-6-14	RBR XR-620	low slack (1.0)	22	Y	N
23	5-13-14	RBR Concerto	flood (4.1)	19	N	S
24	6-17-14	RBR Concerto	flood (2.1)	13	Y	S
25	7-24-14	RBR Concerto	high slack (6.9)	17	Y	S

^a: This is the time in hours since lower low water (LLW) at t = 0 hrs. One complete tidal cycle is 24.84 hours for a semidiurnal M₂ tide that dominates Coos Bay (see Fig. S1).

Table 2. Percentage of total number of DO measurements <6.5 mg L⁻¹ for each logger location. Dates for BLM and EMP span Oct-2011 to Sept-2013, while the SSNERR (SS) Charleston logger spans Apr-2002 to Dec-2013.

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
SS	0	0	0	2	2	4	6	10	10	9	1	1
EMP	1	0	0	0	<1	3	7	4	38	0	0	0
BLM	0	0	0	2	<1	2	1	3	6	0	0	0

FIGURE CAPTIONS

Figure 1. Map of Coos Bay with physical features and bathymetry contoured. Distance along-estuary (x) is numbered (km), with select sensor positions indicated. Inset shows Coos Bay's location along with other PNW estuaries, as well as Heceta Bank (HB) offshore and the location the Newport C-MAN meteorological station and NOAA buoys 46015 and 46050.

Figure 2. (a) Daily river discharge, Q ($\text{m}^3 \text{s}^{-1}$), for the Coos River watershed over water years 2012 and 2013 (Oct-2011 to Nov-2013). Black vertical lines divide the calendar years. (b) Daily alongshore wind stress, τ (N m^{-2}), measured at the Newport C-MAN station. Green ticks indicate upwelling events, while gray shading indicates time span of upwelling season. (c) Dissolved oxygen (DO, mg L^{-1}) from the BLM, EMP, and SSNERR Charleston Bridge sensor locations. The yellow and orange horizontal dashed lines are various cited thresholds (see text). (d) Same as c but for salinity. (e) Same as c but for temperature.

Figure 3. (a) Along-estuary salinity section (color) in the channel during the fall months, starting at the mouth ($x = 0$ km) and ending near the Coos River ($x = 22$ km). Black lines are dissolved oxygen contours every 0.5 mg L^{-1} . Black triangles show the locations of CTD casts. (b) Same as in a, but for a representative wintertime transect. (c) Same as in a, but for a representative springtime transect. (d) Same as in a, but for a representative late summer transect.

Figure 4. (a) Depth-averaged salinity (S) normalized by the observed salinity at the mouth (S_{mouth}) as a function of along-estuary distance for the monthly transects (colored by season). (b) The mean value of the along-estuary salinity gradient for each monthly section as a function of river discharge, Q_r . The black line shows a power law fit. (c) Vertical stratification as a function of along-estuary distance for all of the monthly transects (color). (d) The mean vertical stratification for each monthly section as a function of river discharge, Q_r . The black line shows a power law fit.

Figure 5. (a) Type 1 upwelling response of dissolved oxygen (DO, mg L^{-1}) from the Empire (red), BLM (blue), and SSNERR (gray) Charleston Bridge sensors, and sea surface height, H (m), from the Charleston tide gauge (black) as a function of time in days since an upwelling event. Gray shading indicates timing of upwelling event. Note H is offset by 6 m for plotting purposes. (b-c) Same as in a, but for events starting on dates indicated. (d) T and S anomalies from the BLM sensor over the same upwelling period in a. Note that here H (black line) is offset by -3 m. (e-f) Same as d, but corresponding to time periods shown in b and c, respectively.

Figure 6. (a-c) Mean along-estuary salinity from surveys taken on the same day, but hours apart. Hours past LLW is indicated in parentheses. (d-f) Same as a-c, but showing the minimum along-estuary DO measured.

Figure 7. (a) Minimum dissolved oxygen observed during each monthly transect versus distance along the estuary ($x = 0$ km is at the mouth), colored by month. (b) Dissolved oxygen versus temperature observations from all monthly transect data (colors same as in a) and the OR DEQ historic dataset (pink dots). (c) Same as in b but for dissolved oxygen versus salinity.

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Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7