

RESEARCH ARTICLE

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Key Points:

- Mid-water oxygen minima are studied using both observation and simulation
- The buoyancy-driven convergence is essential in forming the oxygen minima
- The mid-water oxygen minima is actually a bottom low-oxygen water intrusion

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Processes controlling mid-water column oxygen minima over the Texas-Louisiana shelf

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Abstract We investigate distributions of dissolved oxygen over the Texas-Louisiana shelf using spatially highly resolved observations in combination with a regional circulation model with simple oxygen dynamics. The observations were collected using a towed, undulating CTD during the Mechanisms Controlling Hypoxia (MCH) program. Mid-water oxygen minimum layers (dissolved oxygen lower than 3.2 mL L^{-1}) were detected in many transects. These oxygen minimum layers are connected with the bottom boundary layer and follow the pycnocline seaward as a tongue of low oxygen into the mid-water column. T-S diagrams highlighting the low oxygen minima in both observations and simulations imply direct connections between low-oxygen bottom water and the oxygen minimum layer. The dynamics of these oxygen minimum layers in the mid-water column are examined using a three-dimensional hydrodynamic model, based on the Regional Ocean Modeling System (ROMS). Convergence within the bottom boundary layer relative to density surfaces is calculated, results show that there is a convergence in the bottom boundary layer at the location where the pycnocline intersects the bottom. Buoyancy advection forced by bottom Ekman transport creates this convergent flow, and the corresponding low-oxygen intrusion. Similar intrusions of near-bottom water into the pycnocline are observed in other regions. The presence of hypoxia within the bottom boundary layer in the northern Gulf of Mexico creates a unique situation in which these intrusions are also associated with low dissolved oxygen.

1. Introduction

The Texas-Louisiana continental shelf is located in the northern Gulf of Mexico, and the Mississippi and Atchafalaya River system is the major source of freshwater, sediment, nutrients, and pollutants for this region. The Mississippi and Atchafalaya Rivers drain around 42% of the continental watershed of the United States. Every year, the Mississippi and Atchafalaya River system delivers about $5.3 \times 10^2 \text{ km}^3$ of freshwater, $2.1 \times 10^8 \text{ t}$ (metric ton) of sediment, and $0.95 \times 10^6 \text{ t}$ of nitrogen to this region [Milliman and Meade, 1983; Meade, 1996; Bianchi *et al.*, 2010]. The outflow usually peaks in spring and is at a minimum in early fall. The large flux of nutrients leads to eutrophication within the Mississippi/Atchafalaya plume [Lohrenz *et al.*, 1990; Turner and Rabalais, 1994; Fennel *et al.*, 2011], and hypoxic conditions form in the near-bottom waters beneath the plume over the mid-shelf [Rabalais *et al.*, 2002a; Bianchi *et al.*, 2010], at depths ranging from roughly 10 to 50 m. Hypoxia occurs as early as late February and can last to early October, but is most severe, continuous, and widely distributed during June–August [e.g., Rabalais and Turner, 2001].

The wind over the shelf changes seasonally with weak, upwelling-favorable conditions in summer (June–August) and stronger, downwelling-favorable conditions during non-summer (September–May) [Cochrane and Kelly, 1986; Cho *et al.*, 1998; Nowlin *et al.*, 1998, 2005]. The seasonal wind patterns, in addition to a strong contribution from buoyancy forcing near the Mississippi and Atchafalaya Rivers, drive the low-frequency circulation on the shelf [Cochrane and Kelly, 1986; Cho *et al.*, 1998; Zhang and Hetland, 2012]. A significant portion of the freshwater discharge is pooled on the Louisiana shelf and transported downcoast (westward) by a coastal current [Cochrane and Kelly, 1986; Nowlin *et al.*, 2005; Zhang and Hetland, 2012]. During summer, when winds tend to be upwelling-favorable, stratification intensifies and inhibits the exchange of oxygen-rich surface water to subpycnocline waters [Wiseman *et al.*, 1997; Hetland and DiMarco, 2008; Bianchi *et al.*, 2010]. The stratification due to the wind-modified freshwater distribution and the eutrophication due to nutrient loading from the Mississippi River system combine to facilitate bottom hypoxia

formation over the shelf [Wiseman *et al.*, 1997; Hetland and DiMarco, 2008; DiMarco *et al.*, 2010; Zhang *et al.*, 2012; Fennel *et al.*, 2013].

The bottom boundary layer is a thin, well-mixed layer known to exchange water with the interior water column in places where the pycnocline intersects the seafloor [e.g., Pinkel, 1997; Houghton and Visbeck, 1998; Pickart, 2000; Siedlecki *et al.*, 2011]. Although the bottom boundary layer is typically much thinner than the total water column depth [e.g., Perlin *et al.*, 2005], dynamical processes acting in this layer are essential to many biogeochemical processes, such as hypoxia [Wiseman *et al.*, 1997; Fennel *et al.*, 2013]. Hypoxia is defined as a low dissolved oxygen concentration less than 1.4 mL L^{-1} (equivalent to $63 \text{ } \mu\text{M L}^{-1}$ or 2 mg L^{-1}), and it is the threshold at which most oceanic animals evacuate [Rabalais *et al.*, 2002b]. Bottom hypoxia is a common seasonal phenomenon over the Texas-Louisiana shelf. The areal extent of bottom hypoxia has been measured annually during the past three decades, with sizes ranging from 7200 to over 20,000 km^2 [Rabalais *et al.*, 2007]. Notably, seasonal hypoxia in the northern Gulf of Mexico is typically associated with the bottom boundary layer [e.g., Wiseman *et al.*, 1997; Hetland and DiMarco, 2008; Fennel *et al.*, 2013].

On the continental shelf, dynamical processes within the bottom boundary layer, such as buoyancy advection, can span the entire water column and be essential to momentum balance [e.g., Yankovsky and Chapman, 1997; Trowbridge and Lentz, 1998]. Previous studies documented that the vertical shear, generated by buoyancy advection via thermal wind adjustment, reduces the bottom current velocity and thus the bottom stress, changing the Ekman transport within the bottom boundary layer. This, in turn, alters the strength and structure of the interior flow [Trowbridge and Lentz, 1991; MacCready and Rhines, 1993; Garrett *et al.*, 1993; Middleton and Ramsden, 1996]. Chapman and Lentz [1997] extended these theories to a three-dimensional circulation; the feedback between the interior flow and the bottom boundary layer plays a crucial role in their model. Time series measurements collected on the northern California shelf show how the boundary layer flow may have an asymmetric response to the direction of the interior flow, demonstrating that the cross-shore momentum balance includes a dynamically significant buoyancy force resulting from the distortion of isopycnal surfaces within the bottom boundary layer [Trowbridge and Lentz, 1998].

A convergent zone, driven by Ekman transport within the bottom boundary layer, can be generated at the location where the pycnocline intersects the bottom [e.g., Houghton and Visbeck, 1998]. A numerical study by Chapman and Lentz [1994] released buoyant particles into the freshwater inflow, 1.36 m above the bottom at different alongshore locations after the offshore migration of the front was arrested. Some of these remained in the bottom boundary layer until they approached the shoreward edge of the front and were carried vertically upward into the alongshore jet, indicating the location of the convergence. This convergent flow at the foot of the shelfbreak front and the corresponding upwelling circulation within the front in the Middle Atlantic Bight were subsequently studied using a dye tracer injected into the bottom boundary layer [Houghton, 1997; Houghton and Visbeck, 1998]. The dye was later found to have been injected up into the water column within the front. Barth *et al.* [1998] also detected a mid-water region of suspended bottom material emanating from the foot of the front and extending to within 35 m of the surface, 80 m above bottom, providing further evidence of strong convergence within the bottom boundary layer. Pickart [2000] used a synoptic hydrographic section to investigate the characteristic of water injected into the front from the bottom boundary layer, associated with the shelfbreak jet in the Middle Atlantic Bight. Pickart [2000] found the location where the bottom boundary layer detached from the bottom, at the frontal region, and bottom water spread into the interior along an isopycnal layer abutting the front at a pumping speed of 23 m d^{-1} . Also, Pickart [2000] indicates that the detached bottom boundary layer tracks the line of maximum interior convergence, which suggests that intrusions driven by convergence might contribute to the flow along this isopycnal layer.

The primary goal of this paper is to identify the dynamical mechanisms that control the mid-water oxygen minima over the Texas-Louisiana shelf. In the model, oxygen is assumed to be saturated at the sea surface, and benthic oxygen demand is the only imposed sink of oxygen. The model reproduces the mid-water oxygen minima qualitatively. Our results demonstrate that the mid-water oxygen minima are actually low-oxygen bottom water intrusions, that the bottom boundary layer dynamics is crucial to the generation of these mid-water oxygen minima, and that the buoyancy advection in the bottom boundary layer is of fundamental importance in injecting low-oxygen bottom water up into the water column along the main pycnocline that intersects the bottom.

2. Observations

Eight shelf-wide survey cruises (Table 1) over the Texas-Louisiana Shelf were conducted in June and August from 2010 through 2013 to estimate the horizontal areal extent of hypoxia and vertical distribution of DO (Dissolved Oxygen) and other hydrographic properties. Towed, undulating CTD profiles were collected during these cruises (see Table 1 for details) to measure water properties along cross-shelf transects. The towed instrument was equipped with a Sea-Bird SBE 43 DO sensor, RINKO DO sensor, Sea-Bird 55 CTD, WET-labs Fluorometer/Turbidity, and CDOM (Chromophoric Dissolved Organic Matter) sensors. The system sampled water properties every 0.25 second, and was towed between 2 m above the bottom and 2 m below the surface. During each transect, the ship maintained a constant heading and a tow velocity of 9.26 km/h, along cross-shelf transects designed to be roughly perpendicular to the local bathymetric lines. Each transect took 4–6 h to complete. CTD casts were made at the inshore and offshore ends of each cross-shelf transect. Water samples from surface, mid-depth, and bottom were collected and analyzed for nutrients, DO concentrations. The Winkler titration method was used to measure DO concentration of the water samples and the results were used to calibrate the SBE 43 electronic sensors. The properties measured through the towed instrument were compared with the measurement in the lab from discrete water samples to calibrate the electronic probe DO measurements. Cross-shelf transects were not completed in cruises MS1 and MS3 due to equipment failures and weather (S. F. DiMarco, unpublished data, 2010–2014).

3. Hydrodynamic Model Description

The Regional Ocean Modeling System (ROMS) is configured for the Texas-Louisiana shelf, as described by Zhang *et al.* [2012]. The model has 30 vertical layers, and the coordinates are spaced closer near the surface and bottom to better resolve the two boundary layers. The model has a minimum mean water depth of 3 m, and a maximum slightly greater than 3000 m, with a resolution of about ~500 m in the northern Gulf in the region of the Mississippi and Atchafalaya river sources, and ~1–2 km further west offshore of the Texas coast. At the three open boundaries, a nudging zone of six cells was used to relax the model temperature, salinity, and baroclinic velocities toward the Gulf of Mexico Hybrid Coordinate Ocean Model (GOM-HYCOM) (<http://www.hycom.org>). The nudging time scale used was 8 h at the outermost boundary point, with successively weaker nudging in interior points. Sea surface height and barotropic currents from HYCOM were imposed at the boundaries as Chapman [1985] and Flather [1976] boundary conditions.

The hindcast model is forced with surface momentum, heat, and freshwater fluxes from the North American Regional Reanalysis (NARR) data set. Freshwater fluxes from the Mississippi and Atchafalaya Rivers are specified using daily measurements of Mississippi River Transport at Tarbert Landing by the U.S. Army Corps of Engineers. Freshwater fluxes from the other seven rivers (the Nueces, San Antonio, Lavaca, Brazos, Trinity, Sabine, and Calcasieu Rivers) are specified based on the USGS (U.S. Geological Survey) Real-Time Water Data for the Nation. For further details on model configuration, we refer reader to Zhang *et al.* [2012].

The model has a simple parameterization of benthic oxygen demand and air-sea gas exchange of oxygen in order to describe oxygen dynamics [Hetland and DiMarco, 2008]. Oxygen is initialized at saturated values everywhere based on temperature and salinity [Weiss, 1970]. Water from the rivers is also saturated with respect to oxygen. Oxygen concentrations at the open boundaries are relaxed to saturation in exactly the same way as the dynamical tracers. The possible effects of photosynthesis increasing water column oxygen above saturated values are not considered [Hetland and DiMarco, 2008].

The model specifies benthic respiration, applied as a negative flux of oxygen at the seafloor, based on bottom temperature and oxygen concentrations. The function is

$$\text{Bottom } O_2 \text{ flux} = 6.0 [O_2 \text{ m}^{-2} \text{d}^{-1}] \times 2^{T/10.0^\circ \text{C}} \times \left[1 - \exp\left(\frac{-O_2}{30.0 \mu\text{M } O_2}\right) \right] \quad (1)$$

This formulation was proposed by Hetland and DiMarco [2008], and is based on data obtained by Rowe *et al.* [2002]. We refer the reader to Hetland and DiMarco [2008] for further details.

Table 1. Survey Cruise Identifiers and Their Corresponding Dates^a

Cruise ID	Start Date	End Date	Transects
MS1	14 Jun 2010	19 Jun 2010	5
MS2	2 Aug 2010	7 Aug 2010	15
MS3	23 Jun 2011	1 Jul 2011	6
MS4	7 Aug 2011	15 Aug 2011	15
MS5	11 Jun 2012	17 Jun 2012	15
MS6	16 Aug 2012	20 Aug 2012	13
MS7	20 Jun 2013	26 Jun 2013	16
MS8	4 Aug 2013	9 Aug 2013	15

^aTotal numbers of transects are also included.

4. Results

Figure 1 shows the study domain with simulated 31 g kg^{-1} isohaline for both surface (blue) and bottom (red) during June 2011. These thin lines denote the approximate extent of freshwater from river plumes. The shelf is broad and gently sloping to the west of Terrebonne Bay, and the surface and bottom isohalines are aligned, indicating that the plume is relatively more mixed. The shelf is steeper in the eastern part, between the Mississippi River Delta and Terrebonne Bay, and here the surface salinity front is further offshore than that at the bottom, indicating that the plume is more stratified.

Two observed transects of dissolved oxygen concentration are shown, with isopycnals overlaid (thin black lines), in Figure 2 (left); transect A is south of Atchafalaya Bay and transect B is west of the Mississippi River Delta. Hypoxic water (orange, $\text{DO} < 1.4 \text{ mL L}^{-1}$) and near-hypoxic, low-oxygen water (yellow, $1.4 \text{ mL L}^{-1} < \text{DO} < 3.2 \text{ mL L}^{-1}$) are both observed near the bottom. In addition to the bottom-trapped hypoxia, layers of oxygen minima in the mid-water column are also detected in the transects. Figure 2 (right) shows the corresponding T-S diagrams. The low-oxygen layer is colocated with the main pycnocline for both transects (Figure 2, left). Another noticeable feature of Figure 2 transect A1 is that the near-bottom, low-oxygen water extends up through the water column at 28.97°N , reaches to within 10–12 m of the surface, and merges into the main pycnocline low-oxygen layer. The same upward extension of low dissolved oxygen is found in Figure 2 transect B1 at 29.06°N . Only transect A will be discussed in detail since the two transects have very similar characteristics.

The isopycnals reveal the location of the bottom-attached front; the mid-water oxygen minimum is bounded within the pycnocline and attached to the seafloor at the foot of the front (Figure 2, left). The thickness of the bottom boundary layer here is operationally defined as the distance from the bottom to the maximum height at which the vertical density gradient is greater than 0.1 kg/m^4 . This is an operational definition that does not significantly change the results discussed, see Wiseman *et al.* [1997]. The bottom boundary layer becomes thinner where the frontal regions intersect the seafloor (the blue lines in Figure 2, left). Also, Figure 2 reveals that the bottom near-hypoxic layer becomes thinner shoreward, these are consistent with the conclusion of previous studies [Wiseman *et al.*, 1997; Fennel *et al.*, 2013; Hetland and DiMarco, 2008]. Wiseman *et al.* [1997], Fennel *et al.* [2013], and Hetland and DiMarco [2008] demonstrate that the bottom hypoxic layer thickness is highly correlated with the bottom boundary layer thickness and the top of the bottom boundary layer is associated with the upper reach of hypoxic bottom water.

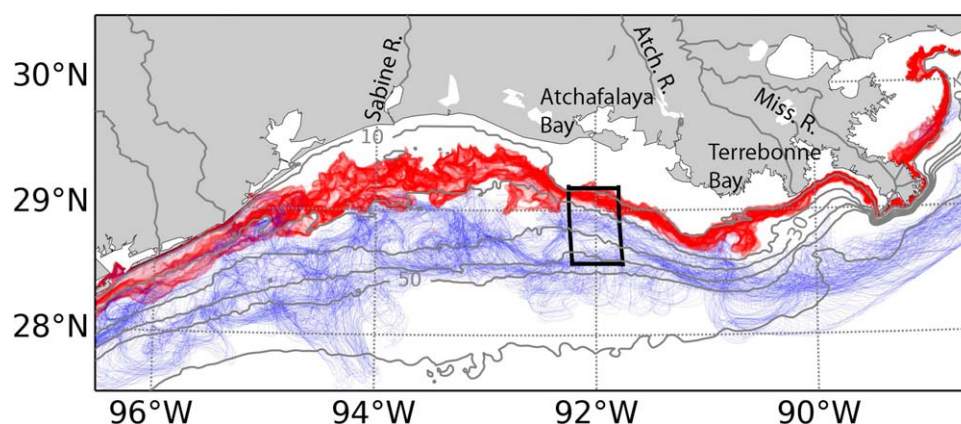


Figure 1. The study domain covers much of the Texas-Louisiana continental shelf. Bathymetric contours of 10, 20, 30, 40, 50, and 200 m depth are shown on the map. The simulated salinity fronts (31 g kg^{-1} isohaline) are shown for both surface (blue lines) and bottom (red lines) of June 2011. The black polygon denotes a region that is analyzed to investigate convergent zone in Figure 7. This same idea as Hetland *et al.* [2012] was used to create this figure.

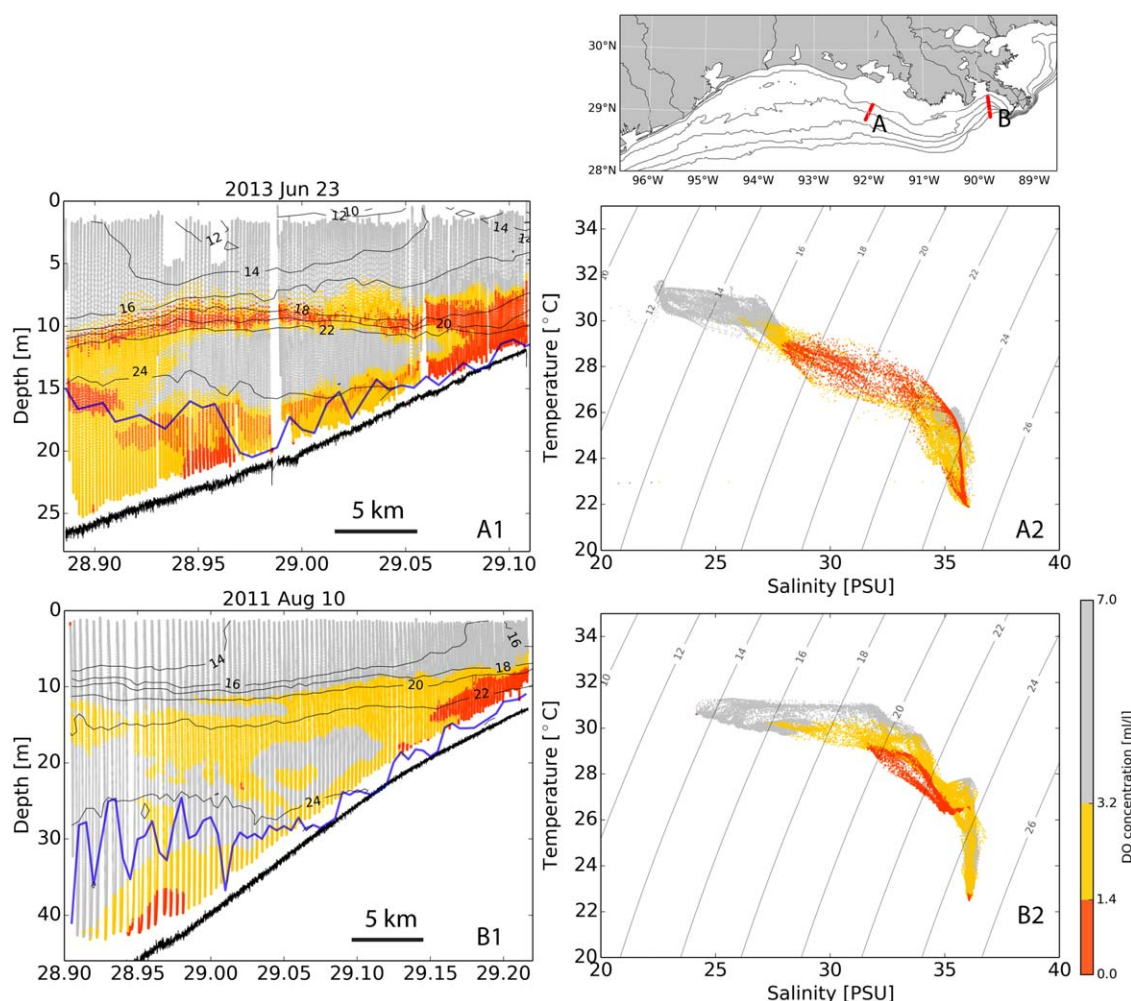


Figure 2. Observed dissolved oxygen concentration on transects A and B. Orange indicates hypoxia ($\text{DO} < 1.4 \text{ mL L}^{-1}$), yellow is defined as near-hypoxic, low oxygen ($1.4 \text{ mL L}^{-1} < \text{DO} < 3.2 \text{ mL L}^{-1}$), and gray is any oxygen concentration higher than 3.2 mL L^{-1} . The critical value, 3.2 mL L^{-1} , is an operational definition that does not significantly change the results discussed. (left) Black lines overlaid with are the isopycnals and the thick blue line is the top of the bottom boundary layer. (right) The corresponding T-S diagram, and the shading of the points are relative to dissolved oxygen concentration. The density contours use an interval of 2 kg m^{-3} for all plots.

We examined transects collected on six cruises, and detected mid-water oxygen minimum layers in many of them. Sometimes the mid-water oxygen minimum layers are similar in extent to those shown in Figure 2, sometimes they are less pronounced and similar to the ones in Figure 3 transect C. Sometimes, there are more than one mid-water oxygen minimum layer in a single transect (Figure 3, transect D).

There are two possible reasons that can cause mid-water dissolved oxygen minima. First, dissolved oxygen could be depressed locally by water column respiration, exacerbated by inhibited mixing and reduced ventilation within the pycnocline. Second, the intermediate low-oxygen region may be formed by advection from the bottom boundary layer along isopycnals connecting the bottom and the intermediate low-oxygen layer, caused by a convergence in the bottom boundary layer. However, it is unlikely that the mid-water dissolved oxygen minima could be generated locally, since we would not expect the water column respiration to occur at significantly higher rates only in the main pycnocline compared to other regions. Figure 4b shows the observed CDOM (chromophoric dissolved organic matter) of transect A. CDOM forms a significant constituent of the dissolved organic matter pool in natural waters and is derived allochthonously from terrestrial environments or autochthonously from in situ phytoplankton primary production or its microbial decomposition [D'Sa and DiMarco, 2009]. Relatively lower levels (generally lower than 8 mg m^{-3}) of CDOM within the front, compared to high CDOM levels in the water

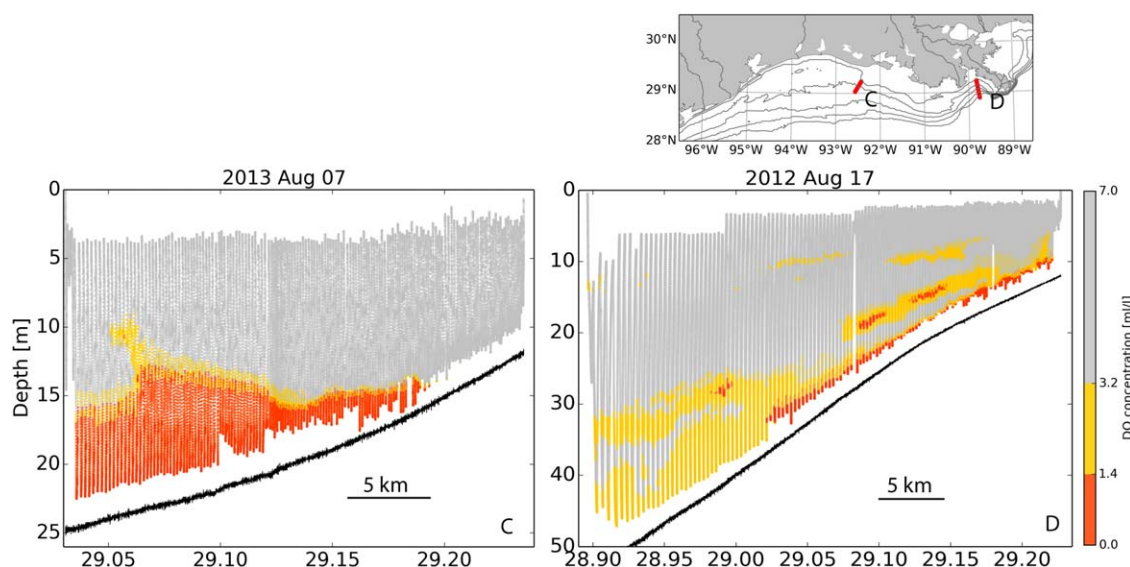


Figure 3. Observed dissolved oxygen concentrations of transects C and D, the same color map from Figure 2 is used.

overlying, indicate that there is not likely an elevated source of respiration due to a locally enhanced supply of organic material.

Inspection of the observed T-S diagrams in Figure 2 implies a direct connection between low-oxygen water within the bottom boundary layer and that within the mid-water layer. The high-density, low dissolved oxygen water masses from the bottom of Figure 2 transect A2 corresponds to the low-oxygen bottom water underneath the 24 kg m^{-3} isopycnal in A1. Moving to fresher isopycnals along the bottom as well as into the frontal layer (Figure 2, transect A1), the corresponding low-oxygen points in the T-S diagram are all connected, without any discernible gap, while crossing multiple isopycnals. This suggests that the mid-water low-oxygen layer is directly connected with the low-oxygen bottom water.

We also analyzed the hydrodynamic model results to identify the source of the mid-water column oxygen minima. A transect of simulated oxygen concentration in the vicinity of the observed transect A is shown in Figure 5. The simulated dissolved oxygen concentration is generally higher than the actual observed dissolved oxygen concentration, most likely because the model only includes benthic respiration and does not account for water column respiration. The model also shows mid-water oxygen minimum (Figure 5) along isopycnals 16 through 18 kg m^{-3} within the stratified front and extending offshore along frontal isopycnals and shoaling toward the surface.

Because the model only uses benthic respiration, the tongue of low dissolved oxygen concentration must ultimately come from the bottom (Figure 5). So the model results suggest that the mid-water low-oxygen layer is advected from the low-oxygen bottom water; and the low-oxygen water is associated with a circulation from the bottom boundary layer into the frontal region. The corresponding simulated T-S diagram (right panel of Figure 5) also shows that the connection between the low oxygen intrusion and bottom water is continuous. Both the observed mid-water low-oxygen layer structure and the T-S diagram distribution are similar in the simulation, suggesting that the same mechanism is behind observed and simulated mid-water oxygen minima.

Similar processes are described by Houghton and Visbeck [1998], Barth *et al.* [1998], and Pickart [2000]; a convergent flow in the bottom boundary layer is found to create a flux of water into the pycnocline. Stronger alongshore flow in shallower regions creates an offshore flux of water in the bottom boundary layer. However, beneath the pycnocline, the flow is halted or reversed, so that there is an onshore flow in the bottom boundary layer. This creates a convergence in the bottom boundary layer at the point where the pycnocline intersects the seafloor, and water is injected from the bottom boundary layer into the pycnocline. We believe this is the process that is responsible for creating the mid-water column dissolved oxygen minimum.

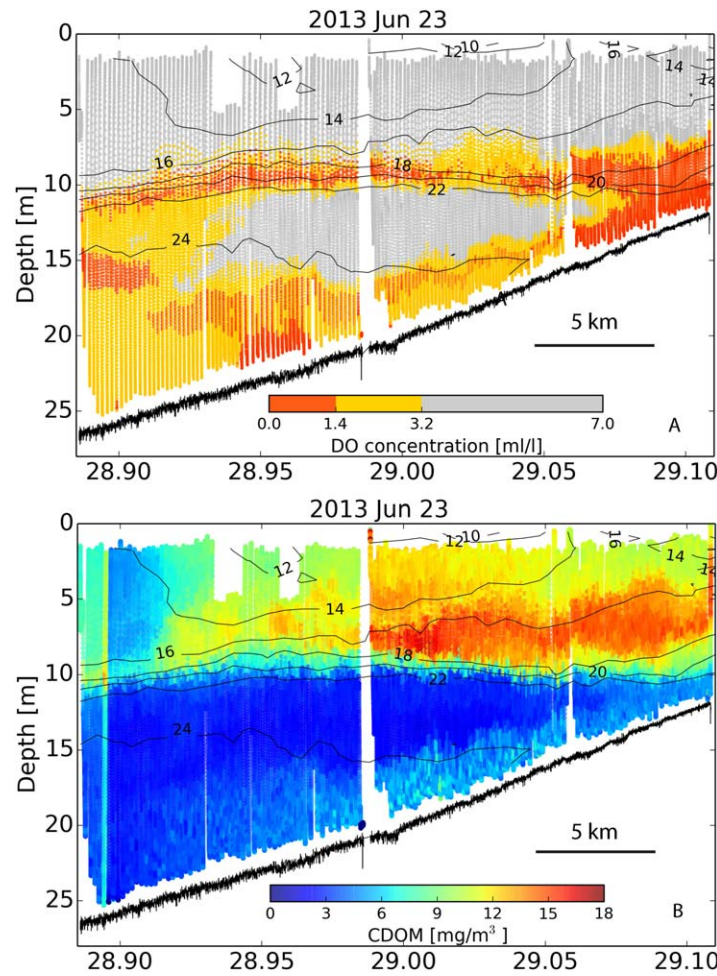


Figure 4. Observed (a) DO and (b) CDOM for transect A, black lines overlaid with are the isopycnals.

To find the convergent zones within the bottom boundary layer over the Texas-Louisiana shelf, first, flow through density surfaces for the bottom cell of the model is calculated using simulated results. In the ocean, the density equation is

$$\frac{\partial \rho}{\partial t} + \vec{u} \cdot \vec{\nabla} \rho = \vec{\nabla} \cdot \vec{F} \quad (2)$$

In equation (2), ρ is the density field, $\vec{u}$ is the simulated current velocity vector, and $\vec{F}$ is the diffusivity term. The total current velocity vector along the bottom, $\vec{u}$, can be divided into three parts, the velocity of the density surface itself $\vec{u}_I$, the flow through the density surface $\vec{u}_M$, and the flow along the density surface $\vec{u}_P$ (Figure 6). Since the dot product of $\vec{u}_P$ and density gradient, $\vec{\nabla} \rho$, is zero, $\vec{u}_P$ can be ignored in the equation. In other words, $\vec{u}_I + \vec{u}_M$ is the normal component (in the sense of density surface) of bottom current velocity, $\vec{u}_c$. So, equation (2) can be rewritten as

$$\frac{\partial \rho}{\partial t} + (\vec{u}_I + \vec{u}_M) \cdot \vec{\nabla} \rho = \vec{\nabla} \cdot \vec{F} \quad (3)$$

In equation (3), $\vec{u}_I$ is associated with inviscid motions and is reversible, which means the density surface itself can move on and offshore as well as go back to its original location; $\vec{u}_M$ is associated with mixing and is irreversible. Such that, equation (3) can be divided into two components

$$\frac{\partial \rho}{\partial t} + \vec{u}_I \cdot \vec{\nabla} \rho = 0 \quad (4)$$

and

$$\vec{u}_M \cdot \vec{\nabla} \rho = \vec{\nabla} \cdot \vec{F} \quad (5)$$

Figure 6 shows the bottom layer in x-y plane; the gray line represents a density surface along the bottom. The isopycnal motions are estimated using equation (4) describing the motions of a density surface at the seafloor. Diffusion is ignored, because this equation follows the isopycnal surfaces instead of water parcels. The magnitude of isopycnal motion will be

$$u_I = \frac{|\frac{\partial \rho}{\partial t}|}{|\vec{\nabla} \rho|} \quad (6)$$

The motion is perpendicular to the bottom density surface, and the direction of isopycnal motion depends on the sign of the $\frac{\partial \rho}{\partial t}$ term. Water motion along the bottom is calculated directly from the simulated current

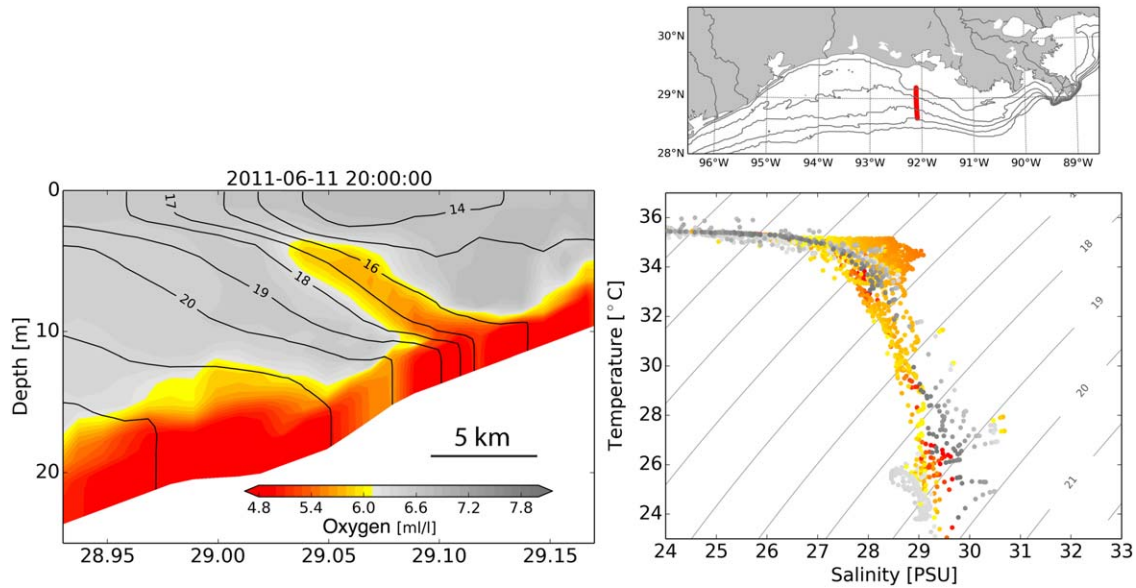


Figure 5. (left) The simulated dye tracer concentration of a transect in the vicinity of the observed transect A (around Atchafalaya river plume, see the map on top right), the black lines overlaid with are the isopycnals. (right) The corresponding T-S diagram, and the shading of the points are relative to dye tracer concentration. The density contours use an interval of 1 kg/m^3 for both plots.

velocity in the bottom layer. Only the cross-isopycnal flow will advect or flow through the surface isopycnal, and the cross isopycnal component of water motion is obtained using

$$\frac{\vec{u}_c = \vec{u} \cdot \vec{\nabla} \rho}{|\vec{\nabla} \rho|} \quad (7)$$

In this equation, $\vec{u}_c$ represents the cross isopycnal component of water motion, and $\frac{\vec{\nabla} \rho}{|\vec{\nabla} \rho|}$ is the unit vector perpendicular to the isopycnal surface. The flow through density surface is obtained by subtracting the isopycnal motion along the bottom from cross isopycnal component of water motion

$$\frac{\vec{u}_M = \vec{u} \cdot \vec{\nabla} \rho}{|\vec{\nabla} \rho| - \vec{u}_l} \quad (8)$$

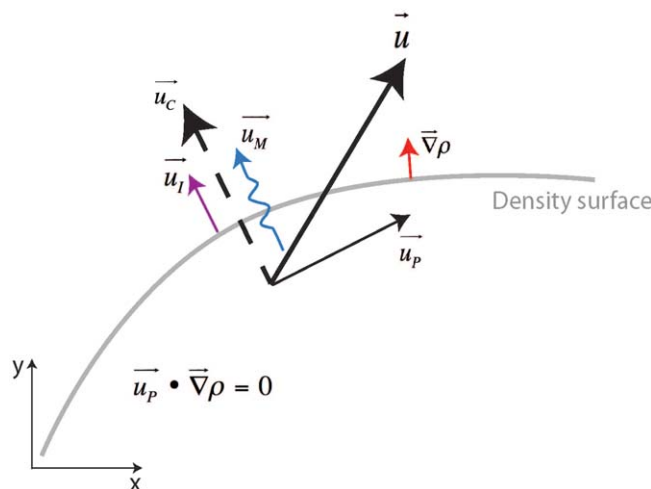


Figure 6. Sketch depicting flow along the bottom cell of the model. The gray line is a density surface, the thin black arrow represents current velocity along density surface, the dashed black arrow represents the cross isopycnal component of water motion, the blue arrow represents flow through density surface, and the purple arrow represents density surface motion along the bottom (equations (3) and (7)).

Instead of the strength of mixing, our interest in this study is the convergence of the flow, relative to the density surfaces, which will then be

$$\xi = \vec{\nabla} \cdot (\vec{u}_c - \vec{u}_l) \quad (9)$$

Following equation (9), the convergence along the bottom cell of the whole model domain can be computed.

Previous studies used different ways to detect convergent flow and the subsequent intrusion at the foot of the shelf-break front in the Middle Atlantic Bight [Houghton, 1997; Houghton and Visbeck, 1998;

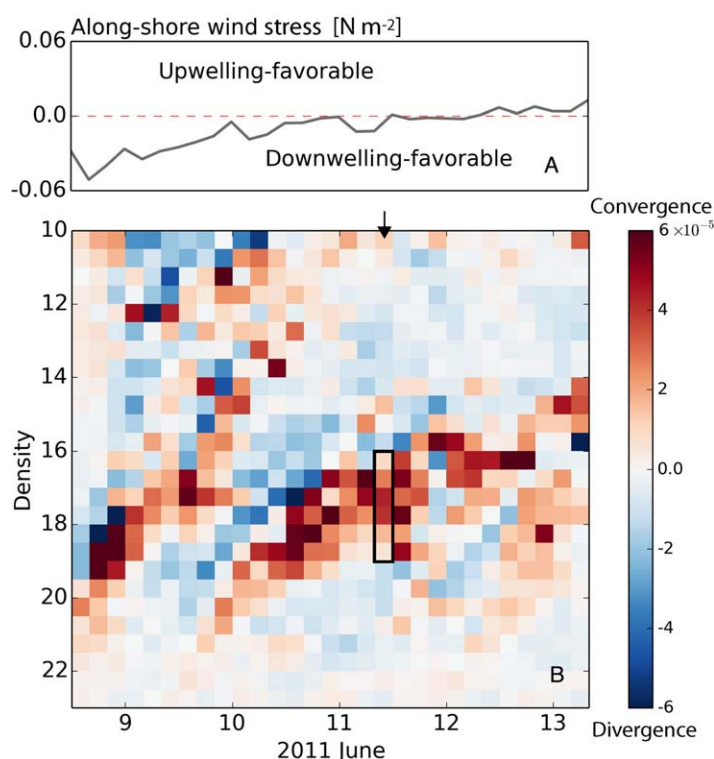


Figure 7. (a) The alongshore wind stress during the analyzed time period; (b) Time series of the simulated convergence relative to isopycnals (result of equation (9)), cast in density space (with a density interval of 0.5 kg m^{-3}). The black arrow denotes the time step in Figure 5 (11 June 2011 20:00), and the black rectangle marks out the convergence range for that time step.

Barth *et al.*, 1998; Pickart, 2000]. The measured intrusion into the front had an averaged velocity of $9 \pm 2 \text{ m d}^{-1}$ [Barth *et al.*, 1998]. However, the Middle Atlantic Bight shelfbreak front is a relatively stationary feature. In the northern Gulf, the temporal change of fresh-water discharge alters the stratification structure and the location where the main pycnocline intersects the seafloor. Wind stress over the shelf changes on seasonal time scales, but there is also considerable variability in the weather band, with time scales of a few days to weeks [Hetland and DiMarco, 2008]. Wind forcing can also modify the distribution of buoyancy on the shelf, and shift the water column stratification and the pycnocline. Furthermore, in summer, the land-sea breeze excites strong near-inertial flows with 24 h periodicity over the entire Texas-Louisiana shelf [Zhang *et al.*, 2009]. Combined,

these factors lead to a very transient stratification structure, thus a transient front over the Texas-Louisiana shelf [Gawarkiewicz and Chapman, 1992]; isopycnals are pushed on and offshore, and bottom boundary layer convergence can occur along different isopycnals at various times and locations. Simulated results indicate that a corresponding intrusion of bottom water usually is generated a day after the formation of a convergence.

Since the distribution of convergence is patchy as described above, convergence properties in Cartesian coordinates may be quite different even for transects in the same vicinity. This will make the interpretation of the convergence and a verification of the consistence of convergence and corresponding intrusion difficult. However, as described above, the intrusion of low-oxygen bottom water is along certain isopycnals. A coordinate system that follows density coordinates and is less sensitive to the geographical location of the convergence is efficient and necessary to address these issues. By identifying regions of persistently convergent flow in a particular range of isopycnals, regions where water is injected into the interior water column can be identified. The way we build a density coordinate is to bin the density field of an analyzed region with a density interval of 0.5 kg m^{-3} , and then the corresponding convergence for each bottom grid cell is cast into the binned density space. Multiple grid cells can be cast into the same density space, and the volume of the grid cells takes a variety of values since the horizontal and vertical resolution of the model changes spatially. Thus, the convergences are weighted by the volume of each model grid cell and then the sum of the weighted convergences represents the convergence in the density space. This metric considers the water mass structure of the convergence as a whole and identifies the isopycnals where convergence occurs.

The method described above was employed on the simulated fields to estimate the convergence relative to isopycnals over the Texas-Louisiana shelf during June 2011. A region south of Atchafalaya Bay is analyzed (black polygon in the map of Figure 1); and then the calculated convergence is cast into a density

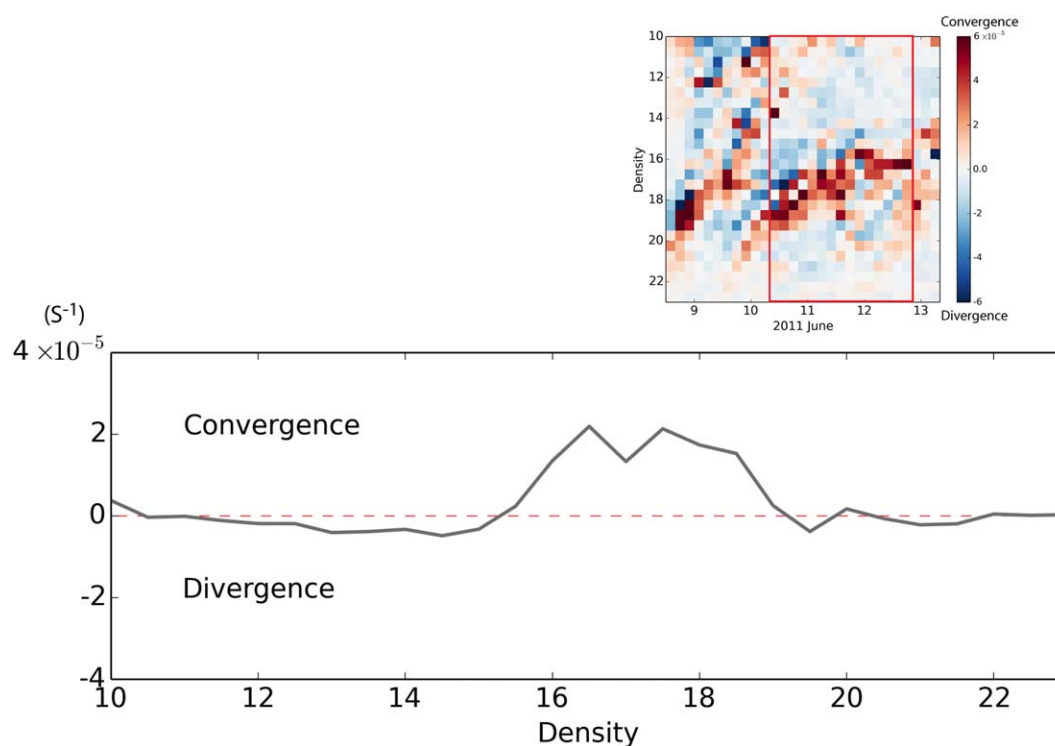


Figure 8. A 2.5 days average of the convergence in density coordinate, the averaged time period is denoted by the red rectangle.

coordinate. Note that, the transect shown in Figure 5 is in this black region. Figure 7a shows the time series of alongshore wind stress (averaged over the analyzed area), downwelling favorable, for this time period. Atmospheric length scales are large enough that the wind stress over the shelf is quite homogenous, thus averaging over the analyzed area is appropriate given the spatial and temporal scales of the local wind field [Wang *et al.*, 1998]. The downwelling-favorable alongshore wind excited downslope Ekman transport, which could cause buoyancy transport along the bottom and support the formation of a convergence. Figure 7b is a time series of the convergent flow in density space; the time shown in Figure 5 is marked with a black arrow. Colors represent convergent or divergent flows; convergent flow implies a net gain of water along a specific isopycnal and divergent flow implies a net loss of water.

A divergent flow is generated on 9 June along isopycnal 23 kg m^{-3} , and this divergence lasts for 4 days while it moves to fresher isopycnals. This divergent flow is bounded by two convergent flows. A time-averaged convergence field in density space clearly shows that the convergent flow is along frontal isopycnals and the divergent flows are along isopycnals underneath the front as well as above the front (Figures 5 and 8). At the time of the transect shown in Figure 5, as well as for the time average shown in Figure 8, convergence occurs along isopycnals 16.5 through 19 kg m^{-3} . This convergence range is slightly denser than what Figure 5 suggests, because of averaging over a large region instead of a single transect.

5. Discussion

Mid-water oxygen minimum layers have often been observed in summer over the Texas-Louisiana shelf. Analysis of the water properties along these minima indicates that these low-oxygen layers are connected with the low dissolved oxygen bottom water through mixing of water masses. These mid-water oxygen minima are also found in simulated results from a hydrodynamic model based on ROMS, which uses only benthic oxygen respiration [Hetland and DiMarco, 2008; Zhang *et al.*, 2012], with similar characteristics to observed oxygen minima. In the model the low-oxygen layer in the mid-water column must ultimately come from the bottom, as that is its only sink of oxygen. The model is also used to address the dynamical mechanisms of these mid-water low-oxygen layers. Convergent flow at the bottom, relative to isopycnal

surfaces, is strongest in the density classes associated with the oxygen minimum layer. We believe these mid-water oxygen minima are actually intrusions of low oxygen protruding from the bottom boundary layer via buoyancy advection-driven convergence, following the main pycnocline.

The temporal evolution of the simulated convergent regions over the Texas-Louisiana shelf indicates that the formation and destruction of convergence is fast, typically lasting only a few days, most likely related to the transient nature of the nearshore frontal region that occurs in water less than 50 m [Hetland and DiMarco, 2008; Bianchi et al., 2010; DiMarco et al., 2010]. Further offshore and further from the freshwater sources, convergence becomes weaker or even vanishes, due to the weakening of stratification associated with the Mississippi-Atchafalaya River plume system. The time series of convergence in Figure 7 shows two periods of convergence; the first occurs during 9–11 June (3 days) and the other during 10–13 June (4 days).

The evolution of both convergences shown in Figure 7 indicates that the convergent flow first forms along denser isopycnals; the convergent zone moves to fresher isopycnals over time. The mid-water layer structure of properties is also described and supported by Houghton [1997], Houghton and Visbeck [1998], Barth et al. [1998], Pickart [2000], Linder et al. [2004], and Hales et al. [2009]. In contrast to convergent flow at the foot of the stationary front in the Middle Atlantic Bight [Houghton, 1997; Houghton and Visbeck, 1998; Barth et al., 1998; Pickart, 2000], this transient feature is unique to the Texas-Louisiana shelf as the region affected by strong lateral density gradients spans most of the inner shelf, inshore of 50 m. A convergence in the bottom boundary layer transport can form at the location where the arrest or reversal in the Ekman transport occurs, which is most likely the location where the front intersects the seafloor. The migrating convergent zone corresponds to shifting frontal features over the Texas-Louisiana shelf. A 1 day long stable convergence occurs along isopycnals 16.5 through 19 kg m⁻³, facilitating the large low oxygen intrusion (Figure 5). After that the convergence becomes weaker and the intrusion is halted.

6. Conclusions

We have investigated the dynamical mechanisms of the mid-water column dissolved oxygen minima that have been observed using high-resolution towed, undulating CTD profiles. The convergence relative to density surfaces are calculated based on hydrodynamic model results, revealing that the mid-water low-oxygen layer is consistent with the bottom convergent flow and the corresponding intrusion of bottom water (Figures 5 and 7). These demonstrate the existence of the near-bottom convergence and subsequent coastal intrusions over the Texas-Louisiana shelf.

The dynamical mechanism of the mid-layer, low-oxygen tongue is a convergent flow and the corresponding intrusion of bottom water produced by buoyancy flux via bottom Ekman transport. Hetland and DiMarco [2008] indicate that water column respiration could be important for the formation of hypoxia over the Texas-Louisiana shelf, especially for the region between the Mississippi River Delta and Terrebonne Bay. It is possible that this type of respiration could also contribute to the formation of the detected mid-water low-oxygen layer, especially for the large main pycnocline bounded low-oxygen layers. However, it seems unlikely that water column respiration would occur at significantly higher rates only in the main pycnocline. Also, low oxygen intrusions were observed in the hydrodynamic model, which ignores water column respiration. The buoyancy-driven convergence in the bottom boundary layer and the corresponding intrusion are essential in forming the mid-water column low-oxygen layer, and the mid-water oxygen minimum layer is actually the intrusion of bottom low-oxygen water.

The temporal evolution of the simulated low oxygen intrusion in Figure 5 reveals that the low oxygen shoals 8 m along the frontal isopycnals from the bottom in 20 h, equivalent to an intrusion velocity of about 10 m d⁻¹, which is similar to what Houghton and Visbeck [1998] and Barth et al. [1998] suggested. The buoyancy advection in the bottom boundary layer is essential to the bottom boundary layer structure itself [Gawarkiewicz and Chapman, 1992]; meanwhile, the frontogenesis driven by this can alter the overlying flow field. Furthermore, the buoyancy advection in the bottom boundary layer has important implications for the concentration of material on the offshore side of the front and the biogeochemistry of the shelf; convergence generates intrusions of nutrients or other materials from the bottom [Siedlecki et al., 2011].

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References

- Barth, A., D. Bogucki, S. D. Pierce, and P. M. Kosro (1998), Secondary circulation associated with a shelfbreak front, *Geophys. Res. Lett.*, **25**(15), 2761–2764.
- Bianchi, T. S., S. F. DiMarco, J. H. Cowan Jr., R. D. Hetland, P. Chapman, J. W. Day, and M. A. Allison (2010), The science of hypoxia in the Northern Gulf of Mexico: A review, *Sci. Total Environ.*, **408**(7), 1471–1484, doi:10.1016/j.scitotenv.2009.11.047.
- Chapman, D. C. (1985), Numerical treatment of cross-shelf open boundaries in a Barotropic Coastal Ocean Model, *J. Phys. Oceanogr.*, **15**(8), 1060–1075.
- Chapman, D. C., and S. J. Lentz (1994), Trapping of a coastal density front by the bottom boundary layer, *J. Phys. Oceanogr.*, **24**, 1464–1479.
- Chapman, D. C., and S. J. Lentz (1997), Adjustment of Stratified Flow over a Sloping Bottom, *J. Phys. Oceanogr.*, **27**(2), 340–356.
- Cho, K., R. O. Reid, and W. D. Nowlin (1998), Objectively mapped stream function fields on the Texas-Louisiana shelf based on 32 months of moored current meter data, *J. Geophys. Res.*, **103**(C5), 10,377–10,390.
- Cochrane, J. D., and F. J. Kelly (1986), Low-frequency circulation on the Texas-Louisiana continental shelf, *J. Geophys. Res.*, **91**(C9), 10,645–10,659.
- DiMarco, S. F., P. Chapman, N. Walker, and R. D. Hetland (2010), Does local topography control hypoxia on the eastern Texas–Louisiana shelf?, *J. Mar. Syst.*, **80**, 25–35.
- D'Sa, E. J., and S. F. DiMarco (2009), Seasonal variability and controls on chromophoric dissolved organic matter in a large river-dominated coastal margin, *Limnol. Oceanogr. Methods*, **54**, 2233–2242.
- Fennel, K., R. Hetland, Y. Feng, and S. DiMarco (2011), A coupled physical-biological model of the Northern Gulf of Mexico shelf: Model description, validation and analysis of phytoplankton variability, *Biogeosciences*, **8**, 1881–1899, doi:10.5194/bg-8-1881-2011.
- Fennel, K., J. Hu, A. Laurent, M. Marta-Almeida, and R. Hetland (2013), Sensitivity of hypoxia predictions for the Northern Gulf of Mexico to sediment oxygen consumption and model nesting, *J. Geophys. Res. Oceans*, **118**, 990–1002, doi:10.1002/jgrc.20077.
- Flather, R. A. (1976), A tidal model of the northwest European continental shelf, *Memo. Soc. R. Sci. Liege*, **10**(6), 141–164.
- Garrett, C., P. MacCready, and P. B. Rhines (1993), Boundary mixing and arrested Ekman layers: Rotating stratified flow near a sloping boundary, *Annu. Rev. Fluid Mech.*, **25**, 291–323.
- Gawarkiewicz, G., and D. C. Chapman (1992), The role of stratification in the formation and maintenance of shelf-break fronts, *J. Phys. Oceanogr.*, **22**, 753–772.
- Hales, B., D. Hebert, and J. Marra (2009), Turbulent supply of nutrients to phytoplankton at the New England shelf break front, *J. Geophys. Res.*, **114**, C05010, doi:10.1029/2008JC005011.
- Hetland, R. D., and S. F. DiMarco (2008), How does the character of oxygen demand control the structure of hypoxia on the Texas–Louisiana continental shelf?, *J. Mar. Syst.*, **70**, 49–62, doi:10.1016/j.marsys.2007.03.002.
- Hetland, R. D., K. Fennel, C. K. Harris, J. Kaihatu, K. Xu, and S. F. DiMarco (2012), Integrated bio-physical modeling of Louisiana-Texas (LATEX) Shelf, BOEM 2012-108, technical report, U. S. Dep. of the Inter., Bur. of Ocean Energy Manage. Gulf of Mexico OCS Reg., New Orleans, La.
- Houghton, R. W. (1997), Lagrangian flow at the foot of a shelfbreak front using a dye tracer injected into the bottom boundary layer, *Geophys. Res. Lett.*, **24**(16), 2035–2038.
- Houghton, R. W., and M. Visbeck (1998), Upwelling and convergence in the Middle Atlantic Bight shelfbreak front, *Geophys. Res. Lett.*, **25**(15), 2765–2768.
- Linder, C. A., G. G. Gawarkiewicz, and R. S. Pickart (2004), Seasonal characteristics of bottom boundary layer detachment at the shelfbreak front in the Middle Atlantic Bight, *J. Geophys. Res.*, **109**, C03049, doi:10.1029/2003JC002032.
- Lohrenz, S. E., M. J. Dagg, and T. E. Whitledge (1990), Enhanced primary production at the plume/oceanic interface of the Mississippi River, *Cont. Shelf Res.*, **10**, 639–664.
- MacCready, P., and P. B. Rhines (1993), Slippery Bottom Boundary Layers on a Slope, *J. Phys. Oceanogr.*, **23**, 5–22.
- Meade, R. (1996), River-sediment inputs to major deltas, in J. D. Milliman and B. U. Haq (eds), *Sea-Level Rise and Coastal Subsidence*, pp. 63–85, Kluwer Acad., London, U. K.
- Middleton, J. F., and D. Ramsden (1996), The evolution of the bottom boundary layer on the sloping continental shelf: A numerical study, *J. Geophys. Res.*, **101**(C8), 18,061–18,077.
- Milliman, J. D., and R. H. Meade (1983), World-wide delivery of river sediment to the ocean, *J. Geol.*, **91**(1), 1–21.
- Nowlin, W. D., A. E. Jochens, R. O. Reid, and S. F. DiMarco (1998), Texas-Louisiana shelf circulation and transport process study: Synthesis report LATEX A, vols. 1 and 2. OCS study MMS 98-0035 and MMS 98-0036, technical report, U. S. Dep. of the Inter., Miner. Manage. Serv., Gulf of Mexico OCS Reg. Off., New Orleans, La.
- Nowlin, W. D., N. D., A. E. Jochens, S. F. DiMarco, R. O. Reid, and M. K. Howard (2005), Low-frequency circulation over the Texas-Louisiana continental shelf, in *Circulation in the Gulf of Mexico: Observations and Models*, edited by W. Sturges and A. Lugo-Fernandez, pp. 219–240, AGU, Washington, D. C.
- Perlin, A., J. N. Moum, and J. M. Kymak (2005), Response of the bottom boundary layer over a sloping shelf to variations in alongshore wind, *J. Geophys. Res.*, **110**, C10S09, doi:10.1029/2004JC002500.
- Pickart, R. S. (2000), Bottom boundary layer structure and detachment in the shelfbreak jet of the middle Atlantic bight, *J. Phys. Oceanogr.*, **30**, 2668–2686.
- Pinkel, R. (1997), *Observations of Coherent Structures in the Oceanic Boundary Layer*, Scripps Inst. of Oceanogr. La Jolla ca Mar. Phys. Lab.
- Rabalais, N., R. Turner, B. Sen Gupta, D. Boesch, P. Chapman, and M. Murrell (2007), Hypoxia in the northern Gulf of Mexico: Does the science support the plan to reduce, mitigate, and control hypoxia?, *Estuaries Coasts*, **30**(5), 753–772.
- Rabalais, N. N., and R. E. Turner (2001), *Coastal Hypoxia: Consequences for Living Resources and Ecosystems*, vol. 58, AGU, Washington, D. C.
- Rabalais, N. N., R. E. Turner, Q. Dortch, D. Justić, V. J. Bierman Jr., and W. J. Wiseman Jr. (2002a), Nutrient-enhanced productivity in the northern Gulf of Mexico: Past, present and future, *Hydrobiologia*, **475/476**, 39–63.
- Rabalais, N. N., R. E. Turner, and W. J. Wiseman Jr. (2002b), Gulf of Mexico Hypoxia, A.K.A. “The Dead Zone”, *Annu. Rev. Ecol. Syst.*, **33**, 235–263.
- Rowe, G. T., M. E. C. Kaeki, J. W. Morse, G. S. Boland, and E. G. E. Briones (2002), Sediment community metabolism associated with continental shelf hypoxia, northern Gulf of Mexico, *Estuaries*, **25**(6), 1097–1106.
- Siedlecki, S. A., D. E. Archer, and A. Mahadevan (2011), Nutrient exchange and ventilation of benthic gases across the continental shelf break, *J. Geophys. Res.*, **116**, C06023, doi:10.1029/2010JC006365.
- Trowbridge, J. H., and S. J. Lentz (1991), Asymmetric behavior of an oceanic boundary layer above a sloping bottom, *J. Phys. Oceanogr.*, **21**, 1171–1185.

- Trowbridge, J. H., and S. J. Lentz (1998), Dynamics of the bottom boundary layer on the Northern California shelf, *J. Phys. Oceanogr.*, **28**, 2075–2093.
- Turner, R. E., and N. N. Rabalais (1994), Coastal eutrophication near the Mississippi river delta, *Nature*, **368**(14), 619–621.
- Wang, W., W. D. Nowlin, and R. O. Reid (1998), Analyzed surface meteorological fields over the northwestern Gulf of Mexico for 1992–94: Mean, seasonal, and monthly patterns, *Mon. Weather Rev.*, **126**(11), 2864–2883.
- Weiss, R. F. (1970), The solubility of nitrogen, oxygen, and argon in water and seawater, *Deep Sea Res. Oceanogr. Abstr.*, **17**, 721–735.
- Wiseman, W. J., Jr., N. N. Rabalais, R. E. Turner, S. P. Dinnel, and A. MacNaughton (1997), Seasonal and interannual variability within the Louisiana coastal current: Stratification and hypoxia, *J. Mar. Syst.*, **12**, 237–248.
- Yankovsky, A. E., and D. C. Chapman (1997), A simple theory for the fate of buoyant coastal discharges, *J. Phys. Oceanogr.*, **27**, 1386–1401.
- Zhang, X., S. DiMarco, D. Smith IV, M. Howard, A. Jochens, and R. Hetland (2009), Near-resonant ocean response to sea breeze on a stratified continental shelf, *J. Phys. Oceanogr.*, **39**(9), 2137–2155.
- Zhang, X., M. Marta-Almeida, and R. Hetland (2012), A high-resolution pre-operational forecast model of circulation on the Texas-Louisiana continental shelf and slope, *J. Oper. Oceanogr.*, **5**(1), 19–34.
- Zhang, Z., and R. D. Hetland (2012), A numerical study on convergence of alongshore flows over the Texas-Louisiana shelf, *J. Geophys. Res.*, **117**, C11010, doi:10.1029/2012JC008145.