COASTAL OCEAN MODELING ACROSS LENGTH SCALES: FROM ESTUARINE CIRCULATION TO BOUNDARY CURRENTS

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COASTAL OCEAN MODELING ACROSS LENGTH SCALES:
FROM ESTUARINE CIRCULATION TO BOUNDARY CURRENTS

BY

KEVIN L. ROSA

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ABSTRACT

From the shoreline to the continental slope, numerical models of coastal physics provide forecasts of storm impacts, inform environmental policy, and drive key processes in climate predictions. The coastal zone is a highly dynamic region characterized by tides, buoyancy-driven flow, bathymetric steering, upwelling, and boundary currents. This complex mix of competing forces has necessitated the development of increasingly advanced realistic numerical models. Realistic models can be used to analyze ocean physics in ways that would not be possible with observational data alone.

The first two chapters investigate water mass exchange between Narragansett Bay and Rhode Island Sound using current meters and regional models. The first chapter focuses on a single hurricane event which led to an intrusion of cold shelf water into the bay. Not only is this the first study to examine storm-driven residual transports in Narragansett Bay, but it also advances the broader literature on estuarine storm response by addressing the questions of local versus non-local wind forcing as well as the importance of baroclinic effects. Results show that local winds drive most of the exchange flow but non-local winds drive most of the storm surge. Additionally, baroclinic effects were necessary to establish realistic vertical shear, despite strong winds vertically mixing the water column. Chapter 2 looks at estuary-shelf exchange more broadly under a range of wind and tide conditions. Our improved calculations of Narragansett Bay exchange flow are at least two times larger than previous estimates and suggest that offshore inputs are a larger source of nitrogen in the bay than
previously reported. The addition of a simulated dye to track shelf water nutrients reveals substantial variations in the dye concentration of the inflow. We find that southward winds drive the strongest exchange flow but that eastward winds drive shelf upwelling which increases the concentration of dye/nutrients entering the bay.

Chapter 3 looks to the future of climate modeling with a new variable-resolution global ocean model which can resolve small-scale coastal processes that would be too computationally expensive for traditional models. We find that enhancing resolution in a band around the coast of North America improves representation of eastern boundary upwelling but creates unrealistic Gulf Stream behavior. We show that the Gulf Stream gets trapped in the enhanced resolution region and impinges on the Labrador Current which prevents deepwater formation, leading to a dramatic weakening in the thermohaline circulation. We provide recommendations on how to change the mesh to avoid this impingement, and how to improve the mesh design process in the future.
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Finally, thank you to my parents Cheryl and Manny Rosa. They have been unceasingly proud of me my entire life regardless of academic or professional accomplishments. They are the voice reminding me to slow down, be easy on myself, and have fun. Grad school is undoubtedly challenging, and I’ve met students who see it as something to endure until “real life” can start. This is your real life and it’s passing by every minute, so try to enjoy the journey.
DEDICATION

Dedicated to the health care workers around the world who have lost their lives during this ongoing COVID-19 pandemic.
PREFACE

This dissertation is prepared in manuscript format and composed of the following three manuscripts:

Chapter 1: “Hurricane-driven shelf water intrusion into Narragansett Bay” was prepared for submission to *Journal of Marine Science and Engineering* with K. Rosa, C. Kincaid, D. Ullman, and I. Ginis as co-authors.

Chapter 2: “Observed and modeled wind-driven variability in shelf-estuary exchange” was prepared for submission to *Continental Shelf Research* with K. Rosa, C. Kincaid, and D. Ullman as co-authors.

Chapter 3: “Boundary current impacts of coastal refinement in a variable-resolution global ocean model” was prepared for submission to *Journal of Advances in Modeling Earth Systems* with K. Rosa, M. Petersen, S. Brus, D. Engwirda, K. Hoch, M. Maltrud, L. Van Roekel, and P. Wolfram as co-authors.
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MANUSCRIPT 1:

Hurricane-driven shelf water intrusion into Narragansett Bay

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Abstract

We explore water mass exchange at the mouth of the Narragansett Bay estuary in response to 1999’s Hurricane Floyd. An acoustic Doppler current profiler (ADCP) observed enhanced deep inflow of cold shelf water into the bay in response to the storm. The Regional Ocean Modeling System (ROMS) is used to develop a three-dimensional hydrodynamic model which shows good agreement with both the tidal and residual components of the currents at the ADCP. To isolate the effects of the storm, we ran the model with and without storm forcing, and separated the storm forcing into local and non-local wind effects. We found that currents at the mouth were primarily driven by local winds, but surge at the mouth was driven by non-local winds. This demonstrates that storm surge is not a prerequisite for shelf water intrusions, and that these intrusion events are driven by specific local wind conditions. The storm generated a two-stage response of outflow and then inflow, with volume transport of these two stages approximately equal and opposite. Although the storm did not produce a net increase in deep inflow, a simulated dye in the deep shelf water showed that 3-4 times more dye entered the bay in the storm scenario compared to the no-storm scenario. This agrees with the observed drop in temperature, since the deep inflow is colder than the bay. We also investigated whether this exchange event was baroclinic or barotropic. Although the storm-driven currents were forced by a barotropic sea surface height gradient, baroclinicity greatly enhanced the two-layer exchange by decoupling the wind-driven surface layer and pressure gradient-driven lower layer.
1.1 Introduction

1.1.1 Hurricanes and coastal circulation

Storm surge has traditionally been the main focus when considering the effects of hurricanes on coastal oceanography, but storms also cause dramatic changes to estuarine circulation which affect the transport of pollutants, organisms, nutrients, sediment, and more. There are substantial differences in the response of individual estuaries as well as within a single estuary between different storms. Operational storm surge modeling by the US National Weather Service has existed in its current state since the 1990s (Jelesnianski, 1992), yet there is still no real-time system for forecasting storm effects on estuarine circulation.

Current/Available storm surge modeling is performed using 2D depth-averaged ocean models which produce realistic sea surface elevations, but do not represent any depth-varying currents. We can separate ocean models into three levels of increasing complexity: 2D depth-averaged, 3D barotropic (constant density), and 3D baroclinic. Each of these steps represents increased computational costs and configuration challenges. 3D barotropic models have been explored for their potential in improving storm surge predictions via more realistic bottom stress formulation (Weaver & Luettich, 2012; Weisberg & Zheng, 2008; Zheng et al., 2013). To add circulation modeling into our pre-existing storm surge model forecasts, we want the simplest setup that will provide accurate results. Strong winds mix away stratification (M. Li et al., 2006), so are baroclinic effects actually important in the coastal response?
Results are mixed regarding the importance of baroclinic effects for storm-driven coastal circulation. Waterhouse et al. (2013) observed that flow through the weakly-stratified subtropical St. Augustine Inlet became baroclinic in response to two 2008 storms. Significant freshwater input, reinforced by winds, generated several days of post-storm gravitational circulation. Tutak & Sheng (2011) modeled these same storms, however, and concluded that the baroclinic pressure gradient was not a significant component of the momentum equation. In the Chesapeake Bay, the normal baroclinic estuarine exchange was overwhelmed by strong barotropic pressure gradients that even overcame the semi-diurnal flood tide (Valle-Levinson et al., 2002). They noted that this was not necessarily the case for all storms, as there was a depth-dependent baroclinic response to Hurricane Agnes reported by the Chesapeake Research Consortium (1976). One goal of our study will be to present both baroclinic and barotropic results in order to quantify and understand their differences.

Water mass exchange between the estuary and the shelf is a primary concern for the hurricane response and can produce changes in the estuary that last days to weeks (Chen et al., 2018). Strong outflow through the mouth of Chesapeake Bay in response to Hurricane Floyd led to a drop in salinity that took about 10 days to return to pre-storm levels (Valle-Levinson et al., 2002). Hurricane Isabel, however, drove a large intrusion of high salinity shelf water into Chesapeake Bay occupying about one-fifth of the bay’s total volume (Li et al., 2006). Differences between storms were also observed for two hurricanes impacting an estuary in Louisiana. The first storm (Gustav) generated a net outward salt flux but the second storm (Ike) generated a net
inward salt flux (Li et al., 2009). Ike made landfall further away, so the local winds were weaker and the non-local effects comprised a larger relative contribution.

Hurricane winds act on the coastal ocean through both local and non-local (remote) effects. Under normal non-storm forcing, Guo & Valle-Levinson (2008) concluded that non-local winds were the dominant driver of variations in transport through the entrance of Chesapeake Bay, yet local winds drove subtidal flow within the bay. Their realistic 3D model agreed with results from the simple linear model by Garvine (1985) which predicted that the sea surface slope and barotropic current variations within an estuary were produced primarily by local winds. These scenarios mostly consider the role of remote effects via alongshore winds driving Ekman transport on the shelf which can enhance or inhibit exchange through the mouth of the estuary via large scale sea level setup along the coast. Hurricanes, however, produce very different remote wind-effects in the form of storm surge. Cho et al. (2012) described a two-part storm surge in Chesapeake Bay driven by initial setup from remote winds and a second stage driven by local winds. For the two storms they considered (Floyd and Isabel), this local wind-driven second stage had sea surface height (SSH) anomalies with opposite signs due to different wind directions.

The varying physical effects described above produce a range of ecological responses. In Chesapeake Bay, Hurricane Isabel (2003) enhanced plankton and fish abundance immediately following the storm and is thought to be responsible for early-onset hypoxia the following spring (Roman et al., 2005). Paerl et al. (2006) looked at eight hurricanes impacting Pamlico Sound, NC and found dramatic post-storm ecological changes lasting multiple months, and even years in some cases. Each of the
eight storms they examined exhibited individualistic effects on residence time, stratification, oxygen, phytoplankton composition, primary productivity, and eventually higher trophic level abundance and health. One of the main predictors of fisheries effects was the amount of rainfall associated with a storm. High rainfall storms generated large nutrient loading which could enhance some shelf fisheries but had largely negative effects within the estuary, particularly hypoxia and the inflow of toxic chemicals. In contrast, low rainfall storms led to only moderate enhancement in phytoplankton productivity and did not generate the strong vertical stratification that leads to hypoxia.

1.1.2 Hurricane Floyd

Hurricane Floyd first made landfall near Cape Fear, NC with maximum sustained winds of 50 m/s at 0900 UTC September 16, 1999. Maximum winds of 13 m/s in Providence, RI occurred one day later. Warnings highlighted the risk of high winds, yet flooding was ultimately the cause of the majority of loss of life and property (Atallah & Bosart, 2003). Over 20 cm of rain fell on parts of the mid-Atlantic, and over 10 cm on parts of New England. The legacy of Hurricane Floyd is that the storm surprised forecasters with larger than expected rainfall despite its fast translation speed.

In Narragansett Bay, storm surge was minimal, with peak surge of about one meter above the astronomical tide. Maximum water levels were similar to a high spring tide. The most notable aspect of Floyd oceanographically was that it was the
first instance of a tropical cyclone occurring during the deployment of a current meter near the mouth of Narragansett Bay. This provides the unique opportunity to study the effects of the storm on the bay’s shelf-estuary exchange.

1.1.3 Study site: Narragansett Bay

The Narragansett Bay estuary is an important natural and economic resource for Rhode Island and Massachusetts. Like many estuaries, Narragansett Bay has experienced hypoxic events due to eutrophication (Bricker et al., 2008; Deacutis, 2008; Melrose et al., 2007). Maintaining the health of the Bay amid continuous anthropogenic pressures is a top priority for many stakeholders including fisheries, the tourism industry, and preservationists. Effective coastal management requires understanding the Bay’s complex circulation.

Narragansett Bay is a 45 km long, north-south oriented estuary categorized as partially- to well-mixed (Goodrich, 1988). At its southern end, the Bay connects to Rhode Island Sound via three narrow north-south passages (Figure 1-1). The Sakonnet River and the West Passage are both shallow, with mean depths of 7.5 m and 10 m MLLW, respectively (Spaulding & Swanson, 2008). The Sakonnet River has limited exchange with the rest of the bay due to a narrow constriction in the north (Deleo, 2001). The East Passage is the deepest of the three passages and is the main conduit for deep inflow into the Bay (Kincaid et al., 2003). A bathymetric cross-section from Rhode Island Sound into the East Passage reveals a fjord-like morphology with a
relatively shallow entrance (less than 30 m) and a rapid drop to maximum depth of 48 m just inside the bay.

A prevailing anticyclonic pattern exists in the mean depth-averaged flow, with inflow through the East Passage and outflow through the West Passage (Kincaid et al., 2008; Pfeiffer-Herbert et al., 2015; Rogers, 2008). In addition to the anticyclonic transport between the passages, residual flow in Narragansett Bay resembles the characteristic two-layer density-driven estuarine circulation (Hicks, 1959). Winds play a dominant role in modifying the background non-tidal circulation on the timescale of days (Weisberg & Sturges, 1976). Kincaid et al. (2003) used a ship-mounted ADCP to characterize the lateral structure of flow at the mouth of the bay and revealed substantial cross-channel velocity structure. The complex temporally- and spatially-varying patterns of exchange with the shelf and within the bay are important to (1) flush anthropogenically-impacted waters from the upper bay and (2) deliver nutrient-rich water from Rhode Island Sound into the bay.
Figure 1-1. Map view of model domain and observational stations. Upper right panel (a) shows the location of Rhode Island’s Narragansett Bay, east of Long Island Sound and west of Buzzard’s Bay and Cape Cod. Panel (b) has a zoomed-in view of Narragansett Bay with the Providence Tide Gauge station at the head of the Bay indicated with a red square. A further zoomed-in view of the mouth of the Bay is shown in panel (c) along with the locations of the Newport Tide Gauge (orange square) and the Acoustic Doppler Current Profiler (ADCP) (blue triangle). Depth contours plotted at 10 m and 30 m depths.
1.2 Methods

1.2.1 Data collection

The key dataset that motivated this study is the velocity data collected by an Acoustic Doppler Current Profiler (ADCP) located near the mouth of Narragansett Bay. This ADCP station has been included in previous studies but no one has investigated the response during 1999’s Hurricane Floyd (Kincaid et al., 2008; Pfeiffer-Herbert et al., 2015; Rosenberger, 2001). The ADCP was located in Narragansett Bay’s East Passage at 41°30.33’N, 71°21.08’W (Figure 1-1). At 40 m depth, the site was chosen at approximately the deepest point in the channel at this latitude. The current profiler was a 300 kHz, 4-beam self-recording RD Instruments ADCP which collected water velocity data at 2-meter vertical bins. We removed bins within 5 m of the surface due to poor data quality. Data were obtained every 6 minutes by averaging a 10-second, 10-burst ensemble. A thermistor attached to the ADCP housing measured near-bottom water temperature at the same 6-minute interval.

Tide gauge observations of the storm surge in Narragansett Bay were obtained by the two NOAA stations operating at the time: Newport (station 8452660: 41°30.2’ N, 71°19.6’W) and Providence (station 8454000: 41°48.4’ N, 71°24.0’W). Wind data are not available at these stations until October 1999 (after the passing of Floyd) but wind data were recorded at TF Green Airport (41°43.2’ N 71°25.8’ W) in the upper bay and the Buzzard’s Bay NOAA station (41°23.8’ N 71°2’ W) 30 km east of the Narragansett Bay entrance.
1.2.2 Model configuration

A two-model nested approach was used to model the effects of Hurricane Floyd in Narragansett Bay. To model the three-dimensional circulation response within the Bay, we used the Regional Ocean Modeling System (ROMS) finite difference model (Rutgers version 3.6, Haidvogel et al., 2008; Shchepetkin & McWilliams, 2005). A large model domain is needed to produce realistic storm surge (Blain et al., 1994; Morey et al., 2006) so we forced the lateral boundaries with elevations and velocities from a basin-scale 2-D depth-integrated ADCIRC model (Luettich et al., 1992). The ADCIRC model was an ideal choice for this application both because it is computationally efficient and because it is an operational storm surge model used by the National Hurricane Center, which means that our ROMS-ADCIRC nesting system could potentially be implemented for future storms in real-time (Fleming et al., 2008).

1.2.2.1 Parent domain: 2D ADCIRC

The ADCIRC finite element model covers the western North Atlantic and is used to compute the “non-local” storm surge and depth-integrated velocities associated with Floyd (Figure 1-2). Atmospheric forcing was obtained from the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA-Interim reanalysis at 1/8° spatial resolution, output at a 3 hour timestep (Dee et al., 2011). We converted the 10 m wind speed into wind stress using the formulation presented by Large & Pond (1981).

Tidal forcing was generated using the US Army Corps of Engineers ADCIRC tidal database (Szpijka et al., 2016). We removed the tidal signal from the ADCIRC
output by running the model twice – once with tides-only and once with tides and atmospheric forcing – and then differenced the two solutions.

Figure 1-2. Full ADCIRC model domain (a) and zoomed-in view highlighting enhanced resolution around Narragansett Bay and Rhode Island Sound (b). ADCIRC model cells are indicated by blue points and Hurricane Floyd track indicated by black and yellow line. The hurricane track is defined as the locations of minimum sea level pressure in the atmospheric model.

1.2.2.2 Child domain: 3D ROMS

Our ROMS model domain (Figure 1-3) covers approximately 90 km x 80 km and was designed to include all of Narragansett Bay and extend out into Rhode Island Sound to track shelf water that enters the Bay. The spatial resolution of the grid determines the scales of processes that the model can represent: finer resolution can produce more realistic simulations but increases the computational cost of running the
model. Across-channel cross-sections in the East Passage reveal significant lateral and vertical structure in the along-channel velocities (Kincaid et al., 2003). We used 150 m cross-channel resolution at the mouth to resolve cross-channel flow structure through the East Passage. Grid spacing varies from a minimum of 47 m in the upper Bay to a maximum of 358 m at the southern open boundary. The grid is curvilinear with 325x425 grid cells in the horizontal directions and 14 terrain-following sigma layers in the vertical direction, with enhanced vertical resolution near the surface.

Figure 1-3. ROMS model domain boundaries are indicated by the blue line. Red area indicates the spatial extent of the deep “dye” initialized in the model. Thick black line shows the location of the cross-section used in Figure 1-16.

Bathymetry data from NOAA’s U.S. Coastal Relief Model (National Geophysical Data Center, 1999) were interpolated to grid cells using bilinear interpolation.
Bathymetry was then smoothed for numerical stability using the LP Bathymetry program (Sikirić et al., 2009) set to a minimum Beckmann and Haidvogel number (Beckmann & Haidvogel, 1993) of 0.2 as recommended by Shchepetkin & McWilliams (2003). This bathymetry smoothing method provided numerical stability without removing key features such as the dredged shipping channels.

In our model configuration, each cell is defined as either “land” or “water” – there is no wetting and drying of cells. This approach offered increased numerical stability and decreased computational cost. Wetting and drying was not a focus of this study since Floyd did not cause coastal flooding in Narragansett Bay. The grid has 80,852 water cells and 57,273 land cells. A minimum depth of 2 m was applied to all water cells to avoid drying in the intertidal regions.

To force the surface boundary, momentum flux (wind stress), salt flux, short- and long-wave heat fluxes, and atmospheric pressure were applied. These input fields were obtained from the same ECMWF ERA-Interim reanalysis product used for the ADCIRC wind forcing. The coarser ~13 km resolution of the atmospheric model did not resolve the land-sea boundaries of Narragansett Bay and resulted in unrealistic surface heat and salt fluxes in the Bay. For this reason, spatially-uniform heat and salt fluxes were applied across the ROMS domain based on fluxes at the ERA-Interim water grid node located at 41°N, 71°W in Rhode Island Sound. Wind stress and atmospheric pressure were allowed to vary spatially across the domain. Wind stress was calculated from the ECMWF 10 m winds using the same formulation (Large & Pond, 1981) used by the ADCIRC model.
The lateral open boundary forcing was a linear superposition of three components: (1) tides, (2) non-tidal temperature and salinity, and (3) de-tided Hurricane Floyd SSH and velocities. Tidal forcing data was obtained from the TPXO8-atlas 1/30° global tide model (Egbert & Erofeeva, 2002) which makes use of satellite data from TOPEX/Poseidon and Jason. Elevations and transports for the M2, S2, N2, K2, K1, O1, P1, Q1, and M4 tidal constituents were included. Temperature and salinity fields at the lateral boundaries were obtained from the GOFS 3.0 HYCOM Global 1/12° Reanalysis (Metzger et al., 2014) and applied using a radiation boundary condition with nudging. De-tided ADCIRC depth-integrated velocities were applied using the Flather (1976) boundary condition and free-surface perturbations were applied using the Chapman (1985) boundary condition.

Freshwater point sources were included in the model to simulate river discharge from seven major rivers emptying into Narragansett Bay: the Blackstone, Moshassuck, Woonasquatucket, Pawtuxet, Taunton, Ten Mile, and Hunt Rivers. River discharge data were obtained from the United States Geological Survey. For the simulated time period (1999), discharge data is available only for the Taunton, Woonasquatucket, and upper Blackstone Rivers. Using discharge data from recent years, we trained a linear a stepwise regression model in order to estimate discharge at the un-gauged rivers in 1999.

The temperature and salinity fields in the model were initialized using a 6-month spin-up run. The spin-up was forced with tides, rivers, and reanalysis surface forcing for 1999 in order to establish realistic stratification and gravitational circulation.
Finally, a passive “dye” tracer was added to the initial conditions to track the transport of dense shelf water. The dye was initialized in the region shown in Figure 1-3. Dye concentration was set to 1000 kg/m3 in the bottom 6 sigma layers for cells with depth greater than 28 m. This dye is a simple proxy for deep offshore nutrients originating in Rhode Island Sound.

1.2.2.3 Model runs

We used a set of different ROMS model configurations to separate out the effects of barotropic tides, gravitational circulation, local winds, and non-local winds. Table 1-1 shows the list of runs and their forcing configurations. We then difference the velocities and SSH from different runs in order to isolate the effects of local winds, non-local winds, etc.

To calculate the non-tidal gravitational circulation, we subtract the barotropic tidal signal (run200) from the no-storm baroclinic model (run300). To isolate the contributions of hurricane local winds vs. non-local winds, run301 has local winds but no non-local boundary forcing, and run310 has non-local boundary forcing but no local winds. Velocities and SSH from run300 can be subtracted from run301 and run310 to remove tides and background gravitational circulation. The realistic case run311 includes both local and non-local wind effects.

In addition to the tides-only run200, we also ran barotropic simulations with local (run201), non-local (run210), and realistic (run211) forcing in order to assess the importance of barotropic vs. baroclinic effects in the bay’s storm response.
Table 1-1. ROMS model run configurations. Different combinations of forcings were applied to the models in order to isolate the effects of barotropic tides, gravitational (i.e. estuarine) circulation, local winds, and non-local wind.

<table>
<thead>
<tr>
<th>Run code</th>
<th>Description</th>
<th>Stratification</th>
<th>Local wind forcing</th>
<th>Boundary forcing</th>
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</thead>
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<tr>
<td>run200</td>
<td>no-wind barotropic tides</td>
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<td>no</td>
<td>tides only</td>
</tr>
<tr>
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<td>local winds (barotropic)</td>
<td>no</td>
<td>yes</td>
<td>tides only</td>
</tr>
<tr>
<td>run210</td>
<td>non-local winds (barotropic)</td>
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<td>no</td>
<td>tides + ADCIRC surge</td>
</tr>
<tr>
<td>run211</td>
<td>full storm (barotropic)</td>
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<td>yes</td>
<td>tides + ADCIRC surge</td>
</tr>
<tr>
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<td>no-wind gravitational exchange</td>
<td>yes</td>
<td>no</td>
<td>tides only</td>
</tr>
<tr>
<td>run301</td>
<td>local winds (baroclinic)</td>
<td>yes</td>
<td>yes</td>
<td>tides only</td>
</tr>
<tr>
<td>run310</td>
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<td>tides + ADCIRC surge</td>
</tr>
<tr>
<td>run311</td>
<td>full storm (baroclinic)</td>
<td>yes</td>
<td>yes</td>
<td>tides + ADCIRC surge</td>
</tr>
</tbody>
</table>

1.2.3 Analysis methods

1.2.3.1 Rotate velocities

Velocity data at the ADCP site were rotated into “up-bay” and “cross-channel” coordinates for the analysis. The up-bay direction is defined as the direction of maximum velocity variance, using all good vertical bins for the full ADCP deployment. Using this method, the up-bay direction for the ADCP data was found to be 24.2° east of north. This is consistent with the angle of the channel’s isobaths near the site.

The same method was used to compute the up-bay direction for the modeled velocities, which was found to be 8.6° east of north. While we expect bathymetric steering to result in roughly along-isobath flow, the model isobaths are limited by the
resolution of the grid, so some disagreement between the model and observations was expected. The 8.6° up-bay direction is similar to the 12.4° angle of the model grid cells in the area of the ADCP site.

1.2.3.2 Volume transport

When calculating the modeled deep transport into the bay, a cross-channel East Passage section was defined at the latitude of the ADCP site and flux was calculated for cells deeper than 10 m below mean sea level. The deep transport was used instead of the full water column depth-integrated transport in order to preserve the bi-directional transports. This depth cut-off approach is different than the usual method for computing the Eulerian sub-tidal volume transport. Normally, we would let the inflow portions of the cross-section vary in time. Instead, the 10 m cut-off will include some outflow cells at times, and at other times will miss some inflow cells. However, this allows for clearer comparisons between different runs, particularly with barotropic runs which don’t have stronger vertical shear.

1.2.3.3 Willmott skill score

When quantifying model-observation agreement, we use the “Willmott skill score” as defined by Willmott (1981), which has been a popular metric in other estuarine modeling studies (e.g. Warner et al., 2005). The skill formulation is seen in Equation 1 where, for any prognostic quantity, \( M \) represents the model’s predicted value and \( O \) represents the observed value. Skill can vary between 1 (perfect agreement) and 0.
$\text{Skill} \equiv 1 - \frac{\sum_{i=1}^{N} (M_i - O_i)^2}{\sum_{i=1}^{N} (|M_i - \bar{O}| + |O_i - \bar{O}|)^2}$

1.2.3.4 Storm surge

We follow the definition of “storm surge” as the sea surface height minus the expected astronomical tide. For tide gauge observations, we use the NOAA-provided predicted SSH as the astronomical tide to subtract from the observed SSH. In the numerical results, since there are slight differences in the astronomical tides between the real-world and the model, we obtain our astronomical SSH from a tides-only model run and then follow the same process of subtracting this SSH from the SSH under the various storm forcing configuration to get the modeled storm surge.

1.2.3.5 Subtidal/nontidal filter

Residual, de-tided velocities were obtained in two different ways. To compute “subtidal” velocities, we applied a 33-hour Butterworth low-pass filter to the data. This was done on ADCP observations but also some model results when used for model-data comparisons. “Nontidal” velocities where only the tidal signal was removed (preserving both low-frequency and high-frequency nontidal variability) were computed for model results only. Nontidal velocities were determined by subtracting the velocity from a tides-only barotropic model run. These nontidal velocities were desirable for analyzing the high-frequency effects of the storm on velocities since the peak surge occurred over less than one day and the signal would have been altered by a 33-hour lowpass filter.
1.3 Results

1.3.1 Observational data

Subtidal flow at the ADCP was directed up-bay throughout most of the deployment (Figure 1-4) but the arrival of Hurricane Floyd on September 17, 1999 generated a strong subtidal outflow event lasting approximately one day. The outflow event was followed by two days of slightly enhanced deep inflow and a sharp drop in bottom temperature of 3.8°C over 2.25 days. This was the largest temperature change over any two-day period in the deployment.

![Figure 1-4. ADCP data near the mouth of the Narragansett Bay. 33-hour lowpass-filtered velocities are shown in (a) and bottom temperature at the ADCP is shown in](image-url)
(b). Velocities have been rotated into up-bay (roughly along-channel) direction in which positive values indicate flow directed into the bay. The ADCP bins have 2 m spacing and are represented by circles in (a). Additionally, the sea surface height observations from nearby Newport, RI NOAA station are shown as a black line in (a). The dashed lines have been added to demarcate the different phases of the storm response.

Maximum observed storm surge at the NOAA tide gauges was 1.1 m at Providence in the upper bay and 0.76 m at Newport near the mouth of the bay (Figure 1-5). The maximum storm surge arrived after low tide when the predicted astronomical tide was approximately 0 m above mean sea level. The de-tided sea surface height in the bay reached a maximum on Sep-17 at approximately 2:00 UTC, then a minimum 15 hours later at 17:00 UTC, and then steadily increased over the following 2+ days until returning to mean sea level on Sep-20.
Figure 1-5. Sea surface height observations at Providence (a-b) in the upper bay and Newport (c-d) near the mouth. Red dashed lines indicate the timing of maximum surge. Maximum surge occurred shortly after low tide when the astronomical tide was slightly below mean sea level and therefore limited the magnitude of the storm tide.

1.3.2 Model validation

The ROMS model with full realistic forcing (run311) was used to evaluate the skill of the numerical model compared to observations. For velocity comparisons at
the ADCP location, both the observations and the model velocities were vertically averaged across bins deeper than 10 m below the surface. The instantaneous (not lowpass-filtered) along-channel model velocities (Figure 1-6a) had a Willmott skill score of 0.946 and the subtidal (lowpass-filtered) along-channel model velocities (Figure 1-6b) had a Willmott skill score of 0.948. The modeled residual velocities had larger magnitude than the ADCP data, both for the strong outflow event and the inflow.

![Figure 1-6. Velocities averaged across depths below 10 m and rotated into up-bay coordinated. Instantaneous velocity (a) and 33-hour lowpass filtered velocity (b). Black line indicates observations from ADCP and blue line is from ROMS model.](image)

The vertical structure of the observed and modeled velocities can be seen in Figure 1-7. Both the ADCP and ROMS velocities show a slight time lag in the vertical
structure of the subtidal outflow event, with outflow beginning and ending earlier in the lower-watercolumn than the upper-watercolumn. The inflow event is characterized by intensification in the lower- to mid-watercolumn with velocities weakening towards the upper-watercolumn. As shown first in Figure 1-6b, the model inflow velocities are stronger than the ADCP observations.

Surface outflow is not detected within the ADCP bins with high data quality, though some outflow velocities are seen in the top bin at times when the surface outflow layer presumably deepens. Based on our calculation of the vertical position of the ADCP bins, it appears that the ROMS model predicts a thicker surface outflow layer.
Figure 1-7. ECMWF model wind vectors (a), ADCP up-bay lowpass-filtered velocities (b), and ROMS model up-bay lowpass-filtered velocities (c).

Maximum model storm surge was only about 60% as large as the observed values — 0.65 m at Providence and 0.46 m at Newport (Figure 1-8). The arrival time of the storm surge agreed with observations but had a longer duration. The observed storm surge had a brief peak, but in the model the maximum surge at both locations persisted for five hours. After the surge, the sea level minimum and subsequent return to normal was well predicted by the model, both in timing and magnitude.
The final model-data comparison is the bottom temperature at the ADCP site (Figure 1-9). Both the model and observations showed rapid warming shortly after Sep-17-1999 0:00 UTC, but the modeled warming is more than twice as large as the observations. The initial warming is likely caused by a combination of vertical mixing and down-bay advection. Model over-prediction indicates problems with one or more of the following: (1) vertical temperature profile at ADCP, (2) vertical mixing, (3) along-channel temperature gradient. Based on the timing of the warming, it seems most likely that the problem is some combination of the vertical temperature profile being too warm in the upper water column and/or over-predicting the vertical mixing.
which mixed this heat down the bottom. The timing of the temperature decrease agrees well with observations: it starts at Sep-18 4:00 UTC and reaches a minimum at Sep-20 7:00 UTC.

![Near-bottom water temperature anomaly at ADCP site, relative to temperature at Sep-15, 1999. Comparing thermistor observations (black line) and ROMS model (blue line).](image)

### 1.3.3 Comparing model configurations

#### 1.3.3.1 Volume transport

The volume transport into the bay for an across-channel section at the ADCP has been separated into its component parts in Figure 1-10. Subtidal transports are integrated through depths below 10 m. The background gravitational circulation is always directed up-bay and contributed a mean transport of 1,569 m$^3$/s during the 6-day period Sept 16-22. Over the 6-day period, the gravitational transport varied
between 2,837 m³/s and 190 m³/s. Gravitational circulation varied throughout the tidal cycle, with maximum exchange during flood tides.

The storm altered the exchange through the mouth of the East Passage via both local and non-local wind effects. The local winds were the largest driver of the outflow event, transporting 331 x 10⁶ m³ out of the bay. This was over two times greater than the gravitational up-bay transport during the same period. Non-local winds contributed 45 x 10⁶ m³ to the outflow. The total non-tidal transport below 10 m for the outflow event was 238x10⁶ m³. For reference, the volume of the bay is 2700x10⁶ m³ (Pilson, 1985) so the non-tidal outflow through the East Passage (below 10 m) flushed nearly 10% of the total volume of the bay.
Figure 1-10. ROMS model deep transport (below 10 m) through mouth of East Passage, separated into forcing components: barotropic tides (a), no-wind non-tidal estuarine exchange (b), local winds (c), non-local winds (d), and full de-tided exchange (e). Integrated volume transports are added (units $10^6$ m$^3$) for the outflow and inflow time periods.
The subsequent inflow was approximately double the normal gravitational transport. Separating the storm-driven fluxes into local and non-local wind components, the local winds were once again the larger contributor, generating 2.5 times greater volume transport than the non-local winds. Summing the separate transport components (gravitational, local winds, and non-local winds) resulted in a net transport of $660 \times 10^6$ m$^3$ into the bay which is 8% greater than the actual modeled non-tidal transport of $610 \times 10^6$ m$^3$. This means that the exchange flow from the separate models was not an exact linear superposition of the flow in the fully-forced model. In both the outflow and the inflow events, the local winds were a larger driver of currents than the non-local wind effects.

Summing the storm-induced inflow and outflow (remove tides and gravitational circulation) results in a weak net outflow of $76 \times 10^6$ m$^3$. However, the modeled dye initialized in the deep shelf water outside of the bay reveals 3-4 times greater transport of dye into the bay under storm forcing than the no-storm control case (Figure 1-11). Four days after Floyd on Sept 21, approximately 8% of the bay’s water originated in the offshore dye patch, compared to only about 2% under the no-storm conditions.
1.3.3.2 Storm surge

Separating the local and non-local forcing components for SSH reveals that the non-local winds generated the larger sea level anomalies (Figure 1-12). At Newport station, near the mouth of the bay, the storm surge was predominantly driven by non-local winds – 0.38 m max surge generated by non-local forcing vs. 0.12 m from local (Table 1-2). At Providence station in the upper bay, the non-local winds contributed a peak storm surge of 0.41 m compared to 0.31 m generated by the local winds. The post-storm SSH set-down at both locations was caused almost entirely by non-local effects. The non-local SSH minima at both stations was similar in magnitude to the
peak maximum SSH. Under local-forcing, the maximum SSH set-down was less than half the magnitude of the set-up.

Figure 1-12. De-tided sea surface height at Providence (a) and Newport (b) from ROMS model experiments. The different forcing conditions shown are local winds only (orange), non-local winds only (green), and full local and non-local winds (blue).
Table 1-2. ROMS model maximum and minimum de-tided SSH anomalies at Newport and Providence tide gauges under non-local and local wind forcing.

<table>
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<tr>
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<th>Newport</th>
<th>Providence</th>
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<tbody>
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<td>Local</td>
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<tr>
<td><strong>Min</strong></td>
<td></td>
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<tr>
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<td>-0.36 m</td>
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<tr>
<td>Local</td>
<td>-0.05 m</td>
<td>-0.14 m</td>
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</table>
1.4 Discussion

1.4.1 Local vs. non-local wind effects

ROMS model results revealed dramatic differences between the effects of local and non-local forcing. Initially, we assumed that the storm surge and the intrusion currents were generated by the same physical forcings and were two manifestations of a single phenomenon. Looking only at observational data (Figure 1-13) provided a misleading picture. We can see that when the SSH is falling after the peak surge, velocities are directed out of the bay. Next, the timing of the SSH minimum and two-day period of SSH returning to normal also corresponds with enhanced flow into the bay. Since the surge and the currents are correlated in time, we assumed that they were driven by the same forcing, i.e. that we would not be able to produce the transports without producing the surge or vice versa.
Figure 1-13. Timing of observed sea surface height anomaly at Newport tide gauge (a) and lowpass-filtered up-bay deep velocity at ADCP (b). This agreement between the timing of decreasing (increasing) SSH and outflow (inflow) led to the initial hypothesis that the surge and currents were driven by the same forcing/mechanism.

Instead, we have shown that we can generate 83% of the maximum storm surge at the mouth of the bay via only the non-local winds, and that the local winds drive the majority of the storm-induced transports (88% of the initial outflow and 72% of the post-storm inflow). The intuition that the changing SSH should be accompanied by a corresponding response in the currents to conserve volume was confirmed in the non-local model (Figure 1-10d), but this transport was found to be of smaller magnitude than the local wind-driven exchange flow. It is an interesting coincidence that in this case the non-local and local effects on currents had the same sign and similar timing.

We could imagine another scenario where the storm passes to the east of Narragansett Bay instead of to the west. In this case, we might expect similar non-local effects on the SSH and currents but the meridional component of the local winds would be reversed. This is similar to the two-stage non-local and local surge in Chesapeake Bay described by Cho et al. (2012). In their analysis of two storms with different tracks, both drove a positive storm surge via non-local effects but had opposite effects on local wind-driven SSH set-up due to different local wind directions (Figure 1-14).
Figure 1-14. Winds in Chesapeake Bay during Hurricanes Floyd and Isabel, compiled from Cho et al. (2012). Storm tracks are shown in (a) for Floyd (red) and Isabel (blue). Timeseries of local wind vectors are shown for Floyd (b) and Isabel (c).

The storm surge at Newport station (in the lower bay) was primarily the result of non-local winds, but at Providence station (in the upper bay) the surge was a combination of local and non-local wind effects. The key to this distinction is the SSH setup within the bay (Figure 1-15). At the time of maximum storm surge (Sep 17 0200 UTC) local winds were directed northward which led to water piling up in the upper bay. The non-local surge was similar at both stations, but the added influence of the local wind-generated SSH setup lead to the larger maximum surge in Providence.
Figure 1-15. De-tided sea surface height (SSH) at the timing of approximately peak SSH in the Bay for local wind case (a), non-local wind case (b), and realistic local+non-local wind case (c). Contours are drawn every 0.02 m. The SSH was de-tided by subtracting the SSH from a tides-only model run. The non-local winds (b) raise the sea level of the entire Bay approximately evenly and explain most of the sea level rise, especially in the southern part of the Bay. The local winds (a) in contrast exhibit strong spatial gradients indicated by the close spacing of SSH contours and the piling up of water in the north of the Bay (due to the northward blowing winds at this time in the storm’s passing). The sum of (a) and (b) results in the realistic SSH results seen in (c), i.e. the SSH gradients of (a) but with sea level lifted up by the near-uniform SSH anomaly from (b). The large SSH gradients in (a) explain why the local winds had a larger effect on currents than the non-local winds (since these SSH gradients generate pressure gradients which drive flow).

The de-tided SSH field in Figure 1-15 also explains why the local winds drove the majority of the transport at the mouth of the bay. The mean SSH across the bay is higher in the non-local wind case, but the SSH gradients are much larger in the local wind case, as seen by the close spacing of SSH contours (Figure 1-15a). The local wind-driven SSH setup generates stronger barotropic pressure gradients compared to
the non-local case where the SSH is relatively flat and the resulting gradients are weak. This is because the non-local storm surge, despite having larger amplitude, also has a longer wavelength and therefore the SSH slope is minimal.

By comparing the wind arrows and the subtidal velocities (Figure 1-7), we can see how the flow was strongly correlated with the along-estuary (meridional) component of the wind. Our Floyd observations and model results agree with the theoretical predictions: up-bay (northward) winds stall/reverse the normal estuarine exchange and down-bay (southward) winds enhance it. Pfeiffer-Herbert et al. (2015) discussed shelf intrusions driven by this two-stage up-bay to down-bay wind sequence for Narragansett Bay under normal non-storm wind speeds.

One aspect that was not investigated is the role of spatially-varying winds within Narragansett Bay. How might our results have changed if we had used uniform winds, or had used a higher resolution wind model with more accurate spatially-varying winds? Based on the north-south oriented geometry of the bay, we expect that wind shadows would have the most significant effect on zonal winds, leading to decreased wind speeds compared to the shelf. Meridional winds would likely be most impacted in the upper bay. Although our analysis focused on flow through the mouth of the bay, future investigations might consider how spatially-varying winds affect local wind-driven transports and surge in the upper bay.
1.4.2 Shelfwater intrusion

If we only consider the net volume transport, we would conclude that the storm generated no net increase of shelfwater in the bay, since the storm-driven outflow and inflow events were approximately equal and opposite. However, model dye results show that the storm initiated an intrusion of shelfwater which resulted in 3-4 times more shelf dye in the bay under storm forcing than no-storm forcing (Figure 1-11). This agreed qualitatively with the observational data which showed a decrease in bottom water temperature at the ADCP. Vertical mixing from the storm would cause the bottom temperature to increase, not decrease, so we hypothesized that the cooling was instead due to enhanced on-shore advection of cool shelf water.

Since shelf dye transport increased but net volume transport did not, there must have been an increase in the concentration of dye in the inflow. We propose that the higher proportion of shelfwater in the inflow was the result of upwelling-favorable winds on the shelf as well as enhanced vertical mixing on the shelf. Referring again to the wind arrows in Figure 1-7a, we can see that the strongest winds were directed eastwards, which should drive offshore Ekman transport and thus deep upwelling on the shelf outside the bay (Pfeiffer-Herbert et al. 2015). 24 hours of strong eastward and southeastward winds starting Sep-17 12:00 drove upwelling on the shelf and transported dye to the mouth of Narragansett Bay. Cross-sections extending onto the shelf show enhanced on-shore velocities in the storm case (Figure 1-16f-h).

Additionally, wind forcing generated turbulence which vertically mixed the water column on the shelf. The entrance to the bay’s East Passage features a shoaling bathymetry that culminates at about 20 m depth before quickly dropping to over 40 m
inside the bay. Vertically mixing shelf dye to above 20 m depth increased concentrations in the mid-depth inflows, compared to the no-storm case where dye was constrained deeper in the water column.

The eastward, upwelling-favorable winds, as well as vertical mixing, created ideal conditions for a shelfwater intrusion into the bay. Model dye results confirmed our hypothesis of increased shelf inflow based on the observed drop in bottom temperature at the ADCP. A simple velocity or volume transport analysis would not have detected this period as an anomalously strong inflow event. Using a numerical model to couple shelf upwelling with transports at the mouth of the bay, we identified an intrusion event that could have important implications for ecology in the bay. Since this shelf water contains elevated levels of dissolved inorganic nitrogen and phosphorous (S. W. Nixon et al., 1995), we predict that Floyd caused a large increase in nutrients entering the bay from offshore which could produce phytoplankton blooms as seen in other regions after storm-driven shelf intrusions.
Figure 1-16. Along-channel cross-sections for the transect line shown in Figure 1-3, left to right moves south to north from Rhode Island Sound in Narragansett Bay. Comparing no-storm case (left column, a-d) and storm case (right column, e-h). Grey contour lines indicate isopycnals drawn every 0.5 kg/m³. Light blue arrows are 33-hour lowpass-filtered northward velocities. Color is dye concentration. Both models were initialized with the same deep dye patch. Blue triangle indicates location of the ADCP in the lower East Passage.
The dye results support the hypothesis that a shelf intrusion caused the observed bottom water cooling at the ADCP, but can we eliminate the alternative hypothesis that the cooling was caused by air-sea heat fluxes? When we estimate the heat flux required to cool the bottom water by 3.8 °C over 2.25 days, we find that the heat flux would have to be unrealistically large. For this estimate, we can ignore vertical temperature gradients and instead assume the water column is fully mixed at all times. This means that heat is instantly exchanged between the bottom and the atmosphere, giving us the maximum theoretical efficiency for cooling the bottom and thus the minimum necessary heat flux. Even under these idealized conditions, the 40 m thick water column would have to flux over $3 \times 10^3$ W/m$^2$ into the atmosphere via latent and sensible heat throughout the 2.25 day cooling period. This heat flux magnitude is larger than any estimates in the literature, including extreme hurricane events (Lin et al., 2009).

1.4.3 Baroclinic vs. barotropic effects

In section 1.4.1, we showed that the storm’s influence on non-tidal bottom velocities can be explained by SSH setup in the bay which imposes a barotropic pressure gradient. Are baroclinic effects important for the storm-induced transports, or could we have obtained similar results using an un-stratified (barotropic) model instead? If baroclinicity is not important to the storm-induced response, then we would expect the transports to simply be the sum of the barotropic model transport plus the background gravitational circulation.
In Figure 1-17 we recreated the volume transport calculations from Figure 1-10 using a barotropic 3D configuration. There is no estuarine gravitational exchange in the barotropic model since that exchange is driven by baroclinic pressure gradients, but the general storm-driven outflow and inflow events appear to be present. Figure 1-18 plots the difference between the baroclinic and barotropic models. The barotropic model produces smaller magnitude deep transports under local wind forcing (Figure 1-18a-c). The non-local wind transports are similar for the two models.

The local wind-driven deep transport is enhanced in the baroclinic model because the stratification reduces vertical mixing of momentum (increases shear) which decouples the wind-driven surface layer from lower layer. The large outflow event was forced by up-bay winds which drove surface water into the bay and set up a SSH gradient which drove deep water out of the bay. The unstratified model does not produce the vertically-sheared bi-directional flow, and as a result there is some canceling out between the competing up-bay wind stress and the down-bay barotropic gradient. We can see this clearly when, instead of looking at the deep transport, we integrate through the entire water column (Figure 1-18d-f). Now the differences between the baroclinic and barotropic local wind-driven transports nearly disappear.
Figure 1-17. Barotropic model transport components similar to Figure 1-10. Deep transport (below 10 m) through mouth of East Passage, separated into forcing components: barotropic tides (a), no-wind non-tidal estuarine exchange (b), local winds (c), non-local winds (d), and full de-tided exchange (e). Integrated volume transports are added (units $10^6$ m$^3$) for the outflow and inflow time periods.
The two-layer wind-driven flow in the baroclinic model was a surprise since we expected that strong wind-driven vertical mixing would destratify the water column and lead to a fully barotropic response. However, cross-sections reveal that the well-mixed water column led to strong horizontal density gradients because of the longitudinal estuarine salinity gradient. These vertical isopycnals were unstable and rapidly tilted to establish vertical stratification (Figure 1-16f). Rapid gravitational adjustment and re-stratification in the presence of a horizontal density gradient has been shown previously both for general fluids (Simpson & Linden, 1989) and for an estuary in response to a hurricane (M. Li et al., 2006).

Figure 1-18. Difference in East Passage transport (m³/s), baroclinic model minus barotropic model. Deep transport below 10 m depth (a-c) and full water column (d-f).
Figure 1-19. Mean along-channel velocities in the baroclinic model (top) and barotropic model (bottom). Cross-sections show flow through the West and East Passages, with the land between the two channels excluded from the figure. The baroclinic model demonstrates two-layer estuarine circulation with inflow in the lower layer and outflow in the upper layer. The barotropic model shows lateral but not vertical shear in the along-channel velocities.

Baroclinic effects also alter sea surface height, with the tendency to result in increased storm surge. Comparing maximum de-tided storm surge from the baroclinic and barotropic models, the baroclinic model has larger maximum surge of between 2 cm to 10 cm in the bay (Figure 1-20). In the upper bay, this results in an increased storm surge of approximately 10%. The main mechanism behind this is thought to be
differences in bottom stress, since baroclinicity changes the vertical structure of the currents (R. L. Gordon, 1982). Despite the less-realistic physics, 3D storm surge studies have typically used barotropic 3D models (Weaver & Luettich, 2012; Weisberg & Zheng, 2008; Zheng et al., 2013). (Bode & Hardy, 1997) predicted that improvements in affordable computing would lead to the wide-spread adoption of baroclinic storm surge models but, over 20 years later, this prediction has still not been realized. (Kodaira et al., 2016) modeled global storm surge using a baroclinic model and found that the inclusion of stratification improved model-data agreement at nearly all of their 257 tide gauges. The recent study by (Ye et al., 2020) also showed improvements in storm surge when using a baroclinic model.

Figure 1-20. Difference in maximum storm surge, baroclinic model minus barotropic model. The baroclinic model shows between 2 cm and 10 cm higher maximum storm
surge everywhere in the bay, with largest differences at the two main freshwater sources: the Providence River and the Taunton River. Surge was calculated by subtracting the sea surface height from the barotropic tide-only model (run200).
Conclusions

In this study, we examined non-tidal transport at the mouth of Narragansett Bay in response to 1999’s Hurricane Floyd. We first compared the numerical model against observations to determine our confidence in the model results. The model showed good agreement with the timing and magnitude of instantaneous and subtidal along-channel deep velocities at the ADCP site. Peak storm surge was underpredicted in the model but the post-storm SSH set-down was well represented. We then ran a series of model configurations turning on/off local wind forcing, non-local wind effects, and stratification in order to determine which processes drove different aspects of the storm response. The most significant contributions from this study can be separated into three main findings:

1. Currents were primarily driven by local winds. Surge at the mouth was driven by non-local winds but the contribution from local wind-driven SSH setup increases as you move north up the bay.

2. Observations show intrusion of cold shelf water into the bay and model dye studies confirm that 3-4 times more shelf dye entered the bay under storm forcing than in the no-storm simulation. Volume transport shows no net deep inflow. Winds on the shelf drove upwelling and vertical mixing which increased dye concentration of East Passage inflow.

3. Baroclinic effects were necessary to represent the shelf water intrusion, despite strong vertical mixing. Deep non-tidal flow was opposite the direction of along-channel winds. The baroclinic model generated realistic two-layer wind-driven flow. The barotropic model agreed with the depth-integrated
transport, but could not represent the bi-directional flow necessary for driving the non-tidal transport.
MANUSCRIPT 2:

Observed and modeled wind-driven variability in shelf-estuary exchange

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To be submitted to Continental Shelf Research

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Abstract

Subtidal estuarine exchange at the estuary-ocean interface controls the residual transport of sediment, plankton, pollutants, and nutrients, yet it is poorly understood for Narragansett Bay. We present new calculations of Narragansett Bay exchange flow based on a high-resolution 3D numerical model that is validated by four current meters. Previous estimates in the literature were based on salt conservation and predicted exchange flow of about 1000 m$^3$/s during the summer. Our results show much larger exchange flows, with mean inflow of 2000 m$^3$/s and episodic events greater than 4000 m$^3$/s. A series of idealized wind forcing experiments testing the influence of wind direction revealed that southward (northward) winds enhance (inhibit) exchange flow at the mouth of the bay. A simulated dye tracked cold, high-nutrient shelfwater and showed maximum dye intrusions under eastward and southeastward winds. Eastward wind forcing drives shelf upwelling, which increases the concentration of dye in the inflow. Our results predict significant temporal variability in the nutrient concentration of the subtidal inflow, with eastward winds driving the highest concentrations. We updated the bay’s nitrogen budget to account for our new calculations of exchange flow and the recent wastewater treatment upgrades. The new estimate places offshore nitrogen contributions at 51% of the bay’s total, compared to 17% in previous estimates. Future work will consider the role of wind-driven shelf intrusions in triggering phytoplankton blooms in the bay.
2.1 Introduction

Estuaries have been referred to as “mixing machines” which combine fresh river water with salty ocean water, giving rise to horizontal density gradients (T. Wang et al., 2017). Gravity acts to flatten the isopycnals, and the resulting pressure gradients drive a bi-directional flow which pulls in ocean water at depth and pushes out a brackish river-ocean mixture near the surface (Geyer & MacCready, 2014). This buoyancy-forced tidally-averaged overturning is referred to as the “exchange flow”, “gravitational circulation”, or “estuarine circulation” and controls the “residence time” and “flushing rate”. The fundamental salt balance of estuarine exchange flow was first described by Knudsen (1900) and then the steady-state equations of motion were developed by Pritchard (1956) and expanded by Hansen & Rattray (1965) and Chatwin (1976). In terms of volume transport, the exchange flow is typically an order of magnitude smaller than the oscillatory tidal flow but an order of magnitude larger than the river input. The exchange flow has a dominant role in the transport of sediment, plankton, pollutants, and nutrients.

One fundamental benefit of understanding estuarine exchange flow is quantifying the importance of nutrient inputs from offshore waters. Coastal and estuarine waters within many parts of the world are impacted by eutrophication, stemming from anthropogenic nitrogen inputs (Bricker et al., 2008; Deacutis, 2008; Melrose et al., 2007). For decades, environmental managers have struggled to quantify the different sources in the estuarine nitrogen budget, and offshore sources are typically the least constrained input in the system. The physics of shelf-estuary intrusions has been studied in major east coast US estuaries including Chesapeake Bay (Valle-Levinson
and Lwiza, 1995; Valle-Levinson et al., 1998; Valle-Levinson et al., 2001; Wong and
Valle-Levinson, 2002; Scully, 2010; 2013), Delaware Bay (Wong and Munchow,
1995), Tampa Bay (Meyers et al., 2007; Weisberg & Zheng, 2006), and Long Island
Sound (Whitney et al., 2016; Gay et al., 2004; O’Donnell et al, 2014). West coast
estuaries such as Puget Sound (Sutherland et al., 2011), San Diego Bay (Delgadillo-
Hinojosa et al., 2008), Willapa Bay (Roegner et al., 2002) and Yaquina Bay (Brown
and Ozretich, 2009) are strongly influenced by nutrient inputs from offshore. In Puget
Sound, offshore nitrogen makes up over 90% of total inputs (Mackas & Harrison,
1997).

Narragansett Bay represents an excellent setting for studying estuarine-shelf
exchange. Years of survey and time series records exist for numerous physical and
biogeochemical parameters. A fixed site buoy network has been collecting
temperature, salinity, chlorophyll, and dissolved oxygen data since 2003 (Codiga et
al., 2009), and years of underway dissolved oxygen surveys have revealed key details
of the spatial-temporal patterns in hypoxia and anoxia throughout the estuary (Prell et
al., 2006). Narragansett Bay is also home to one of the world’s longest-running
plankton surveys (Smayda & Borkman, 2008). Hypoxic conditions occur sporadically
throughout the summer, usually following phytoplankton blooms during neap tides
when stratification is at a maximum (Bergondo et al., 2005). Summertime
phytoplankton blooms are intense and sporadic, unlike the long winter-spring bloom
(Furnas et al. 1976). Despite close monitoring of anthropogenic nutrient sources, these
hypoxia-causing summertime blooms cannot be explained by the existing nitrogen
budget alone (Chaves, 2004).
Narragansett Bay nutrient reduction measures beginning in 2005 have been successful in implementing tertiary treatment at the wastewater treatment facilities (WWTFs) feeding the bay. Nitrogen inputs from wastewater are down 60% (Oviatt et al., 2017). Anthropogenic nutrients from rivers and wastewater treatment facilities discharge into the upper bay, particularly into the narrow Providence River sub-estuary. Nearly every biological and chemical concentration varies along a north-south gradient from the anthropogenically-influenced upper bay to the ocean-influenced mouth (Oczkowski et al., 2008; Oviatt et al., 2002). In recent surveys, these north-south gradients have flattened as a result of reduced inputs in the upper bay and no change in offshore inputs (Oviatt et al., 2017).

Nitrogen inputs from offshore remain the least constrained component of the bay’s nutrient budget. The product of the volumetric inflow and its nitrogen concentration provides the nitrogen mass flux. The mass flux uncertainty is the result of both (1) poor understanding of the amount of ocean water entering the bay and (2) lack of data on the nutrient concentrations of this inflow. Previous attempts to estimate the strength of the Narragansett Bay exchange flow can be separated into two different approaches: hydrographic and hydrodynamic.

The hydrographic approach does not involve any consideration of velocities but instead uses the salinity of the estuarine outflow and the magnitude of river inputs to estimate the exchange flow. If 100 m$^3$/s of river water enters the estuary, then 100 m$^3$/s of river water must exit, but as a result of mixing within the estuary, the outflow is a mix of both ocean water and river water. If we determine from salinity measurements that the ratio is 10 parts ocean water and 1 part river water, then the volumetric
outflow ($Q_{out}$) must be 1100 m$^3$/s. This also tells us that the inflow ($Q_{in}$) must be 1000 m$^3$/s in order to balance the net depth-integrated outflow of 100 m$^3$/s. The advantage of this approach is that salinity and river flow data are relatively easy to measure. The disadvantage is that it assumes a steady state system and cannot capture variability due to other effects such as wind. Hydrographic salt-balance budgets of Narragansett Bay estimate the exchange flow at 900-2500 m$^3$/s, varying as a function of river discharge (Officer & Kester, 1991; Pilson, 1985). Weisberg & Sturges (1976) noted that substantial exchange between the East and West Passages complicated this simple salt budget approach.

The hydrodynamic approach uses velocities to calculate the exchange flow. In Narragansett Bay, this requires calculating transport through continuous sections across the East and West Passages. Significant cross-channel variability in the along-channel velocity makes it difficult to produce a complete estimate. Kincaid et al. (2003) is the only study to use underway ADCP surveys in this region. Their transects just outside the mouth of Narragansett Bay revealed inflow and outflow jets with complex spatial structure, but likely did not provide a reliable estimate of the exchange flow magnitude. The transects’ spatial orientation missed substantial sections of the inflow and resulted in net depth-integrated outflows of 1200-1700 m$^3$/s which were not in balance with the river inputs of 50-300 m$^3$/s. Weisberg & Sturges (1976) estimated transports through the West Passage based on velocity measurements at a single mooring station by assuming a laterally-homogeneous structure across the channel. They estimated West Passage mean $Q_{in}$ and $Q_{out}$ at 110 m$^3$/s and 150 m$^3$/s respectively, but a strong southwestward wind event produced $Q_{out}$ of 1160 m$^3$/s and
then after winds relaxed, led to a maximum Qin of 660 m$^3$/s. This was the first study to describe the dramatic role of winds in modifying estuarine non-tidal circulation. (A.S. Pfeiffer-Herbert et al., 2015) estimated Qin through the East Passage from a single ADCP and, like Weisberg & Sturges (1976), extrapolated a laterally-constant along-channel velocity, though they limited the inflow region to 0.8 km wide (one third of the channel). This method resulted in a mean East Passage Qin of 900 m$^3$/s (and maximum Qin of 2300 m$^3$/s during the largest intrusion event.

We present new results aimed at quantifying estuary-shelf exchange physics for the Narragansett Bay – Rhode Island Sound system. We focus specifically on utilizing a combination of numerical model simulations and observations to address questions related to offshore intrusions of key biogeochemical parameters. Previous numerical models of this region have tended to focus either on currents within the Bay or on the shelf, but have not been appropriate for characterizing the exchange between the two (Bergondo, 2004; R. B. Gordon & Spaulding, 1987; Hess, 1976; Liu et al., 2016; Rogers, 2008). Results presented here utilize a model which resolves the estuary-shelf interface, combined with recent advances in moored ADCP data at the mouth.

During the summer-fall period of 2018, three moored ADCPs were deployed across the East Passage, just inside the mouth of the bay. A fourth ADCP was deployed in the lower West Passage, proximal to the location of earlier moored observations (Weisberg & Sturges, 1976). The use of three ADCPs within the main East Passage inflow channel reveals that the core of intruding water frequently oscillates between eastern and western sides of the passage, most notably with the
spring-neap cycle, but also with wind patterns. These records allow us to make the most accurate estimates of exchange flow at the Narragansett Bay interface to date.
2.2 Methods

2.2.1 Observational data sources

Four Acoustic Doppler Current Profilers (ADCPs) were moored across the estuary-shelf interface in late summer 2018 (Figure 2-1; Kincaid et al., 2020). This field campaign represents the best available observational timeseries of exchange between Narragansett Bay and Rhode Island Sound (RIS). One ADCP site was located in the West Passage (WP1) and the other three ADCPs were placed across the mouth of the East Passage. East Passage sites were spaced 500 m from each in order to capture the cross-channel variability in the along-channel flow. The ADCPs were deployed August 7th through November 9, 2018. At all the sites, several meters of near-surface data were discarded due to low data quality. Locations, bottom depths, and vertical resolution are presented in Table 1. Sea surface height (SSH) observations were taken from NOAA’s Newport, RI tide gauge, located on the eastern shore of the East Passage about 5 km up-bay from the ADCPs.
Figure 2-1. Bathymetry map of lower Narragansett Bay (a) showing that the bay connects to the shelf (Rhode Island Sound) via three channels, called the West Passage, the East Passage, and the Sakonnet River. Of these, the East Passage is the deepest and responsible for most of the inflow. West Passage is responsible for second most exchange flow. A zoomed-in view of the mouth of the West and East Passages shows the locations of the 2018 ADCP data used in this study (b). The numerical model grid (c) covers all of Narragansett Bay and Rhode Island Sound, and extends approximately 100 km offshore of the mouth of the Bay in order to capture the shelf circulation.

Table 2-1. ADCP locations, depths, and vertical resolution. Angle describes the local along-channel angle in degrees clockwise of north.

<table>
<thead>
<tr>
<th></th>
<th>Latitude</th>
<th>Longitude</th>
<th>Depth (m)</th>
<th>Bin size (m)</th>
<th>Angle (ADCP)</th>
<th>Angle (ROMS)</th>
</tr>
</thead>
<tbody>
<tr>
<td>EP1</td>
<td>41.470</td>
<td>-71.370</td>
<td>34.0</td>
<td>1.0</td>
<td>58.08°</td>
<td>59.69°</td>
</tr>
<tr>
<td>EP2</td>
<td>41.466</td>
<td>-71.369</td>
<td>32.5</td>
<td>1.0</td>
<td>51.47°</td>
<td>48.85°</td>
</tr>
<tr>
<td>EP3</td>
<td>41.463</td>
<td>-71.365</td>
<td>49.9</td>
<td>2.0</td>
<td>34.68°</td>
<td>34.46°</td>
</tr>
<tr>
<td>WP1</td>
<td>41.469</td>
<td>-71.403</td>
<td>22.3</td>
<td>1.0</td>
<td>3.34°</td>
<td>3.52°</td>
</tr>
</tbody>
</table>
2.2.2 Model description

The numerical model uses the Regional Ocean Modeling System (ROMS). The grid extends 100 km south of the mouth of Narragansett Bay in order to resolve shelf circulation outside of the bay. Grid cell size varies from a maximum of 390 m at the southern boundary of the grid to 55 m in the Providence River at the head of Narragansett Bay. For vertical resolution, we used 15 terrain-following sigma coordinates with higher resolution in the near-surface layers.

The ocean model is forced at the surface, open boundaries, and rivers with data from both observational sources and model output. Open boundary forcing is separated into tidal and subtidal components. Tidally, the ROMS model is forced with depth-integrated velocities and SSH from a shallow water tidal model of the North Atlantic (Szpilka et al., 2016). Subtidal temperature, salinity, and subtidal velocities are taken from the FVCOM GOM3 model (Chen et al., 2011) and interpolated to the ROMS boundary. Atmospheric fields are obtained from the North American Mesoscale (NAM) atmospheric model and applied using the bulk flux formulation of Fairall et al. (1996). Spatially-varying winds are applied on NAM’s 12 km grid but all other fields are applied uniformly based on a NAM cell located in Rhode Island Sound (41°N 71°W). Finally, river discharge data were obtained from USGS river gauge observations. River salinity was set to 0 and river temperature was estimated based on a moving mean of air temperature.

Forcing files are developed for two styles of numerical model simulations: realistic runs and process runs (Table 2-2). A full-year 2018 realistic simulation using realistic environmental forcing was run to allow for quantitative data-model
comparisons with the 2018 ADCP records. A series of shorter process-oriented simulations was also completed to quantify how different forcing parameters relate to various estuary-shelf exchange metric. These experiments explored the effects of rotation, stratification, tidal amplitude, and, most importantly, wind forcing scenarios.

A primary goal is to better understand and predict shelf intrusions based on forcing conditions. We are able to track characteristics for how RIS water moves into Narragansett Bay using a modeled dye added to water residing outside of the bay. The dye in our ROMS simulations is a simplified proxy for the deep pool of high-concentration dissolved inorganic nitrogen that has been documented in RIS. We choose to start with a physics-only model to understand the spatial-temporal extent of shelf water masses for different intrusion scenarios. Therefore, the dye is treated as a conservative tracer within the models and is not consumed by biology.

Dye is initialized in the subsurface water for all model cells outside of Narragansett Bay. The dye is initialized with a depth-dependent concentration profile that is constant in the horizontal. Starting concentrations are constant at 1000 kg/m$^3$ for depths greater than 30 meters. Concentrations are 0 kg/m$^3$ from 0-5m depth and increase linearly to a value of 1000 kg/m$^3$ at 30 meters depth.

Table 2-2. List of model runs.

<table>
<thead>
<tr>
<th>Run code</th>
<th>Wind forcing</th>
<th>Coriolis?</th>
</tr>
</thead>
<tbody>
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<td>Yes</td>
</tr>
<tr>
<td>osom_101</td>
<td>no wind</td>
<td>Yes</td>
</tr>
<tr>
<td>osom_102</td>
<td>windspeed doubled</td>
<td>Yes</td>
</tr>
<tr>
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<td>winds reversed</td>
<td>Yes</td>
</tr>
<tr>
<td>osom_210</td>
<td>E-ward 5 m/s</td>
<td>Yes</td>
</tr>
<tr>
<td>osom_211</td>
<td>NE-ward 5 m/s</td>
<td>Yes</td>
</tr>
<tr>
<td>osom_212</td>
<td>N-ward 5 m/s</td>
<td>Yes</td>
</tr>
<tr>
<td>osom_213</td>
<td>NW-ward 5 m/s</td>
<td>Yes</td>
</tr>
<tr>
<td>osom_214</td>
<td>W-ward 5 m/s</td>
<td>Yes</td>
</tr>
<tr>
<td>osom_215</td>
<td>SW-ward 5 m/s</td>
<td>Yes</td>
</tr>
<tr>
<td>osom_216</td>
<td>S-ward 5 m/s</td>
<td>Yes</td>
</tr>
<tr>
<td>osom_217</td>
<td>SE-ward 5 m/s</td>
<td>Yes</td>
</tr>
<tr>
<td>osom_248</td>
<td>zero wind</td>
<td>No</td>
</tr>
<tr>
<td>osom_240</td>
<td>E-ward 5 m/s</td>
<td>No</td>
</tr>
<tr>
<td>osom_241</td>
<td>NE-ward 5 m/s</td>
<td>No</td>
</tr>
<tr>
<td>osom_242</td>
<td>N-ward 5 m/s</td>
<td>No</td>
</tr>
<tr>
<td>osom_243</td>
<td>NW-ward 5 m/s</td>
<td>No</td>
</tr>
<tr>
<td>osom_244</td>
<td>W-ward 5 m/s</td>
<td>No</td>
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<tr>
<td>osom_245</td>
<td>SW-ward 5 m/s</td>
<td>No</td>
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<tr>
<td>osom_246</td>
<td>S-ward 5 m/s</td>
<td>No</td>
</tr>
<tr>
<td>osom_247</td>
<td>SE-ward 5 m/s</td>
<td>No</td>
</tr>
</tbody>
</table>

2.2.3 Analysis methods

Flow in the bay is heavily steered by bathymetry so to facilitate comparisons between observed and modeled currents we rotate horizontal velocities into along-channel coordinates, as determined by the direction of maximum variance. Positive along-channel velocities indicate up-bay, or intruding flow. Velocity information is
similarly projected into an across-channel vector that is locally perpendicular to the along-channel direction. Subtidal velocities, or residual velocities, are essential for understanding longer term transport and exchange patterns in coastal waters. Subtidal velocities are the focus of much of our analysis and are calculated using a 33 hour lowpass Butterworth filter.

Since we are interested in deep intrusions of shelf water, and because velocity observations of the upper-water column are limited, we focus on the lower layer for our velocity comparisons. The depth of the lower layer is calculated as the depth where the mean along-channel velocity switches from outflow to inflow. See Figure 2-4 for plots of these mean velocity profiles in which the zero velocity depth can be seen. Model skill was quantified using the skill score defined by (Willmott, 1981) and referred to subsequently as the “Willmott skill score.”

An essential step in this work is to use the enhanced spatial scale of the model to estimate the exchange flow through the mouth of Narragansett Bay. For Narragansett Bay, this involves calculating the subtidal volume transport through cross-channel sections in the lower East and West Passages. To this end, we compute up-bay (Qin) and down-bay (Qout) transports. Transport through each grid cell in the cross-section is calculated by multiplying the section-normal subtidal velocity by the area of the grid cell face. Grid cell face areas vary in time as the vertical layers stretch with the rising and falling tides. At each timestep, Qin is computed as the sum of transports through cells with up-bay flow, and Qout is the sum of down-bay transport.
2.3 Model evaluation

Models have the capacity to be powerful tools for scientific inquiry and management within coastal ecosystems, however it is important to recognize model limitations. All models represent solutions derived from approximations of the full set of hydrodynamic equations, plus they are only as reliable as the forcing conditions we prescribe. It is essential, therefore, to develop quantitative comparisons between modeled and observed data to determine model skill. While it is common to develop data-model comparisons for tidal constituents, it is essential for our purposes to also develop comparisons of subtidal parameters. We compare the ROMS model output against observed sea surface height (SSH), along- and across-channel velocities, and bottom temperature. Quantifying the model skill in the 2018 hindcast will help us determine our confidence in the results when we configure the model with hypothetical forcing scenarios which cannot be validated by observations.

2.3.1 Sea Level

SSH comparisons (Figure 2-2) show that the model agrees very well with observed tidal timing and amplitude, including the spring-neap cycle, but agrees less well with the lowpass filtered SSH. The unfiltered SSH has a Willmott Skill of 0.98 and the lowpass filtered SSH has a Willmott Skill of 0.79. During events with extreme subtidal SSH anomalies in October and November, the model tends to under-predict the magnitude of these anomalies, both positive and negative.
2.3.2 Currents

The angle of the along-channel coordinates, as defined by the direction of maximum variance in velocities, varied greatly between the four ADCP sites (Table 2-1). This was expected based on the differences in the local bathymetry gradient direction at the four sites. Agreement between the observations and the modeled angle was very close -- within 2.62° for all sites.

Comparisons of along-channel instantaneous and subtidal velocities are shown in Figure 2-3. Additionally, skill scores for the along- and across-channel velocities can be found in Table 2-3. In general, the along-channel velocities had higher skill than across-channel velocities, and the instantaneous velocities had higher skill than subtidal velocities. The main focus of analysis is the along-channel velocities and particularly the subtidal along-channel velocities since these will be responsible for the residual exchange flow. Along-channel subtidal skills for the three East Passage ADCPs (from east to west) were 0.89, 0.83, and 0.83 and the West Passage ADCP was 0.72.
Figure 2-3. Instantaneous (left column) and lowpass filtered (right column) deep along-channel velocities at the four ADCP stations. Comparing ADCP observations (black lines) and ROMS model (blue lines).

Table 2-3. Willmott skill scores, ROMS model vs. observations.

<table>
<thead>
<tr>
<th>Site</th>
<th>Field</th>
<th>Filtering</th>
<th>Skill</th>
</tr>
</thead>
<tbody>
<tr>
<td>Newport NOAA</td>
<td>SSH</td>
<td>Instantaneous</td>
<td>0.98</td>
</tr>
<tr>
<td>Newport NOAA</td>
<td>SSH</td>
<td>33hr Lowpass</td>
<td>0.79</td>
</tr>
<tr>
<td>EP1</td>
<td>along-channel velocity</td>
<td>Instantaneous</td>
<td>0.93</td>
</tr>
</tbody>
</table>
All four sites demonstrated two-layer estuarine mean flow of deep inflow and surface outflow. The time-mean vertical structure of the along-channel velocities are shown in Figure 2-4. Model disagreements in bottom depth at the ADCP locations are due to errors in the bathymetry data used to construct the model as well as additional bathymetry smoothing in the model for numerical stability. The zero-velocity depth where the mean flow switches from inflow to outflow agrees well between the model
and data. EP3 had the largest difference with the modeled zero-depth nearly 5 m higher than ADCP observations but note that the model also had 5 m shallower bathymetry.

Figure 2-4. Vertical profiles of mean along-channel velocity from ADCP observations (black line) and ROMS model (blue line) for the four ADCP stations: WP1 (a), EP1 (b), EP2 (c), and EP3 (d). Positive velocity indicates flow into the Bay. The black circles indicate the height of the ADCP bins; the blue circles indicate the mean height of the ROMS vertical layers. An additional data point was added at seafloor with velocity 0. Observations and model output at all four sites show estuarine inflow in the lower layer and outflow in the upper layer.
Figure 2-5. Additional velocity profiles at historical ADCP sites published by Kincaid et al. (2008). Relative to the 2018 ADCPs, these sites are located 3 km up the West Passage (a) and 5 km up the East Passage (b). Plots show mean along-channel velocities for summer 2000 ADCP observations (black line) and summer 2018 ROMS model (blue line).

The vertical structure of the velocity at all four sites shows variability in time. Figure 2-6 shows a timeseries of the depth-varying along-channel subtidal velocity. There are times when EP1 and WP1 are full-watercolumn inflow and times when EP2 and EP3 are briefly full-watercolumn outflow. WP1 has the least variability in the thickness of the two layers and the model and observations both agree that this zero-depth occurs between 5 and 10 meters below the surface. In the East Passage, the thickness of the inflow and outflow layers was much more variable than the West Passage. EP1 had days long periods of strong up-bay velocities which extended to the surface. These periods occurred during spring tides. Simultaneously at EP2 and EP3, these spring tides were correlated with weaker inflow velocities and a deepening of the upper outflow layer. During neap tides, EP2 demonstrated large variability with the inflow layer extending up to 5 meters below the surface or shrinking to 20 meters below the surface.
Figure 2-6. Full-watercolumn subtidal along-channel velocities from ADCP observations (left column) and ROMS model (right column).

Combining the velocity data at the three ADCPs across the East Passage, we can create a sparse picture of the cross-channel structure and compare that to the results from the ROMS model (Figure 2-7). We’ve identified two main modes that occur during low versus high inflow periods, as identified from model Qin. During low inflow periods, the inflow jet shifts to the western side of the channel (EP1 side) where it extends throughout much of the watercolumn. Meanwhile the outflow jet extends deeper into the watercolumn at EP2 in the middle of the channel. During high inflow periods, deep inflow is enhanced at EP2 and EP3 and the outflow jet is constrained to a thin surface layer with high velocities.
Figure 2-7. East Passage cross-channel structure of subtidal velocities from ADCP observations (top row) and ROMS model (bottom row). The four columns highlight four different times: two low Qin (columns 1 and 2) and two high Qin times (columns 3 and 4). There emerged two main regimes that the flow would shift between. During low exchange periods, the outflow layer extended deep into the water column, while during high exchange periods, there were stronger outflow velocities which were confined to a shallow surface layer. The western-most site EP1 behaved significantly differently from EP2 and EP3, with EP1 always flowing into the Bay (this can also be seen in Figure 2-6).

2.3.3 Temperature

Comparing bottom temperature at the four sites shows that the model captures many features of the timeseries (Figure 2-8). The East Passage sites lie along the data trend and the West Passage site is biased 1-2°C warmer than observations. We
recognize two regime changes in the data: first in late September when high frequency variability is reduced, and then in mid-October when there is a rapid cooling. The model represents both these seasonal shifts with impressive accuracy. However, short duration summertime intrusion events are less well-represented (Figure 1-9).

Figure 2-8. Bottom temperature comparisons from ADCP observations (black line) and ROMS model (blue line) for the four ADCP stations: WP1 (a), EP1 (b), EP2 (c), and EP3 (d).

In the East Passage, bottom temperature is a critical proxy for detecting the presence of shelfwater intrusions entering the bay. East Passage temperature records were similar between the three sites, therefore we’ll consider the deepest site, EP3, for our closer look at model bottom temperature performance (Figure 2-9). We identified two periods of significant cooling across multiple tidal periods, and both these cooling events were correlated with enhanced intrusion velocities at EP2 and EP3. The first instance of rapid cooling in this observational record began on Aug-15 when the
temperature fell by 6°C over 3 days. Over the same period, the ROMS bottom temperature fell by only 3°C. Comparing with subtidal velocities, the ROMS model underestimated intrusion velocities at both EP2 and EP3. The second event beginning on Aug-23 showed better agreement: both the model and observations dropped about 4°C over 3 days. The ROMS velocities were still weaker than observations but were stronger than the first cooling event. We conclude that the model captures the mean state and monthly variability but misses some of the more rapid event-driven intrusions.
Figure 2-9. Bottom temperature and subtidal velocities, showing two cold-water intrusion events. The full three month observed bottom temperature timeseries for site EP3 is shown in (a). The blue box indicates the August 12-29, 2018 time period shown in (b-d). Observed (black) and modeled (blue) bottom temperature for the shorter time period is shown in (b) and the timing of the observed temperature drops are highlighted with yellow ovals and a light red shading. Plots (c) and (d) show lowpass-filtered deep along-channel velocity at station EP3 (c) and EP2 (d). The red-shaded intrusion periods
correspond to periods of cooling and enhanced inflow in the observations, which we recognize as deep intrusion events.
2.4 Results

2.4.1 Exchange flow in realistic model

Our comparisons of the model against ADCP data were critical for establishing the skill of the modeled velocities, but those sparse ADCP locations could not give us full volume transport calculations. By integrating model subtidal velocities across sections of the channels, we obtain the exchange flow in m³/s into (Qin) and out of (Qout) the East Passage and West Passage. Our focus will only be on the bi-directional Qin and Qout (instead of the depth-integrated transport) since this is more important for advective material transport between the estuary and the ocean, and has not been sufficiently explored in the literature. The model also allows us to extend our time period to earlier in the summer before the August ADCP deployment.

Histograms of Qin and Qout strength for the East Passage (Figure 2-10) and West Passage (Figure 2-11) show larger mean transports and much greater variability compared to previous Narragansett Bay estimates (Officer & Kester, 1991; Pilson, 1985). The East Passage is the main conduit for inflow, with Qin more than 5x larger than through the West Passage. East Passage Qout is 2-3x stronger than West Passage Qout.
Figure 2-10. Histograms of East Passage Qin (top row) and Qout (bottom row). Horizontal axis shows subtidal transport in m³/s. Data was split into neap tide periods (left column) and spring tide periods (right column).
Figure 2-11. Histograms of West Passage Qin (top row) and Qout (bottom row). Horizontal axis shows subtidal transport in m$^3$/s. Data was split into neap tide periods (left column) and spring tide periods (right column).

There is a strong fortnightly tidal signal showing enhanced exchange during spring tides when there is larger tidal mixing energy (Figure 2-12). The summertime is marked by low river discharge, never exceeding 100 m$^3$/s until September. The Officer & Kester (1991) estimate predicts that these river discharge levels would produce transports under 1500 m$^3$/s however we see transports fluctuating between 2000 m$^3$/s and 3000 m$^3$/s, correlated with the neap-spring cycle.
Figure 2-12. Results from realistic model. (a) Timeseries tidal height (thin grey line) and amplitude of semidiurnal tide (thick grey line). Qin (b) for realistic simulation (thin blue line) with 5-day moving mean (thick blue line). River discharge (c).

In addition to the tidal signal, there is higher-frequency variability driven by the winds. Figure 2-13 shows a 12-day neap tide period in which there is a brief event on August 19 of Qin more than 3000 m$^3$/s through the East Passage alone. This enhanced inflow corresponds to a period of southward winds. A no-wind simulation during this
period produces an inflow approximately half as strong, 1500 m$^3$/s. Reversing the winds to northward produces Qin that is half as strong as the no-wind case. If the wind were acting only as a source of turbulent mixing, we would expect similar results for the realistic and the reversed wind cases since the windspeed magnitudes are the same in both. Instead, the realistic wind and reversed wind transports are negatively correlated, showing that the direction of the wind plays a key role in modifying the exchange flow. In the next section, we analyze the response to different wind directions via a set of simulations forced with idealized winds.

Figure 2-13. Wind vectors for realistic winds (a). East Passage Qin (b) comparing realistic wind forcing (blue line), no winds (grey line), and reversed winds (red line).
2.4.2 Idealized forcing results

Eight idealized model simulations were forced with constant 5 m/s wind from eight different directions (every 45°) as well as a zero-wind case. The mean Qin for the first three days of the simulations are summarized in Figure 2-14a. As expected, the down-bay (southward) winds enhance the estuarine exchange flow and up-bay (northward) winds inhibit this flow. The three wind directions with the largest Qin were southward, south-eastward, and south-westward. The results were approximately east-west symmetric. These constant-wind trends agree with our theoretical predictions for a north-south oriented estuary.

![Figure 2-14. Mean Qin (a) and cumulative dye transport (b) through East Passage [with Coriolis force]. Model winds were spatially and temporally constant at 5 m/s except the zero wind case. Volume and dye transports computed over the first three days of wind forcing.](image URL)
Next, a simulated dye was added on the continental shelf in order to track intrusions of high-nutrient shelfwater. Maps of depth-averaged dye concentrations for each scenario are shown in Figure 2-15. Surprisingly, volume transport was not a predictor of dye transport (Figure 2-14). The dye transport trends have a more eastward-wind preference than Qin. The top three wind directions with largest dye transport are southeastward, eastward, and southward. This agrees with previous observations stating that shelf intrusions into the East Passage are driven by SE-ward winds (Kincaid et al., 2008). If we compare SW-ward versus E-ward winds, a volume transport-only view of the exchange flow would have predicted that the SW-ward scenario delivers more dye. However, despite larger Qin for the SW-ward winds, the E-ward winds deliver more than twice as much dye.
Figure 2-15. Depth-averaged dye concentrations for the 9 wind scenarios, 14 days after initial dye release. The white arrow indicates the wind direction for each simulation.
Figure 2-16. Wind-driven sea surface height (SSH) anomaly for simulations with Coriolis force. The no wind case reference case (e) shows the mean SSH over 6 tidal cycles and the other frames show the wind direction runs with mean SSH minus the reference case. Contours drawn every 1 mm, except the reference case where they are drawn every 2 mm. The white arrows indicate the wind direction.
2.5 Discussion

2.5.1 Spatially-varying role of Coriolis force

Rotational effects in Narragansett Bay versus RI Sound are the key to explaining why exchange flow strength was not a direct predictor of dye intrusions (Figure 2-14). Flow in the bay is directed along narrow channels and experiences limited rotational effects. The first baroclinic Rossby radius in the lower bay is between 2.5 km in the winter and 4 km in the summer, which is larger than the 1-3 km channel widths in the lower bay (Kincaid et al., 2003). In the absence of geostrophic balance, a simple southward wind sets up a north-south gradient in SSH that drives maximum deep inflow surface outflow.

The concentration of dye in the intrusion is largely controlled by processes outside the bay on the shelf, which is not constrained by the same narrow channel geometry. The isobaths run approximately east-west and therefore, in the presence of rotation, eastward winds are the optimal wind direction for upwelling. Offshore Ekman transport in the upper layer drives onshore transport in the lower layer. This lower layer is characterized by colder water, higher nutrient concentrations, and, in the model, high dye concentration. Based on the model results, the higher dye concentrations in the inflow are correlated with more eastward-directed winds. The mouth of the bay represents the interface between an irrotational channel-oriented flow regime and a rotational shelf regime (Dyer, 1997; Pfeiffer-Herbert, 2012).

To test this hypothesis that shelf upwelling affects the concentration of dye in the inflow, we turned off the Coriolis force and re-ran the idealized wind simulations.
(Figure 2-17). These irrotational cases had a dye transport pattern that more closely correlated with the exchange flux results. This is because the optimal Qin winds were now aligned with the optimal shelf upwelling winds.

*Figure 2-17. Mean volume transport (a) and cumulative dye transport (b) through the East Passage with Coriolis force turned off. Model winds were spatially and temporally constant at 5 m/s (except the no-wind case). Volume and dye transports are computed over the first three days of wind forcing. The no-Coriolis volume transport trends are similar to the Coriolis results from Figure 2-14, confirming that the circulation within Narragansett Bay is minimally affected by rotation. The dye transport however is very different from the Coriolis case. In the Coriolis model, the shelf was subject to geostrophy, so eastward winds generated upwelling conditions (onshore transport of deep dye). Now with no Coriolis force, southward winds are both the direction of maximum shelf upwelling and maximum estuarine circulation, and this combination creates the ideal conditions for large dye intrusions.*
The implication of this finding is that winds modify not only on the strength of the bay’s exchange flow, but also the source of the inflow waters. This can explain why some intrusion events cause significant cooling while other relatively strong inflows do not. Ideal intrusion winds are towards the southeast since there is the southward component to enhance the bay’s two-layer exchange which pulls in shelf water, plus the eastward component of the wind to drive shelf upwelling and onshore advection of deep water. This is a rare wind direction in this region during the summer -- winds are predominantly towards the northeast or the southwest, with the exception of storms such as tropical cyclones. In the winter, however, this is a fairly common wind direction. Further investigation is required into how this estuary-shelf exchange behaves under well-mixed winter conditions, but we note the possible connection between winter intrusion-favorable winds and the bay’s large winter-spring phytoplankton bloom. We also ask how this might be affected by changing wind conditions under future climate scenarios.
Figure 2-18. Wind-driven sea surface height (SSH) anomaly for simulations without Coriolis force. As in Figure 2-16, the no wind case reference case (c) shows the mean SSH over 6 tidal cycles and the other frames show the wind direction runs with mean SSH minus the reference case. Contours drawn every 1 mm. The white arrows indicate the wind direction.
Figure 2-19. Schematic summarizing the spatially-varying role of rotation between the channel-dominated Narragansett Bay and the Ekman-driven continental shelf. Southward wind drive maximum exchange flow in the bay but eastward winds drive optimal upwelling on the shelf.

2.5.2 East Passage vs. West Passage exchange flow

Previous work has described Narragansett Bay depth-integrated transport as forming a cyclonic gyre around Conanicut Island characterized by East Passage inflow and West Passage outflow. Previous studies have observed that this gyre is “spun up/down” in response to different wind directions (Kincaid et al., 2008; Rogers, 2008). Our results further expand on previous work by describing the gyre transport within the context of bi-directional transport in both passages. Figure 2-20 shows a
timeseries of Qin and Qout for both the East Passage and West Passage (WP). The difference between inflow and outflow produces the depth-integrated transport.

Spinning up the gyre (increasing depth-integrated transports) occurs via increased EP Qin and simultaneously increased WP Qout. The opposing flows in each passage (EP Qout and WP Qin) are relatively stable and do not increase to compensate, therefore a net transport develops in each passage. This is different than what we would expect in a single-channel estuary. Conservation of volume requires Qin and Qout to approximately cancel, with the depth-integrated outflow equaling the riverine input, which is less than 100 m$^3$/s in our case. Weisberg & Sturges (1976) first noted that the net transport through the West Passage does not equal the riverine input.

Narragansett Bay’s two-channel geometry allows for relatively large depth-integrated transports with a net inflow through the EP and net outflow through WP.
Figure 2-20. Qin (solid line) and Qout (dashed line) through East Passage (a) and West Passage (b). The selected 12 day time period was chosen to highlight a period of enhanced East Passage Qin August 19-22 which corresponds to persistent SW-ward and S-ward winds (Figure 2-13). During this time, East Passage Qin is stronger than Qout resulting in a net depth-averaged inflow (indicated by the black circle and x) and West Passage Qout is stronger than Qin resulting in a net depth-averaged outflow (indicated by the black circle and dot).

This wind-driven control on EP Qin and WP Qout is confirmed when we compute correlations between transport and meridional winds (Table 2-4). The opposing flows in each passage (EP Qout and QP Qin) are much less influenced by the wind, as indicated by their smaller magnitude correlation coefficients. The depth-integrated transport (Qnet) shows high correlation with southward winds spinning up the cyclonic gyre. The West Passage’s shallow bathymetry does not support a strong Qin like the deeper East Passage. It’s less clear though why EP Qout does not have a strong wind dependence like WP Qout. Previous authors have noted some rotational steering of currents, with outflow favoring the western side of channels. We note that the modeled time period had frequent SW-ward winds but not SE-ward. Wind-driven surface outflow would favor the West Passage under SW-ward wind forcing.

Table 2-4. Correlation coefficients between the subtidal exchange flow and the northward component of wind velocity, for realistic model runs. Qin is always positive and Qout is always negative. Qnet is signed, with inflow (outflow) corresponding to positive (negative) Qnet values. Negative correlation for Qin does not mean that northward winds drive negative Qin, it means that northward winds decrease Qin.
Similarly, positive correlation for $Q_{out}$ indicates that northward winds generate outflow that is less negative, i.e. weaker.

<table>
<thead>
<tr>
<th>Correlation with Northward wind</th>
<th>$Q_{in}$</th>
<th>$Q_{out}$</th>
<th>$Q_{net}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>East Passage</td>
<td>-0.44</td>
<td>-0.22</td>
<td>-0.6</td>
</tr>
<tr>
<td>Northward (southward) wind inhibits (enhances)</td>
<td>+0.19</td>
<td>+0.6</td>
<td>+0.55</td>
</tr>
<tr>
<td>West Passage</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Finally, we note the potential role of tidal asymmetry in Narragansett Bay and suggest future investigation into this phenomenon. Asymmetry between flood and ebb tides can result in an additional source of subtidal exchange (Dronkers, 1986; Neill et al., 2014). This could be significant in Narragansett Bay, both on the scale of net inflow/outflow through the East/West Passage, but also for explaining some of the cross-channel structure. The lateral structure at the East Passage ADCP crossing shows a persistent inflow along the western side of the channel (Figure 2-7). We hypothesize that the curving channel at the mouth of the East Passage could be causing a flood-dominant bias on the western side of the channel. In support of this hypothesis, model results from further north up the East Passage where the channel is straighter do not show a western inflow jet (see Figure 1-19).

2.5.3 New estimate of offshore nitrogen inputs

Nixon et al. (1995) estimated that offshore dissolved inorganic nitrogen (DIN) sources make up about 15-20% of the bay’s total DIN input. Later Nixon et al. (2008) estimates did not include offshore contributions, and were based on WWTF nitrogen
loads from before tertiary treatment. We present an updated version of Nixon et al. (2008) which includes offshore sources, and we estimate them to be at least 50% of the total. We do this via changes to three key parameters: (1) mass of anthropogenic DIN inputs, (2) DIN concentration of inflow, and (3) magnitude of inflow volume transport.

Table 2-5. New estimate of Narragansett Bay nitrogen inputs. “Previous estimates” column based on 2003 data obtained from Nixon et al. (2008) and offshore estimate from Nixon et al. (1995).

<table>
<thead>
<tr>
<th>Nitrogen source</th>
<th>Previous estimates</th>
<th>Updated estimates</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Wastewater</strong></td>
<td>363</td>
<td>182 (50% reduction)</td>
</tr>
<tr>
<td><strong>Rivers</strong></td>
<td>145</td>
<td>145</td>
</tr>
<tr>
<td><strong>Atmospheric deposition and urban runoff</strong></td>
<td>67</td>
<td>67</td>
</tr>
<tr>
<td><strong>Offshore</strong></td>
<td>115</td>
<td>413</td>
</tr>
<tr>
<td>Qin = 1040 m$^3$/s$^a$</td>
<td>Conc. = 3.5 µmol/L$^b$</td>
<td>Qin = 2000 m$^3$/s$^c$</td>
</tr>
<tr>
<td><strong>TOTAL</strong></td>
<td>690</td>
<td>807</td>
</tr>
<tr>
<td><strong>Offshore % of total</strong></td>
<td>17%</td>
<td>51%</td>
</tr>
</tbody>
</table>

$^a$ (Officer & Kester, 1991)  
$^b$ (Kremer & Nixon, 1978)  
$^c$ Present study  
$^d$ (Chaves, 2004)
Managed nitrogen reductions began in 2005 and achieved a 60% decrease in nitrogen inputs by 2012 (Krumholz, 2012; Oviatt et al., 2017). As a conservative estimate, we applied a 50% reduction (Table 2-5). Additional reductions will further increase the offshore share of total nitrogen.

The DIN concentrations are the most uncertain parameter in the calculation. DIN concentration used by Nixon et al. (1995) came from Aug 1972-Aug 1973 surveys at two stations in the East Passage. Our updated DIN concentration of 6.55 µmol/L came from June 29-30, 2000 measurements of ammonium (5.38 µmol/L) and nitrite (1.17 µmol/L) in RI Sound (Chaves, 2004). This updated shelf concentration is likely higher than the actual inflow at the mouth, however the sampling of Kremer & Nixon (1978) found highest DIN concentrations November through February, so applying this summertime concentration as an annual mean could potentially be an underestimate. The upcoming publication by Kincaid et al. (in prep) will include greatly expanded nutrient measurements for the lower bay and RI Sound. For all categories except offshore, we use the total nitrogen (TN) values from Nixon et al. (2008). Our offshore concentration measurements only include DIN, but if we had concentrations of TN, the offshore mass flux would be higher.

Finally, we increased mean inflow to a conservative 2000 m³/s, which is lower than the mean inflow found in the present study. The original Nixon et al. (1995) estimates assumed an inflow of 1040 m³/s based on Officer & Kester (1991). That exchange flow was the result of a simple analytical box model based on salinity measurements, and did not consider any hydrodynamic observations or modeling.
Our estimates consider the relative magnitudes over annual means but the most significant implication of our results has been to highlight the degree of temporal variability. The model showed that both the inflow magnitude and nutrient concentration are sensitive to wind forcing. Additionally, we focused on the summer period when rainfall is minimal and therefore river and runoff inputs are lower than the annual mean shown in Table 2-5. Event-driven intrusions with 3,000+ m³/s transports following shelf upwelling could easily overwhelm all other nitrogen sources for multiple days. Future work should look for a connection between intrusion-favorable winds and summertime phytoplankton blooms. These intrusions provide a potential physical mechanism to explain blooms which are not accounted for in the present Narragansett Bay nutrient budget.
2.6 Conclusions

In this paper we presented hydrodynamic estimates of subtidal exchange flow through the mouth of Narragansett Bay. Previous estimates based on salt-balance predicted flushing rates of 800-2600 m$^3$/s, which varied only as a function of river flow (Officer & Kester, 1991; Pilson, 1985). Using a ROMS model validated by ADCP velocity data, we found that the mean exchange flow was approximately 2500 m$^3$/s, and much more variable than previously thought. Qin of 3000 m$^3$/s was common even when river discharge was very low (less than 50 m$^3$/s). We found that tidal amplitude and meridional winds were the dominant controls on exchange flow during the summer.

When comparing the West vs. East Passages, the model showed that 80-90% of the inflow enters via the East Passage. The two-channel geometry generates a net depth-integrated flow in through the East Passage and out through the West Passage. Results confirmed theoretical predictions, with southward winds enhancing exchange and northward winds inhibiting exchange. After removing wind events, fortnightly tides varied the exchange flow from about 2000 m$^3$/s at max neap tide to 3000 m$^3$/s at max spring tide.

We ran a series of idealized simulations to test the effects of different wind conditions and added a hypothetical “dye” to the shelfwater in the model as a proxy for offshore nutrients. We were surprised to find that intrusions of dye are not a simple function of the strength of the bay exchange flow, but that eastward wind directions with weaker inflow could actually input more shelfwater. Comparing runs with and without Coriolis force confirmed that rotational effects on the shelf led to optimal
upwelling under eastward winds, but Narragansett Bay exchange flow is not affected by rotation. This mismatch between optimal estuarine exchange winds and optimal shelf upwelling winds produces significant variation in the concentration of dye in the inflow. The practical significance is that when calculating offshore nutrient inputs, we cannot only measure volume transport and then apply constant nutrient concentration. Our model results predict that both the strength of the inflow and its nutrient concentration are highly variable, and that the wind dependence of the two is 90° out of phase. Future observational campaigns should include multiple cross-channel ADCPs to capture shifting cross-channel structure of the velocities, and add nutrient sampling to monitor the time variability inflowing chemical properties.
2.7 Appendix A: Wind observations

2.7.1 Wind direction climatology

Wind data at Newport, RI NOAA station from 2002 through June 2020 is compiled to show the predominant annual and summertime wind directions.

Figure 2-21. Wind histogram: all speeds, all months, no filtering.
Figure 2-22. Wind histogram: all speeds, summer only (June through September), no filtering.

![Wind histogram: all speeds, summer only (June through September), no filtering.](image)

Figure 2-23. Wind histogram: 5 m/s minimum speed, all months, no filtering.

![Wind histogram: 5 m/s minimum speed, all months, no filtering.](image)

Figure 2-24. Wind histogram: 5 m/s minimum speed, summer only, no filtering.

![Wind histogram: 5 m/s minimum speed, summer only, no filtering.](image)
Figure 2-25. Wind histogram: all speeds, all months, 3 day moving mean.

Figure 2-26. Wind histogram: all speeds, summer only, 3 day moving mean.
2.7.2 Wind model comparisons

North American Mesoscale model compared against wind observations at five stations in Narragansett Bay and the adjacent shelf.
Figure 2-29. Wind observations map showing the five stations in Narragansett Bay and the adjacent shelf.
Figure 2-30. Wind speed histograms comparing observations (black) vs. model (blue). Comparisons for period July 1 through October 1, 2018.
Figure 2-31. Wind direction histograms comparing observations (black) vs. model (blue). Comparisons for period July 1 through October 1, 2018.
2.8 Appendix B: CTD cast comparisons

Hydrographic data-model comparisons are possible thanks to a series of conductivity-temperature-depth (CTD) watercolumn profiles by Dr. Dave Ullman and Dr. Lucie Maranda. Data were collected at seven sites in Narragansett Bay and just outside the bay on ten days during 2018, from May 9 to October 26. For each profile, we compared observations against ROMS model output at the same time and location. The three parameters we considered are temperature, salinity and Brunt–Väisälä frequency ($N^2$). The Brunt–Väisälä frequency is defined as follows:

\[ N^2 = -\frac{g}{\rho} \frac{\partial \rho}{\partial z} \]
Figure 2-32. Map of the seven CTD cast sites.
Figure 2-33. Brenton Reef site. CTD observations (black) and ROMS model (blue) profiles at ten cast times. Temperature (top row), salinity (middle row), and Brunt–Väisälä frequency (bottom row).
Figure 2-34. Lower East Passage site (EPL).
Figure 2-35. Mid East Passage site (EPM).
Figure 2-36. Upper East Passage site (EPU).
Figure 2-37. Lower West Passage site (WPL).
Figure 2-38. Mid West Passage site (WPM).
Figure 2-39. Upper West Passage site (WPU).
Figure 2-40. Model temperature and salinity trends across all seven sites and nine sampling times.
MANUSCRIPT 3:

Boundary current impacts of coastal refinement in a variable-resolution global ocean model

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\textsuperscript{5}Center for Climate Systems Research, Columbia University, New York, NY, USA
Abstract

Variable resolution ocean models are a new tool which may improve the accuracy and computational efficiency of global climate models. The main advantage is the ability to increase resolution in important under-resolved regions, without the high costs of increasing resolution globally. We compare results from three different model meshes: (1) standard low-resolution eddy-closure mesh, (2) prohibitively expensive high-resolution eddy-permitting mesh, and (3) an experimental coastal-refined mesh with low-resolution globally plus 8 km resolution within 400 km from the North American coastline. Running the high-resolution model is 66 times slower than low-resolution, but the coastal-refined model is only 6.5 times slower than the low-resolution model. Our analysis focuses on two main processes: the California Current System upwelling on the Pacific’s eastern boundary and the Gulf Stream on the Atlantic’s western boundary. At the eastern boundary, we see that the coastal-refined model’s fine-scale upwelling patterns agree well with the high-resolution model. Coastal upwelling is the type of locally-forced process which very clearly benefits from regional mesh refinement. At the western boundary, Gulf Stream transport is too weak in the coastal-refined model, but our findings indicate that these results can be improved. The wind-driven equatorward transport in the North Atlantic subtropical gyre is actually similar between all three meshes, suggesting that low resolution over much of the basin is sufficient when designing variable-resolution meshes. The weak Gulf Stream is instead the result of poor deepwater formation in the North Atlantic, despite enhanced resolution in the Labrador Sea deepwater formation region. The coastal-refined mesh appears to artificially steer the Gulf Stream path too close to
land, which traps the Labrador Current, freshens the Labrador Sea, and inhibits deepwater formation. We recommend expanding the area of the refined region in order to avoid interactions between the mesh transition region and the Gulf Stream path.
3.1 Introduction

Ocean modeling in global climate models presents a different set of challenges than the regional modeling discussed in the previous chapters. The model domain covers the entire planet and the simulations span hundreds of years. The high computational costs of running ensembles of long simulations have limited climate models to much lower resolution than regional models. High-resolution (0.1°) eddy-resolving global ocean models have existed now for twenty years (Maltrud & McClean, 2005; Smith et al., 2000) but their slow performance has made them unfeasible for large timescale simulations. A doubling in horizontal resolution requires about 10x increase in computational resources (Ringler et al., 2013) because it requires 4x as many grid cells and at least 2x more timesteps (Δt halved for numerical stability). What this means in practice is that a 500 year pre-industrial spinup (Golaz et al., 2019), run on a global high-resolution 0.1° model with throughput of one simulated year per day (SYPD), would take over 1.5 years. This is not feasible and demonstrates why eddy-resolving models have found only limited application in climate modeling.

Fox-Kemper et al. (2019) predicted that by 2030 we will see more regional downscaling and an increased adoption of models built on “unstructured grids”. Historically, global ocean models have used structured grids, in which resolution for the entire grid must change in lockstep. Ringler et al. (2013) noted that “all 23 global ocean models used in the Intergovernmental Panel on Climate Change (IPCC) 4th Assessment Report were based on structured, conforming quadrilateral meshes.” Unstructured models offer a promising alternative approach in which high-resolution
can be focused on key areas where small-scale physics have an outsized significance, without the restrictive computational cost of global high-resolution. This approach is different from the structured model technique of “nested” grids in which higher-resolution regional grids are nested within larger, coarser grids. Nesting can introduce errors via the boundary conditions and adds additional complications when configuring efficient timestepping for two-way interactions between the grids. Unstructured meshes feature continuous transitions between scales with no sudden downscaling at boundaries or trade-offs between one-way and two-way nesting.

The Energy Exascale Earth System Model (E3SM) is the first global coupled climate model where all components are computed on unstructured meshes capable of variable-resolution. The E3SM project was initiated by the U.S. Department of Energy (DOE) in 2013 to address long-term science drivers including coastal flooding, sea level rise, water availability, and changes in storm frequency and extreme weather. The project identified some major limitations in existing models, particularly the ability to push the frontier into higher resolutions and to bridge the gaps across multi-scale processes. Existing models evolved organically over decades, but E3SM was a complete ground-up. The word “exascale” in the name refers to the upcoming generation of supercomputers which will be able to perform $10^{18}$ floating point operations per second (1 exa-flop). A main focus of the E3SM development has been to optimize the model code for these future high-performance computing platforms.

The ocean model component of E3SM is referred to as MPAS-Ocean, where “MPAS” stands for Model for Prediction Across Scales. Besides running in fully-coupled mode with the other E3SM models, MPAS-Ocean can be run standalone with
prescribed atmospheric forcing. Realistic ocean results require, at minimum, coupling with a sea ice model in order to develop realistic overturning (Yeager & Jochum, 2009). The first published set of realistic MPAS-Ocean simulations were based on meshes with modest variations in resolution, with grid cell size varying smoothly in latitude by a factor of two to three globally (Golaz et al., 2019; Petersen et al., 2019). These served the purpose of standard comparisons with Coordinated Ocean-ice Reference Experiments II (Griffies et al., 2014) and the Coupled Model Intercomparison Project Phase experiments (Eyring et al., 2016). In this manuscript, we present the first realistic MPAS-Ocean ocean/sea ice simulations with strongly varying resolution. We take the standard 30-60 km low-resolution mesh and add a 400 km wide coastal band around North America with 8 km resolution.

The recent study by Hoch et al. (2020) (co-authored by Rosa) validated the numerics of this MPAS-Ocean coastal-refined variable resolution mesh. We focused on the mesh-generation algorithms and problems of how to make the transition between low- and high-resolution regions. The study confirmed that the variable-resolution models are not producing any spurious numerical effects, but there were several limitations to the study. The models were forced with time-constant climatological forcing (no seasons) and there was no sea ice included in the model. Without these components the ocean model was not realistic -- even the reference case with uniform high-resolution had unrealistically weak circulation. For example, mean Gulf Stream transport through the Florida-Cuba transect was 17 Sv compared to 31 Sv in observations (1 Sv = 1x10⁶ m³/s). Instead of focusing on comparing against real-world observations, Hoch et al. (2020) compared the simple models against one
another in order to check for differences in eddy statistics and transports between the
different meshes.

In the present study, we expand on Hoch et al. (2020) by coupling with sea ice,
forcing with realistic atmosphere, and looking more in-depth at the physical
oceanography of the coastal-refined mesh. We compare against results from the
MPAS-Ocean low-resolution model as well as the high-resolution model which has
been shown to agree well with observations but is too expensive to run with today’s
supercomputers (Petersen et al., 2019). We focus our analysis on two weaknesses in
low-resolution ocean models which could potentially benefit from the variable
resolution approach: (1) eastern boundary upwelling systems and (2) western
boundary currents. Both these processes are important for global climate, show biases
in current models, depend on scales too small for most current models, and occur over
relatively constrained geographical areas.

3.1.1 Eastern boundary upwelling

The California Current System is one of the planet’s four Eastern Boundary
Upwelling Systems (Figure 3-1). These highly productive regions are small but
disproportionately important for both the global carbon cycle and human food supply.
They cover only about 0.3% of the ocean area but account for 2% of global marine
productivity (Carr, 2001) and 18.9% of fisheries catch (Pauly & Christensen, 1995,
Table 2). The key spatial scales of these upwelling systems are too fine to be captured
by modern climate models. Cold, upwelled water forms filaments 20-50 km wide that
extend 200-300 km offshore (Brink, 1987; Flament et al., 1985). These filaments are the dominant feature of chlorophyll concentration (Abbott & Zion, 1985), suggesting the biological significance of these upwelled nutrients. Considering that ocean resolution is typically 50-100 km, upwelling physics remain largely unresolved in current climate models.

![Map of mean chlorophyll concentration from SeaWiFS satellite data.](image)

*Figure 3-1. Taken from Messié & Chavez (2015). Map of mean chlorophyll concentration (mg/m$^3$) from SeaWiFS satellite data. White boxes indicate the four Eastern Boundary Upwelling Systems.*

The California upwelling region has been studied in detail using regional models with 5-10 km resolution and domains extending about 1000 km offshore (Debreu et al., 2012; Jacox et al., 2015; Patrick Marchesiello et al., 2003; Penven et al., 2006). Modeling upwelling in a global climate model has not been feasible due to the computational cost of resolving high-resolution globally. By investigating the
California upwelling region, this study seeks to determine the potential benefits of refined resolution in upwelling systems for future unstructured climate modeling applications.

3.1.2 Western boundary current

The Gulf Stream has received extensive attention in the climate modeling literature, as much for its critical role in global climate as for the fact that models consistently fail to accurately represent this important western boundary current. In particular, much of the focus has been on the “separation problem” of simulating a Gulf Stream that properly separates from the coast at Cape Hatteras, NC (Chassignet & Marshall, 2008). Schoonover et al. (2016) provided the following assessment: “Robust and accurate Gulf Stream separation remains an unsolved problem in general circulation modeling whose resolution will positively impact the ocean and climate modeling communities. Oceanographic literature does not face a shortage of plausible hypotheses that attempt to explain the dynamics of the Gulf Stream separation, yet a single theory that the community agrees on is missing.”

Air-sea flux errors as a result of sea surface temperature (SST) biases are one of the largest sources of error in climate models (Chassignet & Marshall, 2008). Most climate models’ largest SST biases occur in the North Atlantic, as a result of misrepresentation of the Gulf Stream path after separation (Keeley et al., 2012; Randall et al., 2007). This is because the Gulf Stream is a meandering SST front and even small errors spatially can result in large errors in air-sea flux.
Like most models, the low-resolution MPAS-Ocean model (Petersen et al., 2019) shows maximum SST bias in the North Atlantic (Figure 3-2). There is consensus that Gulf Stream separation and path improves at eddy-resolving resolution of at least 0.1° (Hurlburt & Hogan, 2000; Smith et al., 2000; Yeager & Jochum, 2009) however this resolution is still too expensive for widespread use in climate modeling. We seek to answer whether a variable-resolution mesh with high-resolution encompassing in the Gulf Stream separation region could produce more realistic paths.

*Figure 3-2. Zoomed-in view of Petersen et al. (2019) Figure 3e. Sea surface temperature (SST) bias in low-resolution model compared to observations. The warm bias (red region) in the North Atlantic is caused by mismatch in the location of the Gulf Stream. The modeled Gulf Stream stays too close to the coast, which is a common problem in non-eddy-resolving ocean models. Observations came from the merged Hadley Center-National Oceanic and Atmospheric Administration Optimum Interpolation data set (Hurrell et al., 2008) for the period 1948–2010.*

In addition to the errors in the Gulf Stream path, the low-resolution model of Petersen et al. (2019) also suffered from weak Gulf Stream transport. Mean transport
through the Florida-Cuba transect was 15 Sv which is less than 50% of observations. This is from the “realistic” configuration with sea ice coupling, not the simple standalone model configuration used by Hoch et al. (2020). This weak Gulf Stream led to severely weakened meridional heat transport in the North Atlantic (Petersen et al., 2019 Figure 12).

Low-resolution models often suffer from poor deep convection, and previous studies have shown how artificially altering the surface density in deepwater formation regions can strengthen the deep western boundary current (DWBC) and, as a result, the Gulf Stream (Gerdes & Köberle, 1995; Stephen G. Yeager & Jochum, 2009). The MPAS-Ocean low-resolution model demonstrated a shallow mixed layer depth (MLD) bias in the Labrador and Irminger Seas (key deepwater formation regions) consistent with weak deep convection. We include the Labrador Sea in our coastal-refined resolution region in the hopes of resolving processes necessary for establishing realistic deep convection.
3.2 Model Configuration

The E3SM Version 1 simulations presented here include active ocean and sea-ice components, MPAS-Ocean and MPAS-SeaIce. Three model configurations were considered (Table 1): (1) a low-resolution approximately 1/2° “eddy-closure” mesh \( EC60to30 \); (2) a variable-resolution North American coastal-refined mesh; and (3) a high-resolution approximately 1/10° “Rossby radius scaled” mesh \( RRS18to6 \). The low- and high-resolution meshes, the general model configuration, and the atmospheric forcing are described in detail in Petersen et al. (2019). The corresponding low-resolution E3SM fully coupled simulation with active atmosphere and land components is presented in (Golaz et al., 2019).

![Figure 3-3. Grid resolution of the three meshes: low-resolution (left), coastal-refined (middle), and high-resolution (right). Colorbar indicates grid cell size, with high-resolution regions appearing red and low-resolution regions appearing blue. Bottom](image)
row shows a zoomed-in view of Iceland which happens to fall on the coastal-refined mesh's resolution transition region.

![MPAS mesh specification](image)

**Figure 3-4.** MPAS-Ocean mesh cell size as a function of latitude for the low-resolution (blue) and high-resolution (red) models. Taken from Petersen et al. (2019) Figure 2.

**Table 3-1.** Models' setup and performance. The coastal-refined model is 6.5 times more computational expensive to run than low-resolution, but 10 times cheaper than global high-resolution

<table>
<thead>
<tr>
<th></th>
<th>Low-resolution</th>
<th>Coastal-refined</th>
<th>High-resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mesh name</td>
<td>EC60to30</td>
<td>CUSP8</td>
<td>RRS18to6</td>
</tr>
<tr>
<td>Horizontal Grid Cells (ocean)</td>
<td>235k</td>
<td>645k</td>
<td>3.69 mil</td>
</tr>
<tr>
<td>Cell Size: min–max</td>
<td>30–60 km</td>
<td>8–60 km</td>
<td>6–18 km</td>
</tr>
<tr>
<td>Vertical Layers</td>
<td>60</td>
<td>60</td>
<td>80</td>
</tr>
<tr>
<td>Time step</td>
<td>30 min</td>
<td>10 min</td>
<td>6 min</td>
</tr>
<tr>
<td>Simulated years per day</td>
<td>13.18</td>
<td>4.55</td>
<td>0.77</td>
</tr>
<tr>
<td>Total cores (ocean + sea ice + coupler)</td>
<td>960</td>
<td>2160</td>
<td>3600</td>
</tr>
<tr>
<td>Million CPU hours per century</td>
<td>0.17 (^a)</td>
<td>1.1 (^b)</td>
<td>11.2 (^b)</td>
</tr>
<tr>
<td>Cost vs. low-resolution</td>
<td>×1.0</td>
<td>×6.5</td>
<td>×65.9</td>
</tr>
</tbody>
</table>

\(^a\)compy mcnodeface
\(^b\)blues
The three meshes for this study appear in Figure 3-3. The low-resolution configuration consists of 30 km resolution in an equatorial band, 60 km resolution at mid-latitudes, and 35 km resolution about the poles. It is designed to be run with a turbulence closure parameterization of mesoscale eddy activity. Mesh resolution in the high-resolution configuration is roughly scaled by Rossby radius of deformation and varies from 18 km at the equator to 6 km at high-latitudes. This mesh is a globally eddy-permitting configuration, and all mesoscale eddy parameterizations are turned off.

The variable-resolution mesh incorporates a coast-following 400 km-wide band of enhanced 8km resolution around North America and Greenland. Resolution outside the coastal band is consistent with the low-resolution EC60to30 configuration. The variable-resolution mesh is designed to capture eddy activity in the high-resolution zone, tapering to an eddy-parameterized representation in the global domain.

This transition from eddy-resolving to eddy-closure was accomplished in two ways. First, momentum dissipation was scaled by the horizontal grid cell width $\Delta x$ in both the viscosity ($\nu_2$) and hyperviscosity ($\nu_4$) terms.

$$\nu_2 = 1000[m^2 s^{-1}] \frac{\Delta x}{30[km]} \tag{1}$$

$$\nu_4 = 1.2e11[m^4 s^{-2}] \left( \frac{\Delta x}{30[km]} \right)^3 \tag{2}$$

Non-eddy-resolving ocean models use artificially large viscosity to account for the effects of unresolved turbulence and to satisfy numerical stability constraints (Large et al., 2001). Secondly, there is an additional parameterization used at low-resolution, but
not high-resolution, to account for the effects of baroclinic instabilities in unresolved mesoscale eddies (Gent & Mcwilliams, 1990, hereafter GM). GM parameterization is typically either “on” or “off” in a model, there is no standard way of transitioning between regimes. Our solution was to alter the GM tracer Bolus velocity term, $\kappa$. We set $\kappa = 600 \text{ m}^2\text{s}^{-1}$ for grid cells larger than 30 km, and then taper linearly to $\kappa = 0$ at $\Delta x = 20 \text{ km}$. The only grid cells with $\Delta x$ between 20 km and 30 km are within the narrow resolution-transition region.

The coastal-refined mesh was created using the JIGSAW library (Engwirda, 2017) — an unstructured mesh generator that supports the construction of high quality, variable resolution Voronoi-type tessellations on a sphere. MPAS-Ocean meshes consist of collections of Voronoi polygons and Delaunay triangles, with an unstructured Arakawa C-grid discretization. The low- and high-resolution meshes were created with a previous mesh generation method, based on Lloyd’s algorithm (Jacobsen et al., 2013).

MPAS-Ocean solves the primitive equations, which are incompressible, hydrostatic, and Boussinesq (Ringler et al., 2013). Prognostic equations for momentum, thickness, and tracers are solved with a split-explicit baroclinic/barotropic time stepping scheme. The vertical coordinate is “z-star” (not terrain-following) and layers range from 2 m thick at the surface to 150 m at depth. The vertical mixing scheme is the K-Profile Parameterization (Large et al., 1994; Roekel et al., 2018) and is applied with the CVMix library (Griffies et al., 2017).

Simulations are spun up from an initial condition of Polar Science Center Hydrographic Climatology, version 3 (Steele et al., 2001). Atmospheric forcing and
run-off are from the CORE-II forcing dataset (Yeager & Large, 2008). MPAS-Seaice is based on the CICE sea-ice model 4 (Hunke et al., 2017) but has been adapted from the CICE quadrilateral grid to the MPAS framework polygonal cells.
3.3 Results

3.3.1 Eastern Boundary: California Current Upwelling

Snapshots of SST at high-resolution show cold upwelled water in thin filaments extending hundreds of kilometers from the shelf, but at low-resolution these filaments are absent (Figure 3-5). The coastal-refined model shows similar behavior to the high-resolution model within the refined region. SST gradient (Figure 3-5 d-f) more clearly shows the distribution of these cold filaments which are characterized by strong SST fronts/gradients. The multi-year summertime (upwelling season) mean SST gradient (Figure 3-5 g-i) shows the mean distribution of these SST fronts. The coastal-refined model has a thinner coastal band of maximum SST gradients, but there is a clear improvement over the low-resolution model.
Figure 3-5. California Current System upwelling seen in sea surface temperature (SST). Plots show early July SST snapshot (top row), early July SST gradient magnitude (middle row), and multi-year summertime (JAS) mean SST gradient (bottom row) for the three runs: low-resolution (left column), coastal-refined (center column), and high-resolution (right column). The grey line on the coastal-refined run (center column) indicates the approximate mesh transition -- it follows location of 12 km wide grid cells. East of the transition is the refined region with 8 km resolution. The SST and SST gradient snapshots show that the coastal-refined model produces a visible improvement in the representation of cold, upwelling filaments. Some filaments in the high-resolution model extend beyond the coastal-refined region and would not be resolved in the
coastal-refined model but, as seen in the mean SST gradient (bottom row), strongest gradients occur within the refined region.

Sea surface height (SSH) snapshots (Figure 3-6 a-c) reveal increased mesoscale eddies in the coastal-resolution within the refined region. The low-resolution SSH snapshot (Figure 3-6 a) closely resembles the low-resolution time-mean (Figure 3-6 d) and shows no discernable eddying. The mean SSH contours in both the high-resolution and coastal-refined models show that the mean geostrophic circulation is characterized by persistent meanders (Figure 3-6 e-f). These are caused by interactions with bathymetry (Marchesiello et al., 2003). The low-resolution mean SSH contours are much straighter and don’t feature these meanders.
Figure 3-6. California Current System SSH snapshot (top row), SSH mean (middle row), and log10 of SSH variance (bottom row) are shown for the three runs: low-resolution (left column), coastal-refined (center column), and high-resolution (right column). For SSH snapshot and mean, contour lines are drawn every 2 cm. The grey line on the coastal-refined run (center column) follows the 12 km grid cell size contour. SSH snapshots show mesoscale eddy activity in the coastal-refined run extending out to the resolution transition contour. The mean SSH patterns also show that the coastal-refined run offers improvements over the low-resolution, with persistent meanders in the SSH contours that agree better with the high-resolution run. Maximum SSH variance for the high-resolution run is 100+ km offshore. The coastal-refined model shows substantial
improvement in SSH variance compared to low-resolution, with increased SSH variance throughout the refined resolution region.

SSH variance in the high-resolution model reaches a local maximum away from the coast, which agrees with observations (P. Marchesiello & Estrade, 2009). SSH variance in the coastal-refined model is lower magnitude but also shows a local maximum away from the coast, before dropping off at the mesh transition. The low-resolution model shows very low SSH variance, with maximum variance closest to the coast.

To quantify the enhanced mesoscale activity in the CR model, we calculated the wavenumber power spectra of SST and SSH (Figure 3-7) following a similar method as D.P. Wang et al. (2009). Power spectra were computed along a 1500 km path approximately following the coast within the refined region from 31°N to 42°N, as indicated on the inset map in Figure 3-7. Data were interpolated to this line at 8 km spacing. Linear trend was removed and a Hanning window function was applied. The time snapshot transects were then separated from each other with a 1500 km pad of zeros and a fast Fourier transform was applied.
Figure 3-7. Wavenumber power spectra along the central California coast to characterize the key lengthscales for SST (top) and SSH (bottom). On the inset map, the red line indicates the track along which the calculations were performed. Notice that the track was defined inside of the refined resolution region (grey contour). Low-resolution (blue), coastal-refined (black), and high-resolution (red) runs were
compared in addition to satellite observations (green). These results confirm that, compared to the low-resolution model, the coastal-refined model provides substantial improvements at representing mesoscale features. In these power spectra, we can see that the coastal-refined model is in good agreement with the high-resolution model.

The power spectra show that the coastal-refined results for both SST and SSH agree with the high-resolution results though with slightly less energy at most wavenumbers. The low-resolution shows substantially lower energy than the other two models, as expected. For a reference, we included the SST power spectra from Advanced Very High Resolution Radiometer (AVHRR) satellite observations. The AVHRR power spectra fall between the low-resolution and the coastal-refined. The lower energy of the observations could be caused by the re-gridding process and/or the numerical models are too eddying. SSH power spectra showed better agreement between observations (AVISO satellite altimetry), coastal-refined model, and high-resolution model. This is in contrast to the analysis by Petersen et al. (2019) which found that the high-resolution MPAS-Ocean model produced higher SSH variance than AVISO observations.

3.3.2 Western Boundary: The Gulf Stream

3.3.2.1 Path and eddying

Looking first at the surface speed (Figure 3-8), we see that the variable-resolution model develops a realistically narrow Gulf Stream, as opposed to the diffuse western boundary current in the low-resolution model. The coastal-refined model surface
speed snapshot (Figure 3-8 b) even includes cold core rings within the refined region. To quantify the eddy activity in the region of the western boundary current, the eddy kinetic energy (EKE) is shown in Figure 3-8 g-i. EKE is uniformly low in the low-resolution model as expected. There is enhanced EKE in the coastal-refined model, particularly where the Gulf Stream separates from the coast. This demonstrates that the refined region is supporting the formation of mesoscale eddies. The eddying regime ends sharply at the transition into the low-resolution region of the model, where the mesoscale eddies are killed by grid-scale dissipation (Danilov & Wang, 2015).
Figure 3-8. Surface speed snapshot (top row), surface speed mean (middle row), and surface eddy kinetic energy (bottom row) are shown for the three runs: low-resolution (left column), coastal-refined (center column), and high-resolution (right column). The grey line on the coastal-refined run (center column) follows the 12 km grid cell size contour, indicating the resolution transition from ~60 km resolution east of the grey contour to 8 km resolution west of the contour. The surface speeds show that the coastal-refined run has a clearly defined Gulf Stream (GS) of similar width to the high-resolution run but that the speeds are too weak. The low-resolution GS is unrealistically diffuse but, integrating the transports, we find that the low-resolution and coastal-refined runs have similar GS volume transport and that the high-resolution GS transport is about 50% stronger. Eddy kinetic energy (EKE) shows improvement in the high-resolution region of the coastal-refined run and is rapidly dissipated outside of the
refined resolution region, as predicted. This is because the low-resolution regions of the mesh are too coarse to resolve mesoscale eddy activity.

Figure 3-9. Sea surface height (SSH) snapshot (top row), SSH mean (middle row), and log10 of SSH variance (bottom row) are shown for the three runs: low-resolution (left column), coastal-refined (center column), and high-resolution (right column). For SSH snapshot and mean, contour lines are drawn every 5 cm. The grey line in the coastal-refined model indicates the resolution transition.
Figure 3-10 provides a detailed look at the path of the Gulf Stream as it separates from the coast and approaches the resolution transition region. The high-resolution model produces a highly realistic Gulf Stream path with separation from the coast at Cape Hatteras and continuing eastward as a meandering jet. In the coastal-refined model, the Gulf Stream continues north to about 38°N before reflecting back southward. This separation meander appears highly persistent. The individual Gulf Stream pathlines reveal a surprising result (Figure 3-10 c). The coastal-refined model Gulf Stream largely stays confined within the refined resolution region. This result is revisited in the Discussion.
Figure 3-10. Gulf Stream separation represented by surface speed root mean square (RMS) in the coastal-refined run (a) and high-resolution run (b). The high-resolution run’s separation path and spread is more realistic, while the coastal-refined run follows the coast too far north and then turns back southward before weakening when it passes into the low-resolution region (west of the grey contour line). The path lines of
individual Gulf Stream instances (c) are shown for the coastal-refined run (black) and the high-resolution run (red). These pathlines were calculated by following the -0.2 m SSH contour for the first record of the month, for 3 years (36 total pathlines for each run are shown).

3.3.2.2 Transports

Despite the realistic Gulf Stream width and eddying, the coastal-refined model Gulf Stream is substantially weaker than observations and the high-resolution model. Mean Gulf-Stream transports at 26.5°N for the low-resolution and coastal-refined models are 28 Sv and 29 Sv, respectively, compared to 46 Sv for the high-resolution model (Figure 3-11). To understand the causes of the weak Gulf Stream transports, we computed meridional transports through a transect at 26.5°N. This latitude was chosen to align with the RAPID-MOCHA cross-basin mooring array.

We separated the flow into three components: (1) Gulf Stream, (2) deep western boundary current (DWBC), and (3) “rest of the transect”. Together, these meridional flows make up the wind- and buoyancy-driven components of the Atlantic Meridional Overturning Circulation (AMOC). We defined the spatial extents of three components as follows: Gulf Stream is west of -73°E and shallower than 1000 m; DWBC is west of -73°E and deeper than 1000 m; rest of transect is east of -73°E and all depths. A schematic of the three integration regions can be seen on the inset plot in Figure 3-11.
Figure 3-11. A summary of meridional transports across 26.5°N. Here, transports are separated into 3 parts as follows: the Gulf Stream (GS) is the sum of flow west of -73E and above 1000 m depth, the deep western boundary current (DWBC) is the sum of flow west of -74 E and below 1000 m depth, and the rest of the basin is the sum of flow east of -73E, integrated through all depths. The inset plot shows the depth across 26.5N and the three flow integration regions superimposed. The bar plots represent the mean flow for each of the three runs: low-resolution (blue), coastal-refined (black), and high-resolution (red). We see that for all regions, the low-resolution and coastal-refined transports are in agreement with each other indicating that this coastal-refined run did not provide a significant improvement over the weak low-resolution transports.

Conservation of mass dictates that the northward transport of the Gulf Stream transport nearly balance the southward transport of the buoyancy-driven DWBC and wind-driven Sverdrup transport through the rest of the transect (net inflow of Pacific water into Arctic Ocean is on the order of 1 Sv). The wind-driven rest of transect showed similar transports between the three models: -17 Sv for low-resolution, -20 for coastal-refined, and -16 Sv for high-resolution. The DWBC however was severely
weakened in the low-resolution and coastal-refined models (-11 Sv) while the high-resolution DWBC was more than double the transport at -28 Sv. This suggests that the primary difference between the models is the strength of the DWBC.

The DWBC is fed by deepwater formation, and wintertime mixed layer depth (MLD) can identify regions of deep convection (deepwater formation). The low-resolution and coastal-refined models show much shallower MLD than the high-resolution model. The Labrador Sea has maximum winter MLD less than 300 m while the high-resolution model has MLD over 1000 m in this key North Atlantic deep water (NADW) formation region. Plotting sea surface salinity (Figure 3-12) shows that these same regions with shallow MLD in the low-resolution and coastal-refined models also have fresher surface waters than the high-resolution model.
Figure 3-12. Left column: North Atlantic mixed layer depth in winter (January, February, March) for low-resolution (top), coastal-refined (middle), and high-resolution (bottom). Right column: Surface salinity in October, preceding the deepwater formation season. Red contour at 34.5 PSU. Deep mixed layer depth (MLD) indicates regions of deepwater formation. The high-resolution run shows two main regions of North Atlantic deepwater formation: the Labrador Sea (red circle) and the Greenland Sea (orange circle). In the low-resolution and coastal-refined runs, there is a somewhat deep MLD in the Greenland Sea but extremely shallow MLD in the Labrador Sea, even though the Labrador Sea is well within the refined resolution region of the coastal-refined mesh. Deepwater formation is critical for the thermohaline circulation and
explains why the low-resolution and coastal-refined runs had such weak deep western boundary currents.
3.4 Discussion

The goal of variable-resolution modeling is to gain the computational benefits of using fewer total grid cells, but without introducing any errors/biases as a result of the multiple resolutions within the model. “Simulated ocean dynamics should be insensitive to whether that scale is present in a multi-resolution simulation or a quasi-uniform simulation” (Ringler et al., 2013). Our results showed that this goal was not satisfied – results within the coastal-refined region were not a faithful reproduction of the high-resolution model results. In this Discussion, we analyze the coastal-refined model’s successes and shortcomings.

First, we discuss the differences in model benefits between the eastern boundary and the western boundary. Second, we present a hypothesis to explain biases in the Gulf Stream path in the coastal-refined model. Third, we describe a mechanism by which this altered Gulf Stream path created a feedback of additional model issues, ultimately weakening the DWBC and AMOC. Finally, we make recommendations to the E3SM leadership (and unstructured model users more generally) on how to design their future variable-resolution models.

3.4.1 Local vs. non-local improvements

The coastal-refined model showed great improvements for California upwelling but more complicated results for the Gulf Stream. We frame these spatially inhomogeneous improvements as the difference between locally-forced and non-locally-forced processes. We expect locally-forced processes, like coastal upwelling,
to benefit the most from local resolution refinement. Non-locally-forced processes, like western boundary currents, depend on processes within as well as outside the refined region, and therefore are more prone to unintended biases when using regional resolution enhancement.

To understand local vs. non-local forcing, it is useful to consider regional ocean models. As shown in Manuscript 1, small-domain regional models apply non-local effects via open boundary conditions. A small ROMS model of California upwelling does not depend much on boundary conditions -- even simple closed boundaries will generate upwelling as long as there are along-shelf winds. The Gulf Stream, however, is a response to non-local processes. A 400 x 400 km (width of the refined region) ROMS models, forced only with local winds, will not develop a realistic western boundary current. ROMS models of this size and smaller are frequently used to study submesoscale Gulf Stream physics (e.g. Gula et al., 2014) but they rely on boundary conditions from basin-scale models.

Our results give hope to the possibility of realistic Gulf Stream representation in a variable-resolution model. The Gulf Stream is largely driven by equatorward Sverdrup transport in the subtropical gyre resulting from wind stress curl (Ekman pumping). Although this non-local forcing occurs over the low-resolution region of the model, we found that the Sverdrup transport in the low-resolution region actually agrees well with the high-resolution model (Figure 3-11). Since this process appears to be well-represented at low-resolution, we do not predict that it will be a barrier towards a realistic coastal-refined Gulf Stream. Non-locally-forced processes do not necessarily
rule out the use of variable resolution, but the model must skillfully represent the non-local effects and allow communication across resolutions.

Although we showed that local resolution refinement improved upwelling in the California Current system, this is not to say that all upwelling systems will be so simple. For example, the equatorial eastern Indian Ocean upwelling is significantly impacted by non-locally-generated Kelvin waves propagating along the equator (G. Chen et al., 2016). If these Kelvin waves are not resolved, the model would likely produce unrealistic upwelling even if there is refined resolution in the upwelling region. Whenever non-local forcing is important, there may be a need to refine additional regions outside the specific upwelling area. There are no ‘one size fits all’ rules; mesh design requires careful consideration of the specific physical processes being investigated.

3.4.2 Gulf Stream path affected by mesh transition

To explain the landward shift in the coastal-refined model Gulf Stream, our initial hypothesis was that the path bias was caused by the weak DWBC. Observational studies have suggested that a stronger DWBC precedes a southward shift in the path of the Gulf Stream (Peña-Molino & Joyce, 2008; Stephen G. Yeager & Jochum, 2009). However, those observed interannual shifts in the mean Gulf Stream path were small relative to the large landward bias in the coastal-refined model.

Instead, it appears more likely that the Gulf Stream path bias is a result of unexpected interactions at the resolution transition, which result in the Gulf Stream
getting “trapped” in the coastal-refined region. Persistent meanders appear along the transition line, with the Gulf Stream flowing eastward, encountering the resolution transition, and then re-directing northward (Figure 3-14a). As described earlier, horizontal viscosity is scaled by grid cell size. Thus, the eastward-flowing Gulf Stream experiences the resolution transition as a boundary into higher viscosity. Flow passing through variable resolution/viscosity regions has been considered by Danilov & Wang (2015) (Figure 3-13), however those experiments featured jets flowing normal to the resolution transition. In our case, it is important that the Gulf Stream encounters the transition at an oblique angle.

![Diagram showing schematic of model with zonal flow imposed on a variable resolution mesh with refined resolution](image)

**Figure 3-13.** From Danilov & Wang (2015) Figure 1a+c. Top frame shows schematic of model with zonal flow imposed on a variable resolution mesh with refined resolution.
in the middle. Bottom frame shows snapshot of relative vorticity normalized by the local value of the Coriolis parameter. The relative vorticity field shows the formation of eddies on the fine mesh which decay on the coarse mesh.

When the Gulf Stream encounters the resolution transition, the “right side” (when looking downstream) is slowed more than the left side. This cross-stream shear has the sign of negative relative vorticity. Conservation of potential vorticity (PV) predicts that, in the absence of a significant bathymetric gradient, there must be a change in planetary vorticity ($f$) equal and opposite to the change in relative vorticity (Figure 3-14b). Northward propagation towards higher $f$ agrees well qualitatively with the observed meanders.

**Figure 3-14.** Coastal-refined model Gulf Stream path interacting with mesh transition region. (a) zoomed-in view of mean SSH from Figure 3-9. Red arrows parallel to SSH contours highlight coherent Gulf Stream meanders in presence of mesh transition. (b) schematic view showing conservation of potential vorticity hypothesis. Gulf Stream encounters mesh transition at an angle. Higher viscosity in the low-resolution region
induces an anti-cyclonic horizontal shear. This negative relative vorticity is balanced by increasing planetary vorticity (northward path).

Does this hypothesis make sense quantitatively based on the observed velocities and length scales? In most large scale flow, planetary vorticity is an order of magnitude larger than relative vorticity, but here we are interested in the change in these quantities. Below we calculate the approximate shear that would be required to balance the change in planetary vorticity associated with a northward meander of 3° latitude from 39°N to 42°N (Figure 3-15).

Relative vorticity for a zonal jet:

\[ \xi = \Delta u / \Delta y \]

Conservation of potential vorticity (assume constant layer thickness):

\[ \Delta f = -\Delta \xi \]

Estimated horizontal shear:

\[ \Delta f = 0.058 \times 10^{-4} \text{ s}^{-1} \]
\[ \Delta y = 50 \text{ km} \]
\[ \Rightarrow \Delta u = -0.29 \text{ m/s} \]

Our scaling analysis predicts that a reduction in speed on the order of 0.3 m/s across the Gulf Stream would be required to elicit a 3° northward meander. Referring to Figure 3-15, this seems like a reasonable order of magnitude. Therefore, we conclude that the scaling analysis does not rule out the hypothesis that the coastal-
refined model’s Gulf Stream path bias is the result of mesh-generated negative relative vorticity at the transition into the high viscosity region. This is the result of the refined region cutting through the Gulf Stream’s intended path.

![Figure 3-15. Close-up of surface speed for a Gulf Stream meander in the coastal-refined model near the mesh transition.](image)

### 3.4.3 DWBC shut-down

The Gulf Stream’s shifted path in the coastal-refined model has wide-reaching effects beyond just the SST bias discussed in the Introduction. In this section, we propose a mechanism by which the shifted path leads to the shutdown in Labrador Sea deepwater formation and weakening of the AMOC.
The Labrador Current (LC) is a surface current which follows the continental slope equatorward out of the Labrador Sea around Newfoundland and towards the Gulf of Maine. This is not to be confused with the DWBC which follows a similar path but is a deep sub-surface current. The LC is fed by freshwater from glaciers and is characterized by colder and fresher water than the Gulf Stream. The Grand Banks is a shallow (<200 m) plateau and the “Tail of the Grand Banks” (TGB, Figure 3-16) represents a choke point between the subpolar and subtropical regions (Fratantoni & McCartney, 2010).

Figure 3-16. Schematic of Labrador Current and Gulf Stream in vicinity of the Grand Banks. Taken from Fratantoni & McCartney (2010).
When the Gulf Stream path is too close to the TGB, it impinges on the LC and prevents the southward advection of low-salinity Labrador Sea water. Figure 3-17 compares the salinity distributions for the coastal-refined and high-resolution models. In the high-resolution model, the high-salinity Gulf Stream is located offshore of the TGB which allows the low-salinity LC to pass around the chokepoint. In the coastal-refined model, the Gulf Stream path is pushed up against the TGB and the LC is forced to retroflect back northward. This leads to a dramatic freshening of the Labrador Sea.

The presence of low-salinity water in the surface layer suppresses deep convection. Multiple modeling studies have performed “hosing” experiments where freshwater is added to the Labrador Sea and the result is a weakening or even shutdown of the thermohaline circulation (Renssen et al., 2002). Our LC retroflection at the TGB essentially led to a freshwater hosing of the Labrador Sea which killed deep convection and severely weakened the DWBC.
Figure 3-17. Salinity at 250 m depth level in coastal-refined model (a) and high-resolution model (b). The white regions are shallower than 250 m. Blue arrow indicates Tail of the Grand Banks (TGB). Under realistic conditions (b), the Labrador Current transports low salinity southward from the Labrador Sea, around the TGB. In the coastal-refined model, the high-salinity Gulf Stream can be seen impinging on the TGB. As a result, the Labrador Current retroflects and leads to a freshening on the Labrador Sea.
A review of the mechanism:

(1) Gulf Stream path trapped in the coastal-refined region (Figure 3-10).
(2) This causes the Gulf Stream to impinge on Labrador Current at TGB (Figure 3-17).
(3) Labrador Current retroflects northward and freshens the Labrador Sea.
(4) Fresh Labrador Sea cannot develop deep convection (Figure 3-12). DWBC is weakened.
(5) Weaker thermohaline circulation [and no change in wind-driven Sverdrup transport] results in a weaker Gulf Stream (Figure 3-11).

3.4.4 Next steps

As a result of this study, I have presented the E3SM development team with recommendations for a redesigned coastal-refined mesh. In Figure 3-18 I propose the new refined region which extends out further in the North Atlantic to avoid interactions with the Gulf Stream. The original coastal-refined mesh applied a simple 400 km-wide band of high-resolution without regard for the local physical oceanography. The new proposed mesh is informed by satellite observations of the Gulf Stream’s path and the results from this study which showed Gulf Stream steering as a result of poorly placed resolution regions. We estimate that the new model would only be about 10% more computationally expensive. This physics-informed approach to mesh design should be considered in future variable-resolution modeling of other phenomena. It is necessary to look beyond the local region of interest and consider if
there are non-local processes that are unresolved or if there are significant currents which might be passing in/out of the resolution transition zone.

Figure 3-18. White line indicates the extent of the expanded coastal-refined region we recommend in order to avoid mesh-transition interactions with Gulf Stream. Color shows mean SSH variability from AVISO satellite.
3.5 Conclusions

In this study, we expanded on the recent MPAS-Ocean coastal-refined mesh experiments by Hoch et al. (2020) by adding in sea ice-coupling and realistic atmospheric forcing. This improved simulation design resulted in a more realistic Gulf Stream and AMOC in our high-resolution model, which was not achieved in the simpler Hoch et al. (2020) setup. We looked at upwelling in the California Current system and found improvements in the representation of cold upwelling filaments, eddy kinetic energy, and mean SSH structure. When looking at the western boundary, we found that the coastal-refined model does not agree with the high-resolution Gulf Stream/AMOC – a conclusion which was missed in the previous study.

Weak Gulf Stream transport in the coastal-refined model is primarily the result of weak DWBC, not weak wind-driven Sverdrup transport in the low-resolution subtropical gyre. From a modeling perspective, it is a good sign that the Sverdrup transport was well represented at low-resolution since this is a large region that would have been expensive to model if eddy-resolving resolution was needed.

The coastal-refined model Gulf Stream was steered by the mesh transition to stay within the refined region. The increased horizontal viscosity at the transition produced a negative relative vorticity which was balanced by a northward meander in order to gain planetary vorticity. This is the first study to document the behavior of a jet being steered upon encountering a resolution transition at an oblique angle.

The altered Gulf Stream path halted the southward transport of the Labrador Current and led to a freshening of the Labrador Sea. As predicted by previous Labrador Sea “hosing” experiments, this freshening shut off deep convection which
led to a slow down in the AMOC and Gulf Stream. Our model accidentally produced an experiment which offers valuable insight into the role of Gulf Stream-Labrador Current interactions on climate. Previous studies have artificially added freshwater to the Labrador Sea but our results show extreme sensitivity of the overturning circulation to changes in the position of the Gulf Stream.

The most immediate outcome of this study is the proposed design of a new coastal-refined mesh. We have suggested that the E3SM team redesign their mesh to feature a larger refined region in the North Atlantic to accommodate the realistic path of the Gulf Stream. We hypothesize that this new mesh would fix the bias in the Gulf Stream path as well as improve deepwater formation in the Labrador Sea and thus increase Gulf Stream transport to realistic levels.
3.6 Appendix A: Eddy propagation

3.6.1 Eastern boundary

In order to understand how eddies behave in the coastal-refined model, we look at the propagation of SSH anomalies using time-longitude (Hovmoeller) diagrams. Figure 3-19 shows a time-latitude diagram at 40°N from the coast of California to about 1500 km west into the North Pacific. SSH anomalies primarily propagate westward, away from the coastal-refined region. The coastal-refined model shows agreement with the high-resolution model in the refined region, with energy getting filtered out at the resolution transition region.

The 2 cm/s reference speed is based on the climatological predictions of the first baroclinic gravity-wave phase speed (Chelton et al., 1998). Marchesiello et al. (2003) also found propagation speeds of about 2 cm/s in their numerical model of this region, and described the propagating signal as more eddy-like than wave-like. This 2 cm/s characteristic speed appears to agree well with the results from our models as well.
Figure 3-19. Sea surface height anomaly Hovmoeller diagrams for low-resolution (left), coastal-refined (center), and high-resolution (right) at 40N in the California Current. Time on the y-axis increases upward and the gray line indicates a reference slope of 2 cm/s westward propagation. Arrows indicate SSH anomalies propagating westward away from the coast at approximately 2 cm/s. The coastal-refined run appears to agree well with the high-resolution inside the refined region and then dissipates some eddy energy at the resolution transition. Outside of the refined region, results resemble the low-resolution model.

3.6.2 Western boundary

Figure 3-20 shows Hovmoeller diagrams for the western North Atlantic at 34.8°N, the latitude of Cape Hatteras where the Gulf Stream is observed separating from the coast. Noticing the colorbar range, the SSH anomalies are of larger magnitude than those seen in the California Current System (Figure 3-19). The other main difference is that energy is primarily propagating into the refined region. This agrees with theory, models, and observations which show that Rossby waves and most eddies propagate westward (Petersen et al., 2013). However, we do see some eastward-propagating anomalies, including some energy exiting the refined region in the coastal-refined model. These patterns are consistent with eddies sheared off of the Gulf Stream and earlier plots of eddy kinetic energy (Figure 3-8h) rapidly dissipating at the mesh transition.

The reference speed of 4 cm/s comes from Osychny & Cornillon (2004) observations of Rossby wave propagation speed in the North Atlantic at 35°N. This speed is about twice the theoretical estimate for the first baroclinic mode from
climatological in situ data (Chelton et al., 1998). The coastal-refined and high-resolution model results appear to show good agreement with the observed 4 cm/s propagation speed as opposed to the slower estimate based on hydrography.

Figure 3-20. Sea surface height anomaly Hovmoeller diagrams for low-resolution (left), coastal-refined (center), and high-resolution (right) in the North Atlantic at 34.8°N, the latitude of Cape Hatteras, where the Gulf Stream should separate from the coast. Time on the y-axis increases upward and the black and white dashed line indicates the reference slope of a 4 cm/s westward propagation. At high-resolution, waves/eddies can be seen propagating westward across the basin while the low-resolution model shows very little activity. The coastal-refined run behaves as predicted: waves/eddies are generated near the transition to the coastal-refined region and propagate westward towards the coast. In addition, there are some signs of eastward-propagating anomalies in all three models.
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