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SUBESTUARINE CIRCULATION AND DISPERSION
IN NARRAGANSETT BAY

BY
CHRISTELLE BALT

A DISSERTATION SUBMITTED IN PARTIAL FULFILLMENT OF THE
REQUIREMENTS FOR THE DEGREE OF
DOCTOR OF PHILOSOPHY
IN
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UNIVERSITY OF RHODE ISLAND

2014

DOCTOR OF PHILOSOPHY DISSERTATION
OF
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2014

ABSTRACT

The subestuaries in the northern reaches of Narragansett Bay (Rhode Island, USA) are susceptible to summertime water quality impairment. High nutrient loading of these systems leads to eutrophication and recurrent hypoxia, which cause ecosystem degradation and incidents of organism mass mortality. Management efforts to improve water quality include reductions in nutrient input to Narragansett Bay. We hypothesize that nutrient reductions may not sufficiently improve the health of severely impacted regions due to physical processes that contribute to water quality deterioration, including high vertical density stratification and low horizontal exchange. Dispersion depends on vertical turbulent mixing and advection. Circulation and dispersion are assessed using high-resolution observational and numerical modeling techniques. A realistic Regional Ocean Modeling System (ROMS) model for Narragansett Bay is validated by comparison with data and by doing a turbulence closure scheme assessment. It is found that the model provides an accurate representation of Narragansett Bay hydrodynamics when a $k - \varepsilon$ scheme is implemented. High agreement exists between the model and data of tidal velocity and sea surface elevation, as well as subtidal velocity, temperature, and salinity. Future optimization of the model involves refinement of some forcing specifications. Observations of hydrodynamics in Greenwich Bay, one of the most severely impaired subestuaries of Narragansett Bay, and the adjacent Warwick Neck channel show the response of the subsystem to different forces. The data show that the wind exerts a dominant influence on the subtidal flow. Wind directed predominantly eastward improves exchange between the basins of Greenwich Bay relative to northward wind. Vertical turbulent mixing by tidal shear is generally low in Greenwich Bay indicating that bottom water may become isolated from the atmosphere for prolonged periods of time. This situation is conducive to

hypoxia. Diurnal wind can cause an increase in vertical turbulent mixing and can force intratidal residuals by its interaction with the semidiurnal tide. The numerical model is used to carry out semi-idealized scenario experiments to determine the influence of specific forcing conditions on circulation, dispersion, and flushing of Greenwich Bay. Model experiments show that residence time depends on wind direction, since flushing occurs largely by wind-driven advective dispersion. The dominant sustained and diurnal wind conditions during the summer months lead to retention by weak flow and horizontal recirculation structures.

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DEDICATION

In memory of Dr Johann Lutjeharms

To Dr Tom Rossby

& my family

PREFACE

The dissertation was prepared in manuscript format. It contains three separate manuscripts, each to be submitted to a specific journal.

Manuscript 1: *Validation of a numerical model for Narragansett Bay with turbulence closure scheme comparisons* by Christelle Balt, Chris Kincaid, David S. Ullman, Dan L. Codiga, James N. Kremer, Jamie M.P. Vaudrey, Justin M. Rogers, and Deanna L. Bergondo was prepared for submission to *Journal of Geophysical Research*

Manuscript 2: *Observations of the physical processes that influence water quality of Greenwich Bay and upper Narragansett Bay* by Christelle Balt and Chris Kincaid was prepared for submission to *Estuaries and Coasts*

Manuscript 3: *Modeling the relationship between Greenwich Bay circulation, flushing efficiency, and environmental forcing conditions* by Christelle Balt, Chris Kincaid, and David S. Ullman was prepared for submission to *Ocean Dynamics*

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MANUSCRIPT 1

**Validation of a numerical model for Narragansett Bay with turbulence
closure scheme comparisons**

by

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Kremer²; Jamie M.P. Vaudrey²; Justin M. Rogers³; Deanna L. Bergondo⁴
is in preparation for submission to *Journal of Geophysical Research*

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Abstract

A three-dimensional hydrodynamic Regional Ocean Modeling System (ROMS) model for Narragansett Bay is validated using observed tidal and subtidal time series of sea surface elevation, currents, and hydrography. The model is forced with observations of tides, freshwater runoff, and atmospheric conditions for the summer of 2006. The model serves as physical driver for a hybrid ecosystem model for the estuary, contributing information about exchange of constituents between discrete sub-regions of the bay. Seeing as exchange is strongly dependent on advection and vertical turbulent mixing, our motivation is to test a number of turbulence closure schemes in ROMS to determine which of the schemes are appropriate for this application. A further goal is to do a detailed validation of the NB-ROMS model, which is the version that implements the most successful scheme. Skill assessments show that the majority of schemes perform similarly well and the $k - \varepsilon$ scheme is selected for NB-ROMS. Accuracy of the model is quantified by a skill parameter, which is defined to indicate good data-model agreement when it exceeds 0.65. The NB-ROMS model accurately predicts tidal sea surface elevation and currents, as well as subtidal currents, temperature, and salinity with spatially averaged skills of 0.98, 0.70, 0.88, and 0.82, respectively. Small discrepancies between the NB-ROMS model and observations of tidal temperature and salinity, and subtidal sea surface elevation cause spatially averaged skills of 0.65, 0.61, and 0.43, respectively. The discrepancies are attributed to forcing specifications. Overall the model is successful in representing tidal and subtidal circulation and hydrography in Narragansett Bay.

1.1 Introduction

Narragansett Bay is a coastal plain estuary and an important natural resource for the communities of Rhode Island and Massachusetts. It is subject to severe anthropogenic pressures, including bacterial contamination, metal pollution, and excessive nutrient loading [1, 2]. These stresses lead to summertime water quality degradation of the hydrodynamically complex subsystems near the head of Narragansett Bay. Prolonged dissolved oxygen deficiency, caused by bacterial decomposition of organic matter produced by blooms, is particularly damaging to Narragansett Bay ecosystems [3, 4, 5].

We hypothesize that physical processes are instrumental in determining water quality. For example, dissolved oxygen depletion of a subsystem can result due to inadequate replenishment of ventilated water when exchange with the main estuary is inhibited. Subestuarine processes that determine dispersion characteristics of Narragansett Bay, namely vertical and lateral advection and mixing, are not well understood. Detailed information regarding the hydrodynamics is important to understand the physical controls that regulate mixing and the exchange of water and its constituents in impaired areas. High-resolution numerical modeling is invaluable for obtaining comprehensive knowledge of estuarine hydrodynamics.

A three-dimensional, high-resolution numerical model of Narragansett Bay is verified in this study. The model was developed from previous modeling efforts [6, 7] using the Regional Ocean Modeling System (ROMS) [8, 9]. The model is a component of the NOAA Coastal Hypoxia Research Project (project NA05NOS4781201) and is used to investigate multi-scale physical processes and to provide the net advective and diffusive tracer exchanges between the spatial domains of an ecological box model [10].

A number of vertical mixing parameterizations, also referred to as turbulence

closure schemes, are available in ROMS [9]. These schemes enable the closure of the Reynolds averaged Navier-Stokes (RANS) equations and determine the solutions of subgrid scale vertical turbulent fluxes [11, 12]. The choice of turbulence closure scheme in models can dramatically affect the accuracy of mixing processes and circulation [13, 14, 15, 16, 17].

The model validation consists of two parts, namely a sensitivity analysis to compare model accuracy when different vertical mixing parameterizations are implemented, and a detailed assessment of the model performance when one successful vertical mixing parameterization is specified. The version of the model that implements the successful turbulence closure scheme is referred to as the NB-ROMS model. Accuracy of the model is determined by comparing model output to observations of tidal and subtidal tracers, velocity, and sea surface elevation. Model performance is quantified using the Willmott skill (WS) [18, 19, 16].

The turbulence closure schemes compared in the sensitivity experiments are summarized here, and described in detail in Section 1.2.3. The experiments include five statistical two-equation models and an empirical model. Statistical and empirical models differ in the procedures used to solve the vertical turbulent fluxes [11, 12]. Two-equation models rely on instantaneous local flow properties in the calculation of fluxes, whereas empirical models incorporate both local and nonlocal effects by employing separate treatments for the surface boundary layer and the ocean interior [11, 12]. Local fluxes result from gradients at a particular location; an example of nonlocal fluxes are convective plumes introduced at an interior location by surface cooling. Two-equation models are named as such because they solve two transport equations, one for turbulent kinetic energy and another for turbulent length scale. The different two-equation models vary in their treatments of turbulent length scale in the calculation of eddy viscosity and eddy diffusivity

coefficients [15]. The turbulent length scale (l) and a length scale-related quantity, the turbulence frequency (ω), are functions of turbulent kinetic energy (k) and dissipation rate (ε). The five statistical two-equation models compared are $k - \varepsilon$ [20, 21, 22, 23], Mellor-Yamada 2.5 (MY25) [24], $k - \omega$ [25, 26, 27, 28], $k - kl$ [24, 28], and *gen*, which is a generic model proposed by [28]. The empirical model is a K-profile parameterization (KPP) by [29] referred to as the LMD scheme [15].

Previous studies have investigated differences between turbulent closure models. A two-dimensional wind-driven model was used to compare the $k - \varepsilon$, MY25, and KPP turbulence closure schemes in an application that approximated the continental shelf [30]. The study found that these models produced similar mesoscale features. Although eddy diffusivity and eddy viscosity coefficients were qualitatively similar between the models, the turbulent structure and the intensities of mixing were not the same.

The MY25 model and a modified KPP were compared in wind-driven one- and two-dimensional shallow ocean conditions [31]. It was found that MY25 caused deeper mixing and higher entrainment than KPP in a highly stratified water column, but MY25 produced relatively less mixing in weakly stratified conditions. This characteristic was ascribed to downward diffusion of turbulent kinetic energy in MY25 in the presence of a strong pycnocline, which is not represented in the KPP model. Relatively stronger mixing of the KPP model lead to faster disintegration of the pycnocline where surface and bottom layers impinged on each other.

Model simulations of the Chesapeake Bay estuary revealed decreased model accuracy under strong vertical stratification, which was attributed to the choice of turbulence parameterization [32]. Four turbulence closure schemes, namely $k - \varepsilon$, $k - \omega$, $k - kl$, and KPP, were compared and shown to produce very similar results. In

each case the vertical stratification of the water column away from boundary layers depended strongly on the background diffusivity value. These vertical mixing parameterizations had difficulty representing a strong pycnocline.

Two-equation schemes were compared in a study that employed three different idealized three-dimensional model scenarios, approximating (i) steady barotropic flow in a rectangular channel, (ii) wind-induced surface mixed-layer deepening in a stratified water column, and (iii) oscillatory estuarine circulation in a rectangular channel [15]. The turbulence closure models tested included $k - \varepsilon$, $k - \omega$, $k - kl$, MY25, and *gen*. The study found that the $k - \varepsilon$, $k - \omega$, and *gen* models performed similarly in cases (i) – (iii). The MY25 scheme did not produce a well-defined surface mixed layer like the other schemes in case (iii), which caused a reduced salinity intrusion. The MY25 model produced relatively less mixing and lower bottom stress in case (i).

Studies of the Red Sea overflow [33] and Columbia River plume [16], which compared a number of two-equation and empirical models, showed that the $k - \varepsilon$ scheme was the most reliable at representing observed mixing and hydrographic processes. In the case of the Red Sea overflow the eddy diffusivities of the KPP model were lower than the observations, whereas those of the MY25 model were higher than observed and caused excessive mixing in the bottom layer of the overflow. The highest overall deviations from observations were seen for the KPP model and the MY25 model. In the Columbia River plume study the $k - \varepsilon$ scheme provided improvements over the MY25 and LMD schemes.

1.2 Method

One aspect of this work is a validation of the NB-ROMS model, a three-dimensional numerical model that represents the hydrodynamics of Narragansett Bay. An important parameterization determining the accuracy of models is the

turbulence closure scheme specification. The NB-ROMS model incorporates a turbulence closure scheme that was chosen for its success in representing Narragansett Bay processes. The procedure used to determine the suitability of a given turbulence closure scheme is another aspect of the work. Model performance is assessed for different turbulence closure scheme scenarios. Model scenarios differ only in the applied turbulence closure scheme. The performance assessment and validation process involve comparisons of model output at specific locations to observational time series of temperature, salinity, velocity, and sea surface elevation at the same locations. The discussion on methods starts with information about the observational time series (Section 1.2.1) and the model configuration (Section 1.2.2). Detail about the turbulence closure schemes is provided in Section 1.2.3. Data analysis techniques and the skill assessment method are explained in Section 1.2.4.

1.2.1 Observations

Observational monitoring sites were predominantly situated within the northern half of Narragansett Bay, which is commonly referred to as the upper Bay (Figure 1.1). Information regarding observations, namely the site names, locations, and depths, as well as data types and sampling periods used for model validation can be found in Table 1.1.

Velocity data were collected from three bottom-mounted Acoustic Doppler Current Profilers (ADCP) for a period in the summer of 2006 [7]. The work was funded by Rhode Island Sea Grant Omnibus 2006-08, grant -R/P-061, and by the Narragansett Bay Commission, grant 330286. The instruments were deployed along a latitude coincident with the northern point of Prudence Island at the following locations: (i) in the West Passage channel (WP) at 21 m depth, (ii) on the relatively shallow shoal of the East Passage (EPs) at 6 m depth, and (iii)

Table 1.1. Summary of observations including the station names (Sta), type of data that were collected at a site (Data), the sampled period in 2006 that is used for model validation (Tme), the geographical coordinates of each site (Pos), and the approximate site depth at mean lower-low water (D).

Sta	Data	Tme (GMT)	Pos (°)	D (m)
WPc	$u(t, z); v(t, z)$	20 Jun – 16 Aug	41.6645; -71.3715	21
EPc	$u(t, z); v(t, z)$	20 Jun – 16 Aug	41.6669; -71.3117	13
EPs	$u(t, z); v(t, z)$	3 Aug – 16 Aug	41.6665; -71.3358	6.0
BR	$T(t, z_s); S(t, z_s)$ $T(t, z_b); S(t, z_b)$	1 Jun – 12 Jul 5 Jun – 16 Aug	41.7406; -71.3714	6.0
CP	$T(t, z_s); S(t, z_s)$ $T(t, z_b); S(t, z_b)$	30 May – 16 Aug 15 Jun – 16 Aug	41.7138; -71.3450	7.0
NP	$T(t, z_s); S(t, z_s)$ $T(t, z_b); S(t, z_b)$	07 Jul – 30 Jul 07 Jul – 30 Jul	41.6704; -71.3547	11
MV	$T(t, z_s); S(t, z_s)$ $T(t, z_b); S(t, z_b)$	30 May – 16 Aug 30 May – 16 Aug	41.6384; -71.3837	7.0
QP	$T(t, z_s); S(t, z_s)$ $T(t, z_b); S(t, z_b)$	30 May – 25 Jul 30 May – 11 Jul	41.5881; -71.3807	7.0
PP	$T(t, z_s); S(t, z_s)$ $T(t, z_b); S(t, z_b)$	30 May – 16 Aug 30 May – 16 Aug	41.6634; -71.3161	8.0
TW	$T(t, z_s); S(t, z_s)$ $T(t, z_b); S(t, z_b)$	30 May – 16 Aug 31 May – 16 Aug	41.5789; -71.3219	6.0
GB	$T(t, z_s); S(t, z_s)$ $T(t, z_b); S(t, z_b)$	06 Jun – 16 Aug 30 May – 16 Aug	41.6826; -71.4401	3.0
MH	$T(t, z_s); S(t, z_s)$ $T(t, z_b); S(t, z_b)$	30 May – 16 Aug 30 May – 16 Aug	41.6801; -71.2152	5.0
WNp	$T(t, z); S(t, z)$	27 Jun – 14 Jul	41.6709; -71.3736	3.7
RSp	$T(t, z); S(t, z)$	27 Jun – 14 Jul	41.6921; -71.2946	4.0
BRp	$T(t, z); S(t, z)$	02 Aug – 16 Aug	41.7333; -71.3735	4.0
RNp	$T(t, z); S(t, z)$	02 Aug – 16 Aug	41.7180; -71.3224	4.3
P	$\zeta(t)$	30 May – 16 Aug	41.8085; -71.4000	
C	$\zeta(t)$	30 May – 16 Aug	41.7166; -71.3433	
Q	$\zeta(t)$	30 May – 16 Aug	41.5850; -71.4083	
N	$\zeta(t)$	30 May – 16 Aug	41.5050; -71.3267

Figure 1.1. Narragansett Bay bathymetry and locations of the following stations: ADCP's (stars) – West Passage (WP), East Passage Shoal (EPs), and East Passage Channel (EPc); NBFSMN buoys with temperature and salinity sondes (diamonds) – Bullock Reach (BR), Conimicut Point (CP), North Prudence (NP), Mount View (MV), Quonset Point (QP), Poppasquash Point (PP), T-Wharf (TW), Greenwich Bay Marina (GB), and Mount Hope Bay (MH); temperature and salinity profilers – Bullock Reach (BRp), Rumstick North (RNp), Rumstick South (RSp), and Warwick Neck (WNp); and tide gauges (squares) – Providence (P), Conimicut Light (C), Quonset Point (Q), and Newport (N). (Bathymetry data from <http://www.geomapapp.org>)

in the East Passage channel (EPc) at 13 m depth (Figure 1.1). Sampling at the West Passage Channel and the East Passage Channel stations were conducted at one-meter depth bins and ten-minute intervals. At the East Passage Shoal station sampling was specified at 0.5-meter depth bins and ten-minute intervals.

Temperature and salinity observations were of two types. Firstly, time series of temperature and salinity were collected near the surface and bottom at nine fixed monitoring sites. Eight of these stations were located near channels in water depths ranging between 5 m and 8.5 m at mean lower-low water. Mean lower-low water is defined as the average of the lower low water height of each tidal day observed between 1983 and 2001 [34]. Secondly, vertical profile time series of temperature and salinity were measured at four shallow, near-shore sites where depths ranged between 3.7 m and 4.3 m at mean lower-low water.

The first set of temperature and salinity observations were obtained from the Narragansett Bay Fixed-Site Monitoring Network (NBFSMN), which is maintained through collaborations between the Rhode Island Department of Environmental Management Water Resources Division, the Narragansett Bay National Estuarine Research Reserve, the Narragansett Bay Commission, the University of Rhode Island Graduate School of Oceanography, Roger Williams University, and the Bay Window Program [35, 36, 37]. At each location (Figure 1.1) the temperature and salinity are measured by a near-surface and near-bottom YSI 6000-series sonde at fifteen-minute intervals. The near-surface and near-bottom sondes are situated within the upper one meter and the lower one meter of the water column, respectively. *In situ* density was calculated from a nonlinear equation of state for seawater [38]. Temperature and salinity stratification were estimated from the difference between surface and bottom tracer time series.

The second set of temperature and salinity data were obtained from moored

SeaBird CTD profilers. The work was funded by the NOAA Coastal Hypoxia Research Program grant NA05NOS4781201. The profilers incorporated a bottom-mounted winch that alternately released and retrieved a buoyant sensor package with a YSI 6000-series sonde. The sensor profiled the water column from within 0.5 m or less of the seafloor to within 0.5 m or less of the sea surface. The duration of each profile measurement, which started and ended at the bottom, was five to ten minutes. Profile measurements commenced every three hours and the average vertical distance between samples was 14 cm. Upward and downward casts of each profile were averaged to obtain a measure of the vertical temperature and salinity structure.

Tide gauge information from the National Oceanic and Atmospheric Administration (NOAA) was acquired at the Providence, Conanicut Light, Quonset Point, and Newport locations [7] (Figure 1.1). Wind velocity and air temperature data were obtained from the Physical Oceanographic Real Time System (NOAA PORTS) [34]. Observations of river transport were accessed from the national water information system of the U.S. Geological Survey (USGS) [39]. Rainfall data at T.F. Green Airport were retrieved from the Weather Underground [40].

1.2.2 Model configuration

The Narragansett Bay numerical model comprised a fine scale grid to calculate circulation and transport at high spatial resolutions. Modeling work was funded by the NOAA Coastal Hypoxia Research Program grant NA05NOS4781201 [10]. The grid encompassed the entire coast of Narragansett Bay with highest horizontal resolution in the Providence River (~ 30 m) and lowest horizontal resolution in Mount Hope Bay (~ 240 m) (Figure 1.2). The open boundary was at the mouth of Narragansett Bay along a line passing 0.5 km north of Whale Rock to Narragansett in the west and to Sakonnet Point in the east (Figure 1.2). The model contained

15 terrain-following vertical levels, including a free surface layer, which produced variable spacing, from 0.13 m in the shallowest regions to 3 m in the deepest channels relative to a horizontal surface. Model bathymetry varied between 2 m in parts of Greenwich Bay, Providence River, and Mount Hope Bay to 45 m in the southern East Passage. A minimum depth of 2 m was defined to avoid inclusion of wetting and drying effects.

The model was set up to represent Narragansett Bay processes at high temporal resolution. A baroclinic time step of 24 s and a barotropic time step of 12 s were specified. Model experiments were run for a summer period from 31 May 2006 to 16 August 2006.

External forcing functions applied in the Narragansett Bay model included wind, tide, river, and atmospheric variables required for bulk formulations [41]. Local wind, air temperature, and barometric pressure observations at six locations around Narragansett Bay were obtained from the Physical Oceanographic Real Time System (NOAA PORTS) [34]. The locations were Fall River, Conimicut Light, Newport, Providence, Potter Cove, and Quonset Point. These observations were spatially averaged and applied across the grid at three-hour intervals. Daily volume transport of the eight largest rivers discharging into Narragansett Bay was obtained from the U.S. Geological Survey [39]. The rivers are the Woonasquatucket and Moshassuck, Blackstone and Ten Mile, Pawtuxet, Palmer, Taunton, and Hunt. Correction factors (M. Brush 2009, pers. comm.) were applied to account for the ungauged sections of each watershed. The freshwater sources included transport from three wastewater treatment facilities (WWTF), namely Bucklin Point, Fields Point, and East Providence (Figure 1.2). Observations of precipitation and relative humidity were obtained from the T.F. Green Airport. Hindcast information of net shortwave radiation and incoming longwave radiation were acquired from NCEP

Figure 1.2. Narragansett Bay model grid (black dots are the rho-points) and bathymetry. Locations of freshwater input in the model are indicated by arrows. The real coastline is shown in grey, along with important landmarks that are mentioned in the text.

Reanalysis of the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, from their Web site at <http://www.esrl.noaa.gov/psd/> [42].

The southern boundary of the Narragansett Bay model was forced by output of hourly temperature, salinity, surface elevation, and current velocity from a coarser ROMS model that extended onto the continental shelf [7]. The eight largest tidal constituents were obtained from the Eastcoast Tidal Constituent Database [43] for the larger domain model. Implicit upstream radiation conditions were specified at the southern boundary of the model. A 30-minute nudging timescale was prescribed for inflowing tracers and momentum, meaning that velocity, temperature, and salinity return to prescribed values over this time period. The nudging timescale for outflowing properties was one hour.

A number of options regarding the computation of momentum in the Narragansett Bay model were specified. They include an upstream-biased horizontal advection and conservative, parabolic spline vertical advection. In addition, the following was specified: Coriolis force, a parabolic spline density Jacobian pressure gradient algorithm [44], harmonic horizontal mixing, and logarithmic bottom friction. Mixing of horizontal momentum was calculated along geopotential surfaces with viscosity coefficients scaled by the grid size. A value of $2.0 \text{ m}^2/\text{s}$ was assigned to the lateral harmonic viscosity coefficient of the largest grid cell in the domain. Vertical mixing of momentum is described in Section 1.2.3.

Specifications for the calculation of tracer equations were assigned in the Narragansett Bay model. Tracers were computed through a three-dimensional positive definite advection transport algorithm. Harmonic horizontal mixing of tracers was calculated along geopotential surfaces with diffusion coefficients that were scaled by the grid size. The lateral harmonic diffusion coefficient for the largest grid cell in the domain was assigned a value of $1.5 \text{ m}^2/\text{s}$. Vertical mixing of tracers is

described in Section 1.2.3.

1.2.3 Turbulence closure specifications

Vertical mixing parameterizations that are fully coded within ROMS (Table 1.2) were used to characterize how various options impact model accuracy. Closure schemes fall into two general categories. Five of the schemes were statistical two-equation models, namely $k-\varepsilon$ [20, 21, 22, 23], MY25 [24], $k-\omega$ [25, 26, 27, 28], $k-kl$ [24, 28], and *gen* [28]. One scheme was an empirical model, called the K-profile parameterization or LMD scheme. Both the statistical and empirical models solve the Reynolds averaged Navier Stokes (RANS) equations. The difference between the two types of models is in how the Reynolds fluxes are solved. The statistical approach to turbulence closure approximates these fluxes in terms of a Friedman-Keller series [11, 12]. For this subset of cases we tested the G88 and KC stability functions (Table 1.2), described by Galperin *et al.* [45] and Kantha and Clayson [46]. Stability functions are derived algebraically from transport equations of the Reynolds stresses and occur in expressions that relate turbulent kinetic energy and length scale to viscosity and diffusivity coefficients. The empirical approach to turbulence closure employs empirical knowledge regarding fluxes in the geophysical boundary layers.

The theory of statistical two-equation and empirical models is explained in detail in Appendix A.

1.2.4 Data analysis techniques

Processing and analysis methods used to compare different aspects of the data included filtering techniques, harmonic analysis, Empirical Orthogonal Function (EOF) analyses, and skill tests. Time series were filtered to compare the tidal and subtidal components of modeled and observed properties. Here "tidal" refers

Table 1.2. Summary of turbulence closure scheme scenarios that distinguish the different model runs.

Turbulence closure	Specifications
$k - \varepsilon$	G88 stability function
MY25	G88 stability function
LMD	KPP surface and bottom boundary layer mixing Convective mixing due to shear instability Convective nonlocal transport Diffusivity due to shear instability
$k - \omega$	G88 stability function
$k - kl$	G88 stability function Parabolic wall function with free surface correction
<i>gen</i>	G88 stability function
$k - \omega$	KC stability function
$k - kl$	KC stability function
$k - \varepsilon$	KC stability function
<i>gen</i>	KC stability function

to the 2.5–33 hour bandpass-filtered time series record and "subtidal" refers to the 33-hour lowpass-filtered time series record. The MATLAB [47, 48] functions 'filtfilt' and 'butter' were used to construct filters. Harmonic analyses of properties were conducted using the MATLAB toolbox T-TIDE [49].

Empirical Orthogonal Function (EOF) analyses further facilitated model-data comparisons [50, 51]. MATLAB coding [50] was used to conduct EOF analyses on velocity data. EOF analysis entails the decomposition of a spatial time series, such as profiles of horizontal velocity collected by an ADCP, into a linear combination of orthogonal spatial modes. It is a method for partitioning the variance of spatial time series. The EOFs are ordered by decreasing eigenvalue so that the first few modes with largest eigenvalues contain the majority of the variance of the data. By extracting the patterns of these functions the dynamical processes involved may become apparent. The orthogonal functions are defined by the covariance structure of the spatial time series.

For a total of N observations collected at times $t = t_i$ ($1 \leq i \leq N$), and a total of M spatial points collected at locations, $\vec{x}_m$ ($1 \leq m \leq M$), the goal of the EOF procedure is to write a data series $\psi_m(t)$ at a location $\vec{x}_m$ as the sum of M orthogonal spatial functions $\phi_i(\vec{x}_m) = \phi_{im}$ such that

$$\psi(\vec{x}_m, t) = \psi_m(t) = \sum_{i=1}^M [a_i(t)\phi_{im}], \quad (1.1)$$

where $a_i(t)$ is the amplitude of the i th orthogonal mode [51]. The time variation of a dependent scalar variable at each location is the linear combination of M spatial functions, ϕ_i , whose amplitudes are weighted by M time-dependent coefficients, $a_i(t)$, which characterizes the temporal variability of the modes [51]. The orthogonality requirement is

$$\sum_{m=1}^M [\phi_{im}\phi_{jm}] = \delta_{ij}, \quad (1.2)$$

where δ_{ij} is the Kronecker delta. Another constraint is that the amplitudes $a_i(t)$ are uncorrelated over the sample data such that

$$\overline{a_i(t)a_j(t)} = \lambda_i\delta_{ij} = \overline{a_i(t)^2}\delta_{ij}, \quad (1.3)$$

where overbar denotes a time averaged value and $\lambda_i = \overline{a_i(t)^2}$ is the variance in each orthogonal mode [51]. Forming a covariance matrix with the time series as $\psi_m(t)\psi_k(t)$ the canonical form of the eigenvalue problem is derived for the i th mode at the m th location [51]:

$$\sum_{k=1}^M \overline{\psi_m(t)\psi_k(t)}\phi_{ik} = \lambda_i\phi_{im}. \quad (1.4)$$

In basic terms the procedure of solving for the EOFs of a single field is to (i) form a matrix of the observations and remove the temporal mean; (ii) construct the covariance matrix; (iii) find the eigenvalues and eigenvectors of the covariance matrix; (iv) find the highest eigenvalues and the corresponding eigenvectors (EOFs); (v) find the time-dependent amplitudes of each mode of the EOF [50, 51].

Model performance was quantified using a Willmott skill [18, 19, 16]. The Willmott skill (WS) was calculated as

$$WS = 1 - \frac{\overline{(m - o)^2}}{(\overline{|m - \bar{o}|} + \overline{|o - \bar{o}|})^2} \quad , \quad (1.5)$$

where m is the modeled time series, o is the observed time series. Note that the nominator in the second term, $\overline{(m - o)^2}$, is the mean squared error. The mean squared error comprises the (i) mean bias, (ii) standard deviation error, and (iii) cross-correlation coefficient error. A WS value of one indicates perfect agreement between modeled and observed time series and a WS value of zero indicates that there is no agreement between modeled and observed time series. WS values of different physical variables can be compared due to the nondimensionality of the Willmott skill [18, 16].

The behavior of the Willmott skill can be demonstrated by comparing simple harmonic functions of different forms (Figure 1.3). A comparison of the signals in Figure 1.3 (a1) and (a2) shows that the WS value increases from 0.55 to 0.89 when the RMSE decreases. This is due to an improved correspondence between signal amplitudes in Figure 1.3 (a2) compared to Figure 1.3 (a1). Decreasing the difference in signal frequency (compare Figure 1.3 (b1) and (b2)) causes CC and $RMSE$ to remain the same and does not improve the WS value. The signals of Figure 1.3 (c1) and (c2) differ in the magnitude of phase discrepancies. In this example the WS value increases from 0.58 to 0.90 with improved agreement in phase and consequent increases in CC and $RMSE$. Finally, when the mean offset between two identical signals is decreased from five to two, the WS value increases from 0.23 to 0.40 (compare Figure 1.3 (d1) and (d2)).

The length of time series used in the skill assessment is an important factor in dictating final outcomes. Note that the skill assessments are conducted with the maximum number of data points available for a particular variable at any given

Figure 1.3. Simple harmonic functions show the characteristics of the Willmott skill (WS) with relation to the correlation coefficient (CC) and the root mean square error ($RMSE$). The differences between the grey and black signals of each specified panel are as follows: (a1) amplitudes vary by a factor of five; (b1) periods vary by a factor of five; (c1) phases differ by 72° ; (d1) the mean values differ by five; (a2) amplitudes vary by a factor of two; (b2) periods vary by a factor of two; (c2) phases differ by 36° ; and (d2) the mean values differ by two.

station. It is necessary to identify the time periods that correspond to the skill values of tracers, velocity, and sea surface elevation at each station.

Time series of surface and bottom temperature and salinity were obtained from a buoy network, which lead to a wide range in time periods when data were available at the different sites (Figure 1.4). Tracer time series that were relatively short included the surface and bottom properties at the North Prudence (23 days) and Quonset Point stations (56 days and 42 days, respectively), as well as the surface tracer time series at the Bullock Reach station (41 days) and bottom time series at the Conimicut Point station (62 days). All other tracer time series had lengths that ranged between 70 days and 78 days.

Figure 1.4. Time periods used in skill assessment of tracers at each of the nine fixed-site stations where, for example, BRs and BRb are the surface and bottom tracer time series at the Bullock Reach station, respectively. Red indicates the duration of surface time series and blue indicates the duration of bottom time series.

ADCP records were from deployments by [7] and had the benefit of coincident

surface and bottom measurements (Figure 1.5). Model output was compared to an observed time series of 57 days at the West Passage and East Passage Channel stations. At the East Passage Shoal station the data record used for model comparisons was 13 days long. The lengths of the four tide gauge records were 78 days.

Figure 1.5. Time periods used in skill assessment of velocity at the three ADCP stations and sea surface elevation at the four tide gauge stations. The tide gauge stations are Providence (P), Conimicut Light (C), Quonset Point (Q), and Newport (N).

In the analysis a distinction was made between barotropic and baroclinic velocity, which were derived from the ADCP profile time series. The barotropic velocity was obtained by averaging the velocity profile at each instant in time. Baroclinic velocity was determined by subtracting the barotropic component from the velocity profile at each instant in time.

Data sets are denoted 'surface' and 'bottom' series to facilitate discussion, but data used in the skill assessment were collected some distance away from the surface

and bottom layers, and the actual data depths depended on the specific variable. Temperature and salinity data were collected within one meter of the surface and the bottom. The depths of velocity data used in the skill assessment is shown in Figure 1.6. Surface and bottom velocity assessments were representations of upper and lower water column characteristics. A number of upper (lower) vertical velocity layers were averaged together to be designated surface (bottom) records. The layers that were averaged together were selected based on the criteria that they were consecutive within the near-surface (near-bottom) and that the resulting instantaneous data-model skill was optimal compared to other near-surface (near-bottom) layer combinations. The number of bins averaged together to represent the upper water column were two for east-west velocity at the West Passage station, and the north-south velocity at the West Passage, East Passage Channel, and East Passage Shoal stations. Two bins were averaged together to represent the lower water column, east-west velocity at the West Passage and the East Passage Shoal sites, and the lower water column, north-south velocity at the East Passage Shoal site. Three bins were averaged together to get upper water column, east-west flow at the East Passage Shoal station. Only one bin was used to represent surface and bottom east-west flow at the East Passage Channel station, and bottom north-south flow at the West Passage and East Passage Channel stations.

1.3 Results and discussion

The model validation involves (i) an assessment of the performance of the NB-ROMS model that implements a successful turbulence closure scheme (Section 1.3.2), and (ii) a comparison of the Narragansett Bay model output when different turbulence closure schemes are specified (Section 1.3.3). The accuracy of the NB-ROMS model and the different turbulence closure scheme scenarios is determined by a skill assessment (Section 1.3.1).

Figure 1.6. Depths of discrete vertical velocity layers that are averaged together to obtain surface (upper boxes) and bottom (lower boxes) velocity time series for data-model comparisons at the three ADCP stations (WP, EPc, and EPs) and for the east-west (u) and north-south (v) velocity components.

1.3.1 Skill assessment

The applicability of the statistical two-equation and empirical models listed in Table 1.2 is examined by considering their accuracy in calculating observed tidal and subtidal tracers, velocity, and sea surface elevation. The Willmott skill (WS) [18, 19, 16] is used to quantify the performance of these models. The maximum WS value of one indicates perfect agreement between modeled and observed properties and a minimum value of zero denotes complete disagreement between a model and the observations [18]. Skill values of 0.65 or higher are considered to indicate a good match between the observed and modeled time series, and skill values below 0.60 show that the model performs poorly.

The accuracy of the different models can be compared comprehensively by averaging skill values over all variables and stations (Figure 1.7). The comparison shows that all two-equation models provide higher skills for subtidal properties than tidal properties. The LMD model performs better at modeling tidal characteristics than subtidal characteristics. The different two-equation models perform very similarly at both tidal and subtidal time scales. The LMD model provides lower accuracy than the two-equation models.

The skill assessment can be used to explore model performance in more detail. The Willmott skills are first averaged across all stations for each variable. These spatially averaged skills provide an indication of the ability of models to reproduce specific properties of upper Narragansett Bay as a whole. Secondly, the skill assessment is considered in a local sense by comparing model performance for each variable at separate stations. This approach highlights spatial trends regarding model performance. A summary of the most important skill findings is presented at the end of this section. The details that describe the ways in which the NB-ROMS model accurately or inaccurately represents hydrodynamic processes are

Figure 1.7. Comparison of the average Willmott skill for tidal (blue) and subtidal (red) properties generated by the different turbulence closure models.

discussed in Section 1.3.2 and the differences in the representation of the dynamics by varying turbulence closure implementations are discussed in Section 1.3.3.

Two-equation model skills for tidal time series

The relative ability of statistical two-equation models to reproduce tidal time series are discussed in this section. It is shown in this section that tidal property skills are generally similar for all turbulence closure implementations. The GLS models offer best skill values overall and skill values of the MY25 scheme deviate slightly from them in a few cases. All models are skillful in predicting tidal sea surface elevation and north-south velocity. In comparison the skills of tidal tracers and east-west velocity are somewhat lower for all models.

In the following assessment the performance of the models for specific variables are first considered in a spatially averaged sense and secondly in the context of spatial skill variability. In addition to focusing on the relative performance of

models that implement different turbulence closure schemes, specific attention is given to the skills of the $k - \varepsilon$ G88 implementation.

Spatial mean skill – tracers The spatially averaged WS values of tidal time series show variability in model performance between water properties (Figure 1.8). The spatially averaged WS values of tidal time series of tracers modeled through GLS schemes vary between 0.60 and 0.70. Very small differences between the GLS schemes and MY25 are seen for surface salinity, temperature, and density. The relative accuracy of the NB-ROMS model to calculate tidal tracers can be assessed by considering WS values of the $k - \varepsilon$ G88 scheme. Lowest skills are for tidal surface salinity and density, both with a value of 0.59. Highest WS of 0.70 is seen for tidal surface temperature.

Figure 1.8. Spatially averaged Willmott skill of tidal time series of surface salinity (S_s), bottom salinity (S_b), surface temperature (T_s), bottom temperature (T_b), surface density (ρ_s), bottom density (ρ_b), surface east-west velocity (u_s), bottom east-west velocity (u_b), surface north-south velocity (v_s), bottom north-south velocity (v_b), and sea surface elevation (z) for each of the turbulence closure experiments. Color scale: 0.40 – 1.0

Spatial mean skill – velocity Skills for the tidal north-south component of velocity are high, but those for the tidal east-west component of velocity are relatively low. All models produce identically high WS values of 0.86 for tidal north-south surface velocity and produce skills between 0.74 and 0.76 for tidal north-south bottom velocity. The models perform relatively poorly when calculating tidal east-west velocity, particularly at the bottom of the water column. Differences between models can be detected for surface east-west velocity skill values, but these differences are small. The highest surface east-west velocity skill value is 0.60 ($k - kl$ G88) and the lowest is 0.57. Skill values for bottom east-west velocity vary between 0.56 and 0.57 for all models.

Spatially varying skill – tracers The accuracy with which models are able to predict tidal tracers at individual stations can be assessed by averaging the tracer skills at each station and viewing the averages as a function of relative position along the longitudinal (lengthwise) axis of Narragansett Bay (Figure 1.9). In general the two-equation models perform similarly at reproducing tidal tracers at each location. The largest discrepancy between observed and modeled tidal tracers is seen at the Greenwich Bay station, where all model skills are less than 0.51. Bullock Reach tidal tracers are also modeled with relatively low accuracy, producing skills below 0.56. Skill values at the southern stations exceed 0.63. The MY25 scheme provides an improvement of 0.01 – 0.03 in the accuracy of modeled tidal tracers at the West Passage stations.

The $k - \varepsilon$ G88 model is used in a discussion of separate spatially varying tidal tracer skills, since it is representative of the other two-equation models (Figure 1.10). Locations where the skill values are low include Greenwich Bay Marina and Bullock Reach. Tidal surface and bottom salinity skills are extremely low at the Greenwich Bay Marina station ($WS \sim 0.20$) and the bottom temperature skill

Figure 1.9. Averaged tidal tracer skills produced by all models as a function of relative station locations in Narragansett Bay. Averages include skills for surface salinity, surface temperature, bottom salinity, and bottom temperature. Stations are arranged from north to south with Bullock Reach (BR) in the north and T-Wharf (TW) in the south. Note that Mount View (MV) and Quonset Point (QP) are West Passage stations, whereas Poppasquash Point (PP) and T-Wharf (TW) are East Passage stations. The Greenwich Bay (GB) and Mount Hope Bay (MH) stations are located away from the longitudinal axis of the Bay.

($WS = 0.50$) is well below the spatial average. At Bullock Reach all tidal tracer skills are below the spatial averages. Relatively low values include the tidal bottom temperature skill of 0.40 and the tidal surface and bottom salinity skills of 0.58 and 0.52, respectively. The Mount Hope Bay station skills are below average for tidal bottom salinity, and surface and bottom temperature. Bottom temperature is calculated at a relatively low skill of 0.57. Tidal surface and bottom temperature skills are relatively low at North Prudence ($WS = 0.58$ and 0.55 , respectively). Tidal surface salinity skills are relatively low at Poppasquash Point and T-Wharf ($WS = 0.54$ and 0.56 , respectively). At the T-Wharf station the tidal surface temperature skill is slightly below the spatial average.

Skills for separate tidal tracers can be organized based on relative station locations from north to south (Figure 1.11). No clear north-south trends can be perceived for tidal tracer skills obtained from the two-equation models. It is noted that all tidal tracers are modeled adequately ($WS > 0.65$) at Conimicut Point, Mount View, and Quonset Point, but some variability exists in the tidal tracer skills at other stations. The low tidal tracer skill at Greenwich Bay is largely due to an inability of the model to capture surface and bottom salinity at this location.

Tidal density skills of the $k - \varepsilon$ G88 model are considered as an additional test of the ability of models to capture Narragansett Bay hydrography (Figure 1.10). The skills for tidal surface (/bottom) density closely resemble those of tidal surface (/bottom) salinity. This indicates that density is largely a function of salinity in Narragansett Bay.

Skills for tidal stratification indicate the capability of models to reproduce the vertical structure of hydrography in Narragansett Bay (Figure 1.10). The surface and bottom tracer time series are subtracted from each other to obtain a proxy for stratification. It should be noted that tidal density stratification skills

Figure 1.10. Willmott skill for tidal surface and bottom salinity, temperature, and density, along with salinity and temperature stratification for each of the turbulence closure experiments at nine stations.

Figure 1.11. Skills for tidal surface salinity (solid red), surface temperature (solid blue), bottom salinity (dashed red), and bottom temperature (dashed blue) are shown. The panels show skills for the $k - \epsilon$ G88 (top), MY25 (middle), and LMD (bottom) models. Stations are arranged from north to south with Bullock Reach (BR) in the north and T-Wharf (TW) in the south. Note that Mount View (MV) and Quonset Point (QP) are West Passage stations, whereas Poppasquash Point (PP) and T-Wharf (TW) are East Passage stations. The Greenwich Bay (GB) and Mount Hope Bay (MH) stations are located away from the longitudinal axis of the Bay.

closely resemble those of tidal salinity stratification, which is in agreement with the finding that density predominantly depends on salinity in Narragansett Bay. Skills of the $k - \varepsilon$ G88 show that tidal stratification is appropriately represented at the Conimicut Point and Quonset Point stations ($WS > 0.67$), but less successfully ($WS < 0.60$) at Bullock Reach, T-Wharf, and Greenwich Bay Marina.

Spatially varying skill – velocity Tidal velocity is reproduced with varying degrees of success at the three ADCP stations (Figure 1.12). Slight differences can be detected in the velocity skills of separate models. Relatively high skills for tidal surface and bottom north-south velocity components are seen at all stations. WS values exceed 0.80 for the surface north-south velocity. Models further produce high accuracy for the tidal surface ($WS = 0.76$) and bottom ($WS = 0.75$) east-west velocity component at the West Passage station. In contrast, the model output of east-west velocity at the surface and bottom of the East Passage Channel and East Passage Shoal stations does not compare well with observations; WS values range between 0.34 and 0.61.

Two-equation model skills for subtidal time series

The relative accuracy of the two-equation models can be assessed for subtidal properties. In this section it is shown that all turbulence closure implementations perform similarly at the subtidal time scales. Different subtidal variables are modeled with different levels of accuracy. In general the skill values of subtidal properties are high, particularly for subtidal tracers. It will be shown that at the subtidal time scale the models perform poorly in computing sea surface elevation and east-west velocity.

Spatial mean skill – tracers Spatially averaged WS values of subtidal time series show that the models capture properties at varying degrees of success

Figure 1.12. Willmott skill of tidal surface and bottom velocity for each of the turbulence closure experiments at three ADCP stations. Note: u-velocity is the east-west component of velocity and v-velocity is the north-south component of velocity. Color scale: 0.40 – 1.0

(Figure 1.13). Skills are high for tracers. Values are highest for subtidal surface temperature at about 0.95 for the GLS and MY25 schemes. Subtidal surface salinity skills are about 0.87 for the GLS and MY25 schemes. Subtidal bottom salinity ranges from 0.74 for the MY25 scheme and 0.75 – 0.77 for the GLS schemes. Subtidal bottom temperature *WS* skills vary between 0.78 (*k – kl* G88) and 0.80. Skills of subtidal surface density of all models are 0.88 and bottom density are 0.77 – 0.78. The *k – ε* G88 predicts subtidal surface temperature, salinity, and thus density successfully. Subtidal bottom salinity, temperature and density are reproduced at slightly lower skills, but *WS* values exceed 0.70.

Figure 1.13. Spatially averaged Willmott skill of 33-hour low pass filtered time series of surface salinity (S_s), bottom salinity (S_b), surface temperature (T_s), bottom temperature (T_b), surface density (rho_s), bottom density (rho_b), surface east-west velocity (u_s), bottom east-west velocity (u_b), surface north-south velocity (v_s), bottom north-south velocity (v_b), and sea surface elevation (z) for each of the turbulence closure experiments. Color scale: 0.40 – 1.0

Spatial mean skill – velocity Skill values of subtidal velocity are similar for all turbulence closure models. Some variability exists in the capacity of the models to calculate the east-west component of subtidal surface velocity. *WS* values are 0.54 – 0.55 for *gen* KC, $k - \varepsilon$ KC, and $k - \omega$ KC, 0.61 for the $k - kl$ G88 scheme, and 0.55 – 0.59 for the other two-equation schemes. The subtidal east-west bottom velocity *WS* values vary between 0.58 and 0.60 for these models. The subtidal north-south component of velocity is represented well by all models. Skill values of surface north-south velocity range between 0.73 and 0.75, and skill values of bottom north-south velocity are between 0.77 and 0.79. The $k - \varepsilon$ G88 model is representative of the other GLS models at the subtidal velocity scales and lead to high skill in computing north-south velocity. It is less successful in accurately predicting east-west subtidal velocity.

Spatially varying skill – tracers The accuracy of models to predict subtidal tracers at individual stations can be assessed by averaging the subtidal skills at each station and viewing the averages as a function of relative position from north to south in Narragansett Bay (Figure 1.14). The two-equation models perform very similarly at reproducing subtidal tracers at each location. Small differences in the performance of models can be detected at the North Prudence station where skills range from 0.80 to 0.84. Skill values at stations south of North Prudence are somewhat lower than stations in the north, particularly at Quonset Point. However, model accuracy can still be considered high at these southern stations, since skill values exceed 0.70.

The spatial variability of skills is further assessed by determining skills for the different subtidal tracers. Two-equation model skills vary markedly from one location to the next, but these models generally exhibit high accuracy (Figure 1.15). Since the differences are minor between the GLS models, we use the $k - \varepsilon$ G88

Figure 1.14. Averaged subtidal tracer skills produced by all models as a function of relative station locations in Narragansett Bay. Averages include skills for surface salinity, surface temperature, bottom salinity, and bottom temperature. Stations are arranged from north to south with Bullock Reach (BR) in the north and T-Wharf (TW) in the south. Note that Mount View (MV) and Quonset Point (QP) are West Passage stations, whereas Poppasquash Point (PP) and T-Wharf (TW) are East Passage stations. The Greenwich Bay (GB) and Mount Hope Bay (MH) stations are located away from the longitudinal axis of the Bay.

model as a representative parameterization to compare skills for different tracers. Locations where subtidal skill values are lower than the spatially averaged skill are predominantly the southern stations. These include subtidal bottom salinity at Poppasquash Point ($WS = 0.60$), Quonset Point ($WS = 0.54$), and T-Wharf ($WS = 0.68$), as well as subtidal bottom temperature at North Prudence ($WS = 0.68$), Poppasquash Point ($WS = 0.69$), and Quonset Point ($WS = 0.59$). Of these time series only three can be considered relatively poorly modeled ($WS < 0.60$).

Skills of separate subtidal tracers can be organized based on relative station locations from north to south (Figure 1.16). It is seen that accuracy of subtidal bottom salinity becomes relatively low southward from North Prudence, notably along the West Passage. Subtidal surface salinity skills show a minor southward decrease. Along the West Passage the decrease is from 0.96 at the Conimicut Point station to 0.75 at Quonset Point. Subtidal tracer skills are consistently high ($WS \geq 0.88$) at Bullock Reach, Greenwich Bay Marina, and Mount Hope Bay.

The ability of the $k - \varepsilon$ G88 model to capture Narragansett Bay hydrography can further be assessed by considering the subtidal density skills (Figure 1.15). Skills of subtidal surface density resemble skills of subtidal surface salinity. This indicates that density is more closely related to salinity than temperature at subtidal scales. Surface subtidal density WS values exceed 0.88 at stations not including Quonset Point ($WS = 0.82$), T-Wharf ($WS = 0.78$), and Greenwich Bay Marina ($WS = 0.80$). At stations where the subtidal bottom salinity and temperature are represented by the model with very high skill, the subtidal bottom density skill is similarly high ($WS > 0.80$). These stations include Bullock Reach and Conimicut Point in the north, as well as Mount View, Greenwich Bay Marina ($WS = 0.91$) and Mount Hope Bay ($WS = 0.92$). The subtidal bottom density WS values vary between 0.73 and 0.79 at the North Prudence and T-Wharf stations. Bottom

Figure 1.15. Willmott skill of subtidal surface and bottom salinity, temperature, and density, along with salinity and temperature stratification for each of the turbulence closure experiments at nine stations.

Figure 1.16. Skills for subtidal surface salinity (solid red), surface temperature (solid blue), bottom salinity (dashed red), and bottom temperature (dashed blue) are shown. The panels show skills for the $k - \epsilon$ G88 (top), MY25 (middle), and LMD (bottom) models. Stations are arranged from north to south with Bullock Reach (BR) in the north and T-Wharf (TW) in the south. Note that Mount View (MV) and Quonset Point (QP) are West Passage stations, whereas Poppasquash Point (PP) and T-Wharf (TW) are East Passage stations. The Greenwich Bay (GB) and Mount Hope Bay (MH) stations are located away from the longitudinal axis of the Bay.

subtidal density skills are 0.56 at Quonset Point and 0.61 at Poppasquash Point.

Skills for subtidal stratification indicate the capability of two-equation models to reproduce the vertical structure of hydrography in Narragansett Bay (Figure 1.15). The $k - \varepsilon$ G88 closure provides subtidal density stratification with similar accuracy than subtidal salinity stratification. Subtidal salinity stratification skills exceed 0.83 at all stations, except Quonset Point ($WS = 0.76$) and Greenwich Bay Marina ($WS = 0.43$). The Greenwich Bay Marina station produces a low skill value even though the subtidal surface and bottom salinity skills are high ($WS > 0.79$). High agreement between modeled and observed subtidal temperature and salinity stratification is seen at Bullock Reach, Conimicut Point, Mount View, and Mount Hope Bay. Subtidal temperature stratification skills are generally lower than those for subtidal salinity stratification. Skills for subtidal temperature stratification are low at T-Wharf and Greenwich Bay Marina (0.50 and 0.54, respectively), even though the subtidal surface and bottom temperature skills are high ($WS > 0.87$). Skill values of subtidal temperature stratification are below 0.65 at the North Prudence, Quonset Point, and Poppasquash Point stations.

Spatially varying skill – velocity Subtidal velocity is reproduced with varying degrees of success at the three ADCP stations (Figure 1.17). Variability among turbulence closure scheme implementations is seen for subtidal surface east-west velocity skills at the East Passage Shoal. For example, the $k - kl$ schemes, the $k - \omega$ G88 scheme, and the *gen* G88 scheme offer improvements over the $k - \varepsilon$ G88 scheme ($WS = 0.49$). Apart from these differences, the performance of the GLS and MY25 models are very similar. The $k - \varepsilon$ G88 scheme is effective in modeling subtidal surface and bottom north-south velocity at the West Passage and East Passage Shallows station ($WS > 0.70$). The model also reproduces the subtidal bottom north-south velocity at the East Passage Channel station very well (WS

$= 0.91$), but less so the subtidal surface north-south velocity at the same station ($WS = 0.65$). The WS value of subtidal surface east-west velocity is relatively low ($WS < 0.64$) at all stations. The subtidal bottom east-west velocity has a low skill value of 0.45 at the East Passage Channel station.

Figure 1.17. Willmott skill of subtidal surface and bottom velocity for each of the turbulence closure experiments at three ADCP stations. Note: u-velocity is the east-west component of velocity and v-velocity is the north-south component of velocity. Color scale: 0.40 – 1.0

Skills for tidal and subtidal sea surface elevation The performance of the two-equation models in reproducing sea surface elevation is different for tidal and subtidal time series. Tidal sea surface elevation is calculated equally well by all models, but modeled subtidal sea surface elevation is in poor agreement with the observations. All models produce spatially averaged WS values of 0.98 for tidal sea surface elevation (Figure 1.8). The spatially averaged WS values for subtidal

sea surface elevation vary between 0.36 and 0.43 for the different models (Figure 1.13). Tidal sea surface elevation skills exceed 0.97 at all tide gauges and subtidal sea surface elevation skills are 0.51, 0.44, 0.40, and 0.39 at Providence, Conimicut Light, Quonset Point, and Newport, respectively.

Skills of empirical model

The empirical LMD model produces properties with generally low skill. The model is as successful as the two-equation models in predicting tidal sea surface elevation and north-south velocity. Similar to those models the LMD scheme has limited skill in modeling tidal tracers and east-west velocity. Specifying the LMD scheme produces subtidal skill values that are generally lower than skills calculated for the two-equation models. The exceptions are subtidal surface temperature and sea surface elevation, which is modeled with similar efficiency as the two-equation models, and subtidal surface east-west velocity, which is modeled with highest accuracy by the LMD scheme.

The relative success of the LMD scheme in modeling tidal tracers can be considered in terms of spatially averaged skills (Figure 1.8). Tidal surface salinity, temperature, and density, as well as bottom temperature are modeled at slightly lower spatially averaged skill than the other models. Tidal bottom temperature modeled via the LMD scheme produces the lowest tidal tracer *WS* value of 0.55. Subtidal tracers are generally modeled at markedly lower accuracy by the LMD model (Figure 1.13). For example, the subtidal surface salinity skill is 0.87 – 0.88 for two-equation models, but 0.55 with the LMD scheme. The exception is subtidal surface temperature that is produced with similar high accuracy by the LMD model.

The spatially averaged tidal velocity skills of the LMD scheme are less than those of the two-equation models (Figure 1.8). The LMD implementation produces

a relatively low WS value of 0.68 for tidal north-south bottom velocity. The accuracy of the LMD scheme to model tidal north-south surface velocity is the same as the two-equation models with a WS value of 0.86. Similar to the other models, the LMD scheme performs poorly in reproducing tidal east-west velocity. For surface east-west velocity the LMD scheme is slightly lower at 0.56 than the highest two-equation skill value of 0.60. The LMD skill value for bottom east-west velocity is 0.51, which is somewhat lower than the lowest two-equation skill value of 0.56.

The spatially averaged skills of the LMD scheme for calculating subtidal velocity vary based on the velocity component (Figure 1.13). The LMD scheme performs better than two-equation models at computing subtidal east-west surface velocity, producing a WS value of 0.67 compared to the highest two-equation skill value of 0.61 for the $k - kl$ G88 scheme. Skill values for the other LMD subtidal velocity components are low compared to the two-equation models.

Spatial characteristics for average tidal and subtidal tracer skills reveal poor LMD performance compared to other models at the majority of the stations (Figures 1.11 and 1.16). The LMD model produces improved spatial subtidal velocity skills in three incidences (Figure 1.17). Two incidences include the subtidal surface east-west velocity at the West Passage ($WS = 0.72$) and East Passage Shoal stations ($WS = 0.83$). In the latter case, implementing the LMD scheme produces considerable improvement over the two-equation models. The other incidence is the skill for subtidal surface north-south velocity at the East Passage Channel station ($WS = 0.77$).

Summary of skill assessment

The skill assessment reveals several interesting aspects of the performance of the models with different turbulence closure scheme specifications. These charac-

teristics are explained with relation to hydrodynamic processes in further detail in Sections 1.3.2 and 1.3.3. It is valuable to summarize here the most important skill findings as a foundation for the continuing discussion.

In general there are very small differences between skills of models that implement the MY25 scheme and the GLS schemes. The $k - \varepsilon$ G88 implementation of the NB-ROMS model is representative of a set of models that provide highest overall skill. The model that specifies the LMD scheme provides output that are generally in poor agreement with the observations.

The only instances where other models perform markedly better than the $k - \varepsilon$ G88 model is when subtidal surface east-west velocity and subtidal surface north-south velocity are analyzed. These are also the only instances where implementing the LMD scheme provides a notable improvement compared to the GLS and MY25 schemes. However, the improvement offered by the LMD scheme only occurs for subtidal surface east-west velocity at the West Passage and East Passage Shoal stations and for subtidal surface north-south velocity at the East Passage Channel station. For most of the other velocity variables and stations the LMD scheme causes decreased skills compared to the other models. At the East Passage Shoal station the subtidal surface east-west velocity is also modeled with higher accuracy than the $k - \varepsilon$ G88 model by the $k - kl$ models, the $k - \omega$ G88 model, and the *gen* G88 model.

Output from the $k - \varepsilon$ G88 model agrees well with observations at tidal and subtidal time scales. The model produces highest agreement with observations of tidal and subtidal north-south velocity, tidal sea surface elevation, and subtidal tracers. Relatively lower model performance is seen for tidal and subtidal east-west velocity, tidal tracers, and subtidal sea surface elevation.

When the skills of all tracer stations are averaged together, the $k - \varepsilon$ G88

model generates relatively low agreement with observations for tidal tracers, except tidal surface temperature. At several individual stations, however, the model computes those tidal tracers at skill values exceeding 0.65 (Figure 1.18). Tidal surface salinity and density are modeled at WS values higher than 0.65 at Conimicut Point and the West Passage stations, although the spatially averaged skills are both 0.59. Spatially averaged tidal bottom salinity, temperature and density skills vary between 0.60 and 0.63, but skills for tidal bottom salinity exceed 0.65 at Conimicut Point, and the West Passage and East Passage stations. Tidal bottom density skills exceed 0.65 at Conimicut Point, and the West Passage and East Passage stations, but excluding North Prudence. Tidal bottom temperature skills exceed 0.65 at Conimicut Point, the East Passage and the West Passage, excluding Quonset Point. Tidal salinity and temperature stratification are adequately represented ($WS > 0.65$) at Conimicut Point and Quonset Point.

Figure 1.18. Summary of Willmott skill analysis for tidal tracer variables modeled by $k - \varepsilon$ G88. Variables that have skills exceeding 0.65 at each of the fixed-site network stations are shown. Surface variables are shown in (a) and bottom variables are shown in (b).

Subtidal tracer skill values are high, but they reveal spatial patterns in $k - \varepsilon$ G88 model performance. Subtidal surface temperature skills are higher than subtidal surface salinity skills. Subtidal surface salinity skills are seen to decrease somewhat from the north to the south in both the West Passage and East Passage. Subtidal bottom salinity skills decrease slightly southward along the West Passage. Subtidal bottom salinity and density skills at Quonset Point and Poppasquash Point, and subtidal bottom temperature skills at Quonset Point are much lower ($WS < 0.62$) than the other tracer skills. Subtidal salinity stratification skills are higher than subtidal temperature stratification skills. Modeling subtidal stratification at the Greenwich Bay Marina station produces poor agreement with the observations, despite the fact that subtidal surface and bottom salinity and temperature skills are high at this station.

For both tidal and subtidal time series the $k - \varepsilon$ G88 model surface density skills closely resemble surface salinity skills with surface temperature skills being higher than the surface density and salinity skills. Similarly, the bottom density skills are similar to the bottom salinity skills, but the bottom temperature skills are generally lower than the bottom density and salinity skills. In the cases of both tidal and subtidal time scales the density stratification skills are similar to the salinity stratification skills.

The spatially averaged skills for tidal surface and bottom east-west velocity are just below 0.60 when $k - \varepsilon$ G88 is implemented. However, at the West Passage station agreement between the model and observations is better for tidal surface and bottom east-west velocity with WS values exceeding 0.70. Similarly, the spatially averaged subtidal surface and bottom east-west velocity skills are slightly lower than 0.60, but the subtidal bottom east-west velocity exceeds 0.65 at the West Passage and East Passage Shoal stations. The model shows particularly high

skill in representing tidal surface north-south velocity at all stations, tidal bottom north-south velocity and subtidal surface north-south velocity at the East Passage Shoal station, and subtidal bottom north-south velocity at the East Passage Channel station.

It is necessary to assess model deficiencies that are not attributable to the choice of turbulence closure scheme. These deficiencies include general poor performance of the model in calculating tidal tracers, tidal and subtidal east-west velocity, and subtidal sea surface elevation irrespective of the turbulence closure implementation. Model shortcomings and the possible causes for relatively low skills are discussed in Section 1.3.2 in the context of the NB-ROMS model validation.

1.3.2 NB-ROMS model validation

Verification of the NB-ROMS model with $k - \varepsilon$ G88 turbulence closure implementation is conducted in this section using the model calculations for sea surface elevation, velocity, and tracers. The performance of the model is considered at tidal and subtidal time scales to test the validity of its application in an operational sense.

The response of the observed system and the comparative model predictions are explained in the context of the environment forcing functions. The observed forcing functions are discussed in the first part of the section, followed by an assessment of the tidal and subtidal characteristics of the model output compared to the observations. The first comparisons are between sea surface elevation constituents, followed by comparisons of tidal and subtidal barotropic and baroclinic velocity. Results of tidal and subtidal tracer comparisons are discussed last.

Environment variables

Estuarine circulation and transport are driven by the tide, wind and other atmospheric forcing conditions, and river runoff. Tides provide the dominant driving force for circulation in Narragansett Bay [52, 53]. The sea surface elevation record at Quonset Point measured relative to mean lower-low water indicates that the model validation period (day 150 – 227) spans four complete spring-neap cycles (Figure 1.19). The tide displays a diurnal inequality for the full duration of the time series with notable amplitude differences between successive high water and low water peaks. The sea surface elevation record can be referenced as the principal forcing function of tidal water property response.

Figure 1.19. Hourly sea surface elevation relative to mean lower-low water observed at Quonset Point for the maximum duration of time series used in the model validation.

The subtidal circulation and transport is dominated by wind and river transport (Figures 1.20 and Figure 1.21). Note that we use the convention of referring to the wind as the direction that it is blowing from, i.e. a southerly wind denotes wind that is blowing from the south. The wind exerts the strongest control on Narragansett Bay subtidal circulation [54]. River transport, atmospheric heating, and precipitation determine internal density differences. A few prominent wind events are observed for the time period of model validation. Two relatively strong wind events, a north-northeasterly and a northeasterly, are seen in short succession prior to day 160. These events are followed by the only sustained west-

northwesterly wind event between day 160.6 and day 162.6. A succession of sea breezes occurs after this event, fluctuating between southerly and southwesterly wind on diurnal time scales. Between day 180 and day 199 the sea breeze fluctuations are interrupted by four relatively short northerly and northeasterly wind events. A more sustained northeasterly wind event is seen between days 199.6 and 201.2. The northeasterly is followed by brief southwesterlies, westerlies, northerlies and northeasterlies. The end of the time series is characterized by a succession of sustained events. A southwesterly around day 218.9 shifts to become a relatively strong northerly around day 220, followed by a reversal to southwesterly around day 221.4. Another reversal leads to a longer sustained north-northwesterly and west-northwesterly event between day 222.1 and day 225.

Figure 1.20. Hourly wind forcing observed at Quonset Point for the maximum duration of time series used in the model validation. Red is the east-west wind component (positive is from the east) and blue is the north-south wind component (positive is from the north).

Pawtuxet River transport can be used to indicate the trend of river flow rates into Narragansett Bay for the period of interest (Figure 1.21). The Pawtuxet River is the third largest freshwater source in Narragansett Bay [55] and although its absolute transport rate is different from the other rivers of the watershed, it represents the general flow patterns of all the rivers. Pawtuxet River transport data show two high runoff events near the beginning of the time period considered for model validation. The maximum flow is observed on day 158 when the transport

of the Pawtuxet River reaches $95.4 \text{ m}^3/\text{s}$, which is an order of magnitude higher than the mean annual flow rate of the river [55]. The maximum transport of the second event occurs on day 176 when the transport of the Pawtuxet River is $51.3 \text{ m}^3/\text{s}$. The high runoff events coincide with periods of relatively high tidal range. After these two events the river transport decreases to about $9.6 \text{ m}^3/\text{s}$ on day 186 and then gradually tapers off to $5.3 \text{ m}^3/\text{s}$ on day 198. A small increase to $8.6 \text{ m}^3/\text{s}$ is seen on day 199, followed by decreasing transport for the remainder of the time period to $3.6 \text{ m}^3/\text{s}$ on day 227.

Two relatively large rainfall events are observed in Providence between days 152 and 158, and a third between days 173 and 176 (Figure 1.21). These rainfall events correspond to the observed high river transport events. For the remainder of the time series a few relatively small rainfall events ($<12.2 \text{ mm}$) are observed.

Air temperature at Quonset Point exhibits diurnal variability and a gradual increasing mean trend for the majority of the time period (Figure 1.21). The mean temperature increases from about 15°C to 30°C between day 160 and day 214. Mean air temperature decreases back to about 20°C from day 214 to the end of the period of interest.

Sea surface elevation

Tidal sea surface characteristics in Narragansett Bay have been well documented [56, 52, 55]. The tide dominates sea surface elevation and accounts for 85% of the total sea surface variance [52]. Observations of tidal sea surface elevation during the validation period agree with previous findings (Figure 1.22). The tidal sea surface elevation is dominated by the M2 semidiurnal harmonic constituent. The diurnal O1 and K1 constituents, semidiurnal N2 and S2 constituents, and the M4 and M6 overtones have markedly lower amplitudes, but are important in the relatively shallow estuarine domain. This is consistent with findings in other

Figure 1.21. Observed environment variables for the maximum duration of time series used in the model validation. Hourly air temperature data from Quonset Point is shown. Pawtuxet River transport and precipitation at the T.F. Green Airport are daily averages.

estuaries [57, 58, 59, 60, 61, 62]. Tidal amplification of the prominent constituents is observed from the mouth to the head of Narragansett Bay. Amplification factors between Newport and Providence are 1.20, 1.80, and 4.42 for the M2, M4, and M6 constituents, respectively. The tidal amplification of the M4 and M6 overtides is large due to resonance of these frequencies by the Narragansett Bay basin [52]. Tidal phase increases from Newport to Providence with phase lags of 6° , 26° , and 85° for the M2, M4, and M6, respectively.

The model is very successful in predicting tidal sea surface elevation. Predictions at the individual tide gauges produce skill values that exceed 0.97. The 78-day time series of the observed and modeled sea surface elevation are decomposed into tidal constituents using the T-TIDE toolbox [49] and the seven largest constituents are compared (Figure 1.22). The modeled tidal sea surface elevations show similar characteristics to the observations. The modeled M2 constituent amplitude dominates at all four stations. The amplitudes of all modeled constituents diminish southward. The observed decrease in M2 amplitude between the Providence and Newport stations is 0.10 m, whereas the modeled decrease is 0.09 m. The model underestimates the M2 amplitude with an error of about 10% at all four the stations. The remainder of the prominent constituents have observed amplitudes less than 0.13 m compared to the M2 amplitude of 0.61 m at Providence. The model overestimates the amplitudes of these constituents. The observed uncertainties of the M4 and M6 overtide amplitudes are relatively large. The minimum errors between observed and modeled amplitudes of the M4 and M6 overtides at the Providence tide gauge are 16% and 10%, respectively. The modeled amplification of M2 and M4 is accurate, but the amplification of M6 is twice as high as observed.

The accuracy of modeled tidal sea surface elevation is also considered in terms

Figure 1.22. Comparisons of observed (blue) and modeled (red) amplitudes of sea surface elevation for the seven major tidal constituents at four tide gauge stations.

of tidal constituent phases (Figure 1.23). The phase error of the model is less than 10% for the O1, K1, N2, and M2 constituents at all tide gauges. M2 phase lag between Newport and Providence is the same as observed. The model is less successful in reproducing the phases of S2, M4, and M6. Disagreement between the model and the data is particularly large for the M4 constituent. The M4 phase lag between Newport and Providence is 10° lower than observed. The difference in the modeled and observed M4 phase varies between 66° and 76° at the four tide gauge stations. Modeled M6 phase lag is not consistent with the observations, which show an 85° increase from Newport to Providence. The modeled M6 phase has a maximum at Quonset Point and a 56° difference between Newport and Providence. Note that relatively large discrepancies between observed and modeled tidal constituent phases are found in some other modeling studies [63, 64, 65, 66].

The relative inaccuracy in modeled phases of overtides may be considered

negligible due to the relatively small amplitudes (< 0.1 m at Newport) of these constituents [64]. However, in Narragansett Bay the overtides are instrumental in tidal current asymmetry. The phase changes of the M4 and M6 overtides relative to the M2 constituent cause double-peaked flood tidal currents [54, 52, 53, 67, 7]. In the next section it is shown that the double-peaked flood is weakly modeled during periods of high tidal range.

To assess the possible causes for inaccuracies of the M4 and M6 amplitudes and phases in the model we can consider the generation mechanisms of these overtides. The M4 overtide is generated through three mechanisms, namely nonlinear flux of continuity, nonlinear advection of momentum, and nonlinear bottom friction [68, 69]. Other studies have shown [69] that nonlinear continuity accounts for 73% of M4, frictional momentum loss due to changes in wave depth accounted for 20% of M4, and nonlinear advection accounted for 7% of M4. The M6 overtide is generated through quadratic friction [68, 69].

A sensitivity study tested the effect of bottom stress, boundary condition, time step, bathymetry, and grid resolution on the modeled accuracy of M4 [58]. The study showed no improvements through tuning of bottom stress and time step, or by increased grid resolution. Only small improvements occurred through bottom drag coefficient tuning. The study found that the M4 overtide was improved when channel representation was refined. Modifying the boundary conditions of the larger domain model by assigning an M4 phase that agrees with observations caused improvement of the M4 overtide in the nested model, but this modification affected all tidal constituents.

A number of possible issues in the NB-ROMS model could lead to inaccurate overtides. Bed shear stress of the ROMS model is based on a logarithmic formulation that uses a constant bottom roughness length. This parameter is known to be

spatially and temporally variable in real estuaries, depending on bed sediment and geometry, wind/wave effects, and density stratification [70, 71, 72, 12]. A variable bottom roughness length could improve overtide representations in the model. The effect of relatively large runoff on the modeled representation of M4 should also be considered, since the validation period occurs during flood events. Large river flow decreases the tidal range of M2 and transfers energy into M4 [69]. It is shown in a later section that the model overestimates freshwater runoff during flood events, which could cause inaccuracy in the relative representations of M2 and M4 and the consequent current asymmetry. However, the modeled ratio of the M4 and M2 amplitudes is the same during periods of high and low runoff, showing that inaccuracy in runoff specification does not cause these errors. Smoothing of the channel bathymetry in the model could lead to a misrepresentation of bed shear stress through its dependence on depth [68, 69]. The model boundary could also be a source of inaccuracies in tidal characteristics.

Figure 1.23. Absolute difference between observed and modeled phases of sea surface elevation for the seven major tidal constituents at four tide gauge stations.

Subtidal sea surface elevation is predicted with relatively low skill. Skill values decrease from the northern to the southern tide gauge stations. Comparisons of observed and modeled subtidal sea surface elevation at the four stations (Figure 1.24) show that the magnitudes of fluctuations about the mean are underestimated by the model. One standard deviation of the observed data sets vary between 0.07 m to 0.09 m, whereas one standard deviation of the modeled output is 0.03 m at all four stations. Furthermore, the correlations between the model output and the observations are low. The correlation coefficient (statistically significant at the 95% confidence level) varies between 0.27 and 0.28 at the three southern stations and is 0.37 at the Providence tide gauge station. The discrepancy between observed and modeled subtidal sea surface elevation may be attributed to the boundary forcing specification of sea surface elevation in the model. The low variance in subtidal sea surface elevation predicted at the tide gauge stations is also seen at grid cells of the model boundary. A mechanism for the inaccuracy of the model boundary is discussed in more detail later, in the context of subtidal barotropic currents. Low variance in the modeled subtidal sea surface elevation could also be due to the lack of atmospheric pressure in the model.

Barotropic and baroclinic velocity at tidal time scales

The tidal currents of Narragansett Bay are a dominant aspect of the flow. Between 80% and 95% of the current variance is explained by the tide [52, 53]. Hypoxic events of Narragansett Bay have been linked to tidal flow dynamics [35]. Tidal current magnitudes vary across the Bay with strongest flow observed in narrow passages and weaker currents in coves [55]. Tidal current harmonics are analogous to tidal sea surface harmonics in terms of the relative dominance of constituents. The combination of relatively strong overtides with the principle lunar semidiurnal tide result in the characteristic double-peaked flood currents

Figure 1.24. Comparison of observed (black) and modeled (grey) subtidal sea surface elevation at the four tide gauge stations.

[54, 52, 53, 67, 7].

The observed tidal current varies between the ADCP sites during the period of the model validation. The maximum tidal current magnitudes occur at the West Passage station where velocities reach 0.65 m/s during the spring tide. Velocities at the East Passage Channel and East Passage Shoal stations reach maxima of 0.35 m/s and 0.20 m/s, respectively. Tidal flow at the West Passage and East Passage Channel sites are confined by the channels, such that very low variance is seen in the direction perpendicular to the channel axes. The double-peaked flood is observed at all three ADCP stations for the duration of the time series.

Skill assessments reveal that the tidal velocity is modeled with varying degrees of success. The model captures the tidal surface north-south velocity component with high accuracy at all three of the ADCP stations. Skills are slightly lower and vary somewhat between stations for the tidal bottom north-south velocity, but the

model nonetheless reproduces this velocity component with high accuracy. The East Passage Shoal station exhibits highest overall tidal north-south velocity skill, and the West Passage station has lowest tidal bottom north-south velocity skill. The tidal surface and bottom east-west velocity component is reproduced well by the model at the West Passage station and less successfully at the East Passage Shoal station. There is low agreement between the modeled and observed tidal east-west velocity component at the East Passage Channel station. Unlike the tidal north-south velocity skills, the tidal east-west velocity skills of each station are similar for surface and bottom currents.

The tidal barotropic velocity is represented well by the model at the West Passage station (Figure 1.25). A small phase difference between the observed and modeled tidal barotropic north-south current is seen at the West Passage station where the maximum ebb occurs about two hours early in the model. The magnitude of the ebb current is slightly underestimated by the model at this station. The double-peaked flood seen in the data is captured well by the model only during times of low tidal range. During periods of high tidal range the double-peaked flood is not as well defined in the model as in the data. These modeled characteristics can also be seen in the tidal barotropic east-west velocity at the West Passage station (Figure 1.26). The east-west component is of similar magnitude than the north-south component during times of low tidal range. During times of high tidal range the east-west component is approximately 18% less than the north-south component. The modeled north-south and east-west tidal barotropic components thus exhibit similar small deviations from the observations at this station. The barotropic component does not explain the difference in overall north-south and east-west velocity component skills at this station.

The idea that the tidal barotropic component is not instrumental in explaining

data-model velocity discrepancies at the West Passage station is supported by the fact that the barotropic north-south velocity at the East Passage Channel and East Passage Shoal stations shows more discernible disagreement with the observations than at the West Passage station (Figure 1.25). The overall tidal north-south velocity skills, which incorporate both the barotropic and baroclinic components, are higher at the East Passage Channel and East Passage Shoal stations. At the East Passage Channel station there is no phase difference between observed and modeled tidal barotropic north-south flow. The amplitude of the ebb current is represented well by the model at this station, but the flood current is overestimated at times. The double-peaked flood current is reproduced well during times of low tidal range, but captured with varying degrees of success during periods of high tidal range.

At the East Passage Shoal station the phase of the tidal barotropic north-south current is modeled accurately, but the flood current is overestimated by the model (Figure 1.25). No evidence of a double-peaked flood can be seen in model output at this station during times of high tidal range. Despite these small discrepancies between modeled and observed tidal barotropic north-south velocity at the East Passage Channel and East Passage Shoal stations, the *WS* values of tidal north-south velocity are high.

The tidal barotropic east-west velocity components are modeled with low accuracy at the East Passage Channel and East Passage Shoal stations (Figure 1.26). The observed tidal barotropic east-west currents at both stations are relatively weak and they do not exhibit regular tidal oscillations, as is the case at the West Passage station. The disagreement with the observations seem higher at the East Passage Channel station, since both the flow variability and flow amplitudes are miscalculated by the model. The flow amplitudes are better represented at the

Figure 1.25. Comparisons of observed (black) and modeled (grey) tidal barotropic north-south velocity at the three ADCP stations for a portion of the time series. Positive is northward flow and negative is southward flow. Note: Velocity scales differ between stations.

East Passage Shoal station, but not the flow variability. The tidal barotropic component of the east-west flow may contribute to low skill values for the tidal east-west currents at these stations.

Figure 1.26. Comparisons of observed (black) and modeled (grey) tidal barotropic east-west velocity at the three ADCP stations for a portion of the time series. Positive is eastward flow and negative is westward flow. Note: Velocity scales differ between stations.

Plots of east-west versus north-south tidal barotropic currents show the small discrepancies between modeled and observed flow (Figure 1.27). The observed tidal barotropic current of the West Passage station is relatively strong and seems to be confined by the channel in a northeast-southwest alignment. The model-predicted flow shows more variance perpendicular to the northeast-southwest axis during the flood current and the ebb current is weaker than observed. In contrast, the model predicts less tidal barotropic variance in a direction perpendicular to the north-south axis of the East Passage Channel than is observed. The observed

flow at this station roughly exhibits two alignments, namely north-south and north-northwest to south-southeast. At the East Passage Shoal station the observed tidal barotropic current is highly variable. The model captures some of the variability, but it overestimates the flood current magnitude at this station.

Figure 1.27. East-west versus north-south tidal barotropic currents as observed (top row) and modeled (bottom row) at the three ADCP stations. Note: axes limits vary between stations.

Characterization of the tidal baroclinic flow is accomplished through Empirical Orthogonal Function (EOF) analyses. EOF analyses of the tidal baroclinic velocity profiles at the West Passage station show that the model captures the general aspects of the EOF modes. The first mode indicates observed linear shearing with a positive sign near the surface and a negative sign near the bottom. The model predicts a similar first mode, but with less linearity (Figure 1.28). The modeled first mode is lower near-surface than the observations and it increases in positive value to a maximum around a normalized depth of -0.25. It has a higher

negative value than the observations near the bottom. The model predicts that the first EOF mode explains 80% of the tidal baroclinic variability, but the observed first EOF mode explains 74% of the variability. The observed and modeled second EOF modes have similar shapes and positive subsurface maxima. The observed subsurface maximum is seen at a normalized depth of -0.40, and the model subsurface maximum is at a normalized depth of -0.45. The principal component time series of the first mode shows that the model often overestimates the intensity of this mode. The correlation coefficient obtained when the observed and modeled principal component time series are compared is 0.69. It is not clear from the EOF analysis why the WS value of the West Passage tidal north-south velocity is higher at the surface ($WS = 0.84$) than the bottom ($WS = 0.69$).

Figure 1.28. EOF analysis of observed (black) and modeled (grey) tidal baroclinic north-south velocity at the West Passage ADCP location. The mean and first two EOF modes of the north-south velocity profile are shown, along with a portion of the time series of the first mode principal component. The variance explained by the EOF modes are Var_d and Var_m for the data and model, respectively.

A comparison of modeled and observed tidal baroclinic north-south velocity profiles at the West Passage station provides insight to the discrepancy in tidal north-south surface and bottom velocity skills (Figure 1.29). The observed tidal baroclinic north-south component exhibits flood and ebb currents that vary in intensity with time. The model resembles the varying amplitudes of the observed flood and ebb currents near the surface. However, the modeled tidal baroclinic north-south bottom flow has relatively less variable maxima and minima in time and these amplitudes exceed the observed flow intensity. The amplitudes of the tidal baroclinic north-south bottom velocity at the East-Passage Channel station is similarly overestimated by the model. This may suggest that the channel bathymetry and friction of the model cause the lower overall skills in tidal north-south velocity near the bottom of the water column at these two deep stations. The shallower East Passage Shoal station does not exhibit the discrepancy between modeled and observed tidal baroclinic north-south bottom velocity and consequently has relatively high tidal north-south bottom velocity skill.

Modeled vertical profiles of tidal baroclinic north-south velocity at the East Passage Channel station give first and second EOF mode structures that resemble those of the observations (Figure 1.30). However, the modeled internal maxima of both modes are shifted slightly deeper in the water column compared to the observations. The variance explained by the first mode is predicted to be 74% and observed to be 64%. Agreement between the observed and modeled principal component is particularly high during the first half of the plotted time period when the tidal range is relatively low. The correlation coefficient for these time series is 0.68 during the sampling period. As in the case of the West Passage station, the EOF modes are represented relatively well by the model and are not instrumental in explaining the slight model deficiency of tidal north-south bottom velocity at

Figure 1.29. Observations (top) and model output (bottom) of tidal baroclinic north-south velocity profiles at the West Passage station for a portion of the time series. Positive is northward flow and negative is southward flow.

this station.

Figure 1.30. EOF analysis of observed (black) and modeled (grey) tidal baroclinic north-south velocity at the East Passage Channel ADCP location. The mean and first two EOF modes of the north-south velocity profile are shown, along with a portion of the time series of the first mode principal component. The variance explained by the EOF modes are Var_d and Var_m for the data and model, respectively.

Modeled EOF modes of tidal baroclinic north-south velocity at the East Passage Shoal station compare well with the observations, except for small differences in maxima near the surface and bottom, and internally (Figure 1.31). The observed modes and the model representations are similar to those seen at the other two stations. The variance explained by the first mode is predicted to be 59% by the model and observed to be 70%. The match between modeled and observed principal component is less good than at the other two stations and produces a correlation coefficient of 0.45.

An EOF analysis of the tidal baroclinic east-west vertical profiles at the West

Figure 1.31. EOF analysis of observed (black) and modeled (grey) tidal baroclinic north-south velocity at the East Passage Shoal ADCP location. The mean and first two EOF modes of the north-south velocity profile are shown, along with a portion of the time series of the first mode principal component. The variance explained by the EOF modes are Var_d and Var_m for the data and model, respectively.

Passage station shows good agreement between the first and second EOF modes of the model and observations (Figure 1.32). The variance explained by the first mode is predicted to be 70% and observed to be 67%. The modeled variance explained by the second mode is 21% and the observed variance is 19%. The principal component time series of the first mode shows that the model captures the magnitude of this mode and some of the variability, notably some prominent peaks during the first half of the plotted time period when the tidal range is relatively low. The correlation coefficient between the observed and modeled principal component time series is 0.34.

Figure 1.32. EOF analysis of observed (black) and modeled (grey) tidal baroclinic east-west velocity at the West Passage ADCP location. The mean and first two EOF modes of the east-west velocity profile are shown, along with a portion of the time series of the first mode principal component. The variance explained by the EOF modes are Var_d and Var_m for the data and model, respectively.

Weak agreement exists between the modeled and observed EOF modes of the tidal baroclinic east-west velocity profiles at the East Passage Channel station.

The modeled first EOF mode explains 41% of the variance and the observed first EOF mode explains 35% of the variance. The modeled second EOF mode explains 24% of the variance and the observed second EOF mode explains 25% of the variance. It appears that the first EOF mode of the data is reproduced by the model as the second EOF mode. Accordingly, the second EOF mode of the data resembles the first EOF mode of the model. It is deduced that a disparity exists between the predicted and observed tidal baroclinic east-west flow regimes at the East Passage Channel station. The mismatch contributes to the relatively low skill of the modeled tidal east-west velocity at this site. This is in addition to the low accuracy obtained for the tidal barotropic east-west velocity, as previously discussed. An assessment of the measurement error reveals that the tidal east-west velocity signal at this station is very low. The average error is found to be 52% of the east-west velocity signal at the East Passage Channel site.

The EOF modes for the modeled and observed tidal baroclinic east-west velocity agree well at the East Passage Shoal station (Figure 1.34). The first EOF modes of the model and observations explain 73% and 65% of the variance, respectively. A comparison of the modeled and observed principal component time series shows high agreement for most of the sampling duration with a correlation coefficient of 0.69. The good agreement between modeled and observed tidal east-west baroclinic flow implies that the relatively low skill seen for tidal east-west velocity at the East Passage Shoal station is due to the tidal barotropic flow disparity.

In summary, the results for the tidal barotropic and baroclinic velocity at the three ADCP stations show the sources of inconsistencies in the model. Slight deficiencies of the modeled tidal north-south bottom velocity at the West Passage and East Passage Channel stations can be explained by the baroclinic component of the flow. The model predicts less temporal variability and higher magnitudes for

Figure 1.33. EOF analysis of observed (black) and modeled (grey) tidal baroclinic east-west velocity at the East Passage Channel ADCP location. The mean and first two EOF modes of the east-west velocity profile are shown, along with a portion of the time series of the first mode principal component. The variance explained by the EOF modes are Var_d and Var_m for the data and model, respectively.

Figure 1.34. EOF analysis of observed (black) and modeled (grey) tidal baroclinic east-west velocity component vertical profiles at the East Passage Shoal ADCP location. The mean and first two EOF modes of the east-west velocity profile are shown, along with a portion of the time series of the first mode principal component. The variance explained by the EOF modes are Var_d and Var_m for the data and model, respectively.

the amplitudes of tidal baroclinic north-south bottom oscillations than is observed. This phenomenon is most prominently seen at the West Passage station where the bottom skill is lowest, and to a slightly lesser extent at the East Passage Channel station where the bottom *WS* value is higher. Both of these two stations are located within relatively deep channel bathymetry (Figure 1.35).

The cause of relatively high tidal current amplitudes may be related to model bathymetry specifications. Energy spectra of the tidal baroclinic north-south bottom water at the West Passage and East Passage Channel stations show an over-estimation of tidal current amplitude by the model. These spectra display higher modeled energy at the M2 tidal period than is observed, even though the M2 sea surface amplitude is underestimated by the model. An energy spectrum of the tidal baroclinic north-south bottom velocity at the East Passage Shoal station shows lower modeled energy at the M2 tidal period than is observed. At the East Passage Shoal station, where bathymetry is relatively flat, the modeled tidal baroclinic north-south velocity matches the observations. The model specification of channel bathymetry may lead to less accurate tidal flow in the deep portions of Narragansett Bay. The modeled and observed bathymetry compare well, but the model bathymetry is somewhat smoother (Figure 1.35). The smoothing leads to shallower channel depths, which would affect the accuracy of bed shear stress [68, 69]. Lateral friction, which can be significant near the bottom of a relatively narrow channel and could lead to an increase in the effective drag coefficient along the channel banks [73], may not be correctly represented by the model.

The observed tidal east-west velocity at the East Passage Channel station has a large uncertainty associated with it and using the data set for model validation appears not to be constructive. The East Passage Channel station is located within a north-south, longitudinally aligned channel. The observed east-west current at

Figure 1.35. NB-ROMS model grid bathymetry (top) compared with observed bathymetry (bottom). ADCP locations are indicated by black dots. (Bathymetry data from <http://www.geomapapp.org>)

this location is about six times weaker than the north-south current component. The average standard deviation of the east-west component of the flow is only two times larger than that of the measurement error. The disagreement between modeled and observed tidal east-west velocity at the surface and bottom of the East Passage Channel station can be explained by both the barotropic and baroclinic components of the flow. Although the observed tidal east-west current at this station has a relatively low magnitude, the model estimates an even lower flow magnitude. The model also does not reproduce the variability of the amplitudes of the observed tidal flow.

The tidal east-west velocity at the East Passage Shoal station is modeled with relatively low success. The model does not capture the spatial variability of the east-west barotropic flow at this site. The modeled barotropic component exhibits a more regularly oscillating flow than is observed.

The general mismatch in the double-peaked flood current is consistent with the disagreement between modeled and observed tidal constituents. Relatively large discrepancies are found between the modeled and observed phases of the M4 and M6 tidal constituents. Since the phase changes of M4 and M6 relative to M2 determine the characteristics of the double-peaked flood, the model representation of the double-peaked flood is compromised.

Barotropic and baroclinic velocity at subtidal time scales

The subtidal currents of Narragansett Bay exert an important influence on the long term transport and exchange of constituents. The subtidal flow has been measured to explain 8% to 45% of the current variance of Narragansett Bay [54, 52]. The wind is a dominant determinant of subtidal current, generally causing the surface to flow in the direction of the wind and the bottom to flow in the opposite direction to the wind [54, 52, 55]. The wind also modifies the average

counterclockwise flow of water entering through the East Passage and exiting from the West Passage [74, 67, 7, 75].

Observations of the subtidal current during the period used in the model validation agree well with previous findings [54, 76, 67, 75]. The influence of the wind dominates the subtidal flow characteristics. The maximum subtidal flow magnitude at the West Passage station is 0.20 m/s. Highest flow rates at this station correspond with southerly and southwesterly wind events. The temporal mean subtidal velocity is northwestward near the surface, weakly southward in the middle of the water column, and west-southwestward near the bottom. At the East Passage Channel and East Passage Shoal stations the maximum subtidal flow rates are 0.25 m/s and 0.13 m/s, respectively. The temporal mean subtidal velocity at the East Passage Channel station is southeastward in a thin surface layer, and the bottom 60% of the water column flows northward. At the East Passage Shoal station the top half of the subtidal water column flows toward the west-southwest and the near-bottom is northwestward.

Model predictions of subtidal velocity indicate variable performance for velocity components at the different stations. The skills in computing subtidal north-south velocity are higher than the skills for east-west velocity. The model is particularly accurate in representing the subtidal bottom north-south velocity component at the East Passage Channel station, which is the dominant inbound current in Narragansett Bay, but it is least successful in predicting the east-west velocity at this station. At the West Passage and East Passage Shoal stations the subtidal surface north-south velocity is reproduced with a Willmott skill that is 0.04 and 0.11 higher than the subtidal bottom north-south velocity, respectively. The model performs very well at calculating the subtidal north-south surface velocity of the East Passage Shoal station.

The subtidal barotropic north-south current at the West Passage station is observed to be more variable in time than is predicted by the model (Figure 1.36). The model underestimates the amplitudes of the observed fluctuations. For portions of the time period, the model also incorrectly predicts the mean flow direction. For example, the model predicts the flow to be northward for most of the first half of the time series, although the observed flow fluctuates between northward and southward events and is more frequently southward during that time period.

At the East Passage Channel station the model underestimates the intensity of the subtidal barotropic north-south flow fluctuations. The observed and modeled temporal average flow direction is northward, but the model fails to capture a southward flow event on day 178.7. The observed subtidal barotropic north-south flow at the East Passage Channel station exhibits patterns that correlate with the north-south wind direction. The southward barotropic flow on day 178.7 corresponds with a relatively strong southerly wind event. Similarly, the prominent wind shifts between days 218 and 227 seem to cause subtidal barotropic flow in the channel that shifts in an opposite sense to the wind stress. It follows that the majority of the channel-aligned water column responds in the opposite direction to the wind stress during prominent wind events. The model response is weaker than is observed. At the East Passage Shoal station the model performs well, but it does not capture the barotropic southward flow event between days 218 and 220.

The modeled subtidal barotropic east-west current at the three ADCP stations show a similar mismatch with the observations as seen for the north-south flow. The observed subtidal barotropic east-west flow at the West Passage station is of comparable magnitude and temporal variability than the north-south component (Figure 1.37). The model generally predicts lower mean subtidal barotropic east-west flow magnitudes at this station than are observed. The modeled fluctuations

Figure 1.36. Comparison of observed (black) and modeled (grey) subtidal barotropic north-south velocity at the three ADCP stations. Positive is northward flow and negative is southward flow. Note: Velocity scales differ between stations.

have lower amplitudes than the observed fluctuations. The data show that this barotropic component exhibits relatively strong eastward and westward flow events, but many of these events are not reproduced by the model. The observed subtidal barotropic east-west velocity at the East Passage Channel station is of very low magnitude and has a high uncertainty. This data set may consequently not be suitable to use for model validation. At the East Passage Shoal station the model closely predicts the subtidal barotropic east-west mean flow, but it underestimates the flow variance.

Figure 1.37. Comparison of observed (black) and modeled (grey) subtidal barotropic east-west velocity at the three ADCP stations. Positive is eastward flow and negative is westward flow. Note: Velocity scales differ between stations.

The model tends to underestimate subtidal barotropic velocity at all three ADCP stations and this disparity can be investigated. If it is assumed that the underestimation is spatially ubiquitous in the NB-ROMS model, a probable cause may be the misrepresentation of the longitudinal (axial) subtidal sea surface ele-

vation gradient. The sea surface elevation gradient is obtained by subtracting sea surface elevation time series, given relative to the NAVD88 datum [34], at Newport from those at Providence Harbor. The NAVD88 datum is a fixed reference for elevations determined by geodetic leveling [34]. Positive sea surface gradient or slope is defined to point in the direction of increasing height. It is seen that the temporal mean trend of the observed subtidal sea surface slope is underestimated by the model (Figure 1.38 (d)). A relationship can be discerned between observations of the wind, the longitudinal subtidal sea surface elevation gradient, and the subtidal barotropic north-south velocity (Figure 1.38). Note that the sea surface elevation gradient is positive for the majority of the time series, i.e. directed from the mouth to the head of the estuary.

Four events are discussed to indicate the relationship between the wind, sea surface gradient, and the subtidal flow. The first event exhibits a relatively strong northward wind that causes a sizable increase in the sea surface elevation gradient followed by an eastward wind, which promotes a decrease in the surface slope. The observed subtidal barotropic north-south component in the East Passage Channel shows a strong southward flow when the subtidal sea surface slope is high and shifts sharply northward with the subsequent lowering of the slope. Despite the fact that the model wind forcing closely resembles the observed wind at Quonset Point, the model underestimates the subtidal sea surface slope and doesn't represent the southward barotropic flow.

The second event marks a period of relatively calm wind on average. The observed subtidal sea surface slope is relatively low and the subtidal barotropic flow in the East Passage Channel is predominantly northward. The model underestimates the subtidal sea surface elevation gradient, but the model-data difference is smaller when the observed slope magnitude is low. The model performs relatively

Figure 1.38. Comparison of observed (black) and modeled (grey) properties: (a) Barotropic north-south baroclinic velocity at EPc; (b) north-south wind velocity (positive is toward north) at Quonset Point; (c) east-west wind velocity (positive is toward east) at Quonset Point; and (d) difference in subtidal sea surface elevation between Providence Harbor and Newport. Orange rectangles mark the events that are discussed in the text.

well in capturing the subtidal barotropic flow.

The third period presents a strong sustained southwestward wind that shifts rapidly to the northeast. The southwestward wind causes a dramatic lowering of the subtidal sea surface slope that leads to a brief reversal of the slope direction, i.e. pointing from the estuary head to the mouth. The subtidal barotropic channel flow is strongly northward during this time. The subsequent northeastward wind is associated with an increase in the slope magnitude in the direction of the estuary head and a concurrent southward barotropic channel flow. The model captures the trends, but underestimates the magnitude of the subtidal sea surface slope and barotropic current for the duration of the third event.

The fourth period shows a sequence of wind fluctuations between northeastward and southward, or southeastward. Similar to previous findings, the northeastward wind leads to an increase in the subtidal sea surface slope and southward or stalled barotropic flow, whereas the southward or southeastward wind causes the estuary slope to decrease and the flow to be northward. The model correctly predicts the fluctuations in the slope and the associated barotropic flow. However, the model underestimates the magnitudes of the sea surface slope and barotropic flow when the slope is relatively high.

The relatively low predicted subtidal sea surface gradient in the model may be attributed to the specification of the model wind forcing. Studies have shown that the estuarine subtidal sea surface slope is primarily determined by the local wind forcing [77]. If the specified axial wind magnitudes are underestimated it would lead to an underestimation of the subtidal sea surface slope. The band averaged spectral energy density of the observed north-south wind component at Newport, Quonset Point, and Potter Cove is compared to that of the model wind forcing (Figure 1.39). The model-specified wind magnitude exceeds the observed

magnitude at Potter Cove at all spectral energies. The wind forcing of the model has a total variance of $5.4 \text{ m}^2/\text{s}^2$, and the variance of the observed wind at Potter Cove is $4.0 \text{ m}^2/\text{s}^2$. The specified wind is similarly high compared to the Newport site, but is somewhat lower than the Quonset Point site where the total variance is $6.7 \text{ m}^2/\text{s}^2$. The wind data show the spatially varying nature of the wind, which is not incorporated in the model. Differences between the modeled and observed wind across the estuary may not be responsible for the disparity in subtidal axial sea surface slope, since the applied wind is the average of observations along the coast of Narragansett Bay.

Figure 1.39. Observed (black) and modeled (grey) band averaged spectral energy density of the north-south wind component. Model wind (a) is compared with observations from (b) Newport, (c) Quonset Point, and (d) Potter Cove. Spectra represent the period from day 170 to day 227. The dashed lines demarcate the 95% confidence interval.

The mechanism by which the model underestimates the subtidal sea surface elevation gradient and barotropic current may relate to wind specifications on the

shelf of the larger domain model [7] that provides boundary conditions to the NB-ROMS model. It is known that subtidal sea surface elevation gradients and subtidal currents are influenced remotely by Ekman processes on the shelf [78, 79, 80]. The wind that is specified on the shelf of the larger domain model is the same spatially uniform wind that is implemented in the NB-ROMS model, which is derived from observations of wind inside the estuary. An observational study showed that the shelf wind is twice as high and rotated clockwise by 15° to 27° as compared to the wind at Newport [75].

The inaccuracy of the modeled shelf wind may lead to a misrepresentation of shelf processes that influence circulation of the estuary. An estuary-shelf study showed that northeastward wind events caused an Ekman response of onshore bottom flow on the shelf and a stalling of the deep inflow of the East Passage [75]. Note that a stalling of the deep inflow is consistent with the observed southward or zero subtidal barotropic flow at the East Passage channel site during northeastward wind events (Figure 1.38 (a)). This Ekman shelf process relates to a lowering of the subtidal sea surface at the mouth, which occurs together with the increased axial estuary sea surface slope from the northward (up-estuary) wind component. Since the model wind is weaker on the shelf than is observed, the sea level decrease caused by Ekman shelf processes may be underestimated. This would cause the subtidal sea surface slope and barotropic current to be underestimated by the model during northeastward wind events.

The estuary-shelf study also showed that southwestward wind caused offshore bottom flow on the shelf and an enhancement of deep estuary-shelf exchange [75]. An enhancement of deep estuary-shelf exchange is consistent with northward subtidal barotropic flow at the East Passage Channel site during southwestward wind events (Figure 1.38 (a)). The Ekman process would lead to an increase in the sub-

tidal sea surface elevation near the mouth that works together with the southward (down-estuary) wind component to decrease the axial estuarine sea surface slope. Since the model wind is weaker on the shelf than is observed, the sea level increase at the mouth due to the shelf processes may be underestimated. This can cause the subtidal sea surface slope and barotropic current to be underestimated during southwestward wind events.

Data-model comparisons of time series profiles of the subtidal baroclinic velocity provide further insight to the performance of the model. These comparisons show disagreement between observed and modeled current shear at the West Passage station (Figure 1.40). Strong southward bottom pulses accompany pulses of northward surface flow in the data. The model captures the surface inflow and bottom outflow pulses for a few isolated events centered on days 172.7, 182.2, 191, 195.6, 198.6, 202.8, 206, and 218.7. These modeled pulses are of shorter duration and lower magnitude than the observed pulses. Discounting these pulses, the modeled surface flow is predominantly southward. Stronger and longer persisting northward pulses are seen near the middle of the water column, not near the surface as the observations suggest. The observations thus suggest two-layered subtidal baroclinic north-south flow at the West Passage station, but the model produces a three-layered flow.

Comparisons of the modeled and observed subtidal baroclinic east-west velocity profiles at the West Passage station show that the model correctly predicts the predominantly westward near-surface flow (Figure 1.41). Near the surface the model also captures the short-lived flow reversals centered on days 182.2, 202.5, and 218.7. The modeled subtidal baroclinic westward flow extends deeper into the water column than the observed westward surface flow. Eastward flow events near the bottom are sustained in the model and contribute to predominantly two-

Figure 1.40. Observations (top) and model output (bottom) of subtidal baroclinic north-south velocity profiles at the West Passage station. Positive is northward flow and negative is southward flow.

layered flow. In the data these eastward flow events are of shorter duration and are alternated by longer sustained westward flow events near the bottom, which are not predicted by the model. However, more sustained eastward flow events are observed near the middle of the water column, such that in some instances the water column has a three-layered flow structure.

Figure 1.41. Observations (top) and model output (bottom) of subtidal baroclinic east-west velocity profiles at the West Passage station. Positive is eastward flow and negative is westward flow.

The time series profiles of subtidal baroclinic north-south velocity at the East Passage Channel site show generally good agreement between observations and the model (Figure 1.42). Both the observed and modeled flow exhibit predominantly southward surface flow and northward bottom flow. The two-layered flow computed by the model is separated by a relatively uniform level of no motion that always occurs near the middle of the water column. The observed position of the level of no motion is more irregular in time and occasionally occurs at shallower

locations in the water column than predicted by the model, such as around days 175, 193, and 209. The modeled northward pulses of the lower layer are sustained for longer periods of time than the observed northward pulses. The model slightly underestimates current magnitude during three occasions of flow regime reversal, i.e. northward upper layer flow and southward lower layer flow, centered on days 182.5, 190.3, and 202.7. These flow reversals are associated with abrupt shifts in wind to southwesterly conditions. The model correctly predicts short periods of three-layered flow regimes where the surface and bottom flow is northward and the middle water column flow is southwards around days 177 and 205.8. The model is also successful in reproducing the vertical structure and magnitude of the flow variability between day 218 and 227, which is associated with strongly shifting wind events. During these wind events the subtidal baroclinic north-south near-surface current flows in the direction of the north-south component of the wind stress and the near-bottom flows in the opposite direction to the wind stress.

Good agreement exists between the modeled and observed time series profiles of subtidal baroclinic north-south velocity at the East Passage Shoal station (Figure 1.43). The agreement is particularly high during two events of southward flow near the surface and northward flow near the bottom centered on days 220.3 and 223.5. However, the model underestimates the magnitudes of surface-inflow bottom-outflow event around day 218.5.

The subtidal baroclinic east-west velocity profiles of the observations and model at the East Passage Shoal station show agreement in the structure of the flow between days 214.7 and 218, but not the magnitude of the flow (Figure 1.44). The model underestimates the intensity of the current during this period. Between days 218 and 219 the observations predict surface westward flow and bottom eastward flow, but the model predicts surface eastward flow and bottom westward

Figure 1.42. Observations (top) and model output (bottom) of subtidal baroclinic north-south velocity profiles at the East Passage Channel station. Positive is northward flow and negative is southward flow.

Figure 1.43. Observations (top) and model output (bottom) of subtidal baroclinic north-south velocity profiles at the East Passage Shoal station. Positive is northward flow and negative is southward flow.

flow. The model captures aspects of the flow, such as the flow events centered on days 220.3, 222, and 226, but the predicted magnitudes are not the same as the observations.

Figure 1.44. Observations (top) and model output (bottom) of subtidal baroclinic east-west velocity profiles at the East Passage Shoal station. Positive is eastward flow and negative is westward flow.

The results in this section reveal strengths and weaknesses of the model that lead to varying performance in predicting subtidal velocity. The modeled subtidal barotropic flow has less variance than observed at all stations. The low subtidal barotropic variance of the model is associated with a lower than observed axial subtidal sea surface gradient. The low modeled surface slope may be due to the boundary conditions that are obtained from a larger domain model. The wind field on the shelf of the larger domain model is the same as the wind applied across the estuary. It is known that the observed wind on the shelf has a higher magnitude and is shifted compared to the wind inside the estuary. Applying inadequate wind

stress on the shelf of the large domain model probably leads to reduced magnitudes in Ekman shelf processes and inadequate subtidal sea surface level changes at the mouth of Narragansett Bay.

Despite the relatively low modeled subtidal barotropic variance the model accurately predicts north-south velocity. The model is successful in predicting subtidal north-south flow at the East Passage Channel station. The vertical structure of the subtidal baroclinic current is captured with reasonable accuracy at this station. The relatively high skill of subtidal north-south velocity at the East Passage Shoal station may be attributable to the accurate match between modeled and observed baroclinic velocity profiles. The short time series used for model verification at this station is concurrent with prominent wind events that dominate the water response.

The moderate skills of the subtidal north-south and east-west velocity components at the West Passage station may be attributable in part to the barotropic velocity disagreement between the model and the observations. In addition, a model-data disparity exists in the vertical baroclinic velocity structure. The observed subtidal north-south baroclinic velocity has a two-layered vertical structure, but the model predicts a three-layered structure. In the case of the subtidal east-west baroclinic velocity the model predicts a two-layered vertical structure when a three-layered structure is observed.

Tracer variability at tidal time scales

The tidal characteristics of tracers in Narragansett Bay have been extensively studied in the context of water quality. A correlation has been found between the neap tide, which is concurrent with increased stratification, and the occurrence of hypoxia [81, 3, 6, 35]. Observed tidal tracers exhibit very small perturbations during the period used in the model validation. The tidal surface temperature

variances vary between $0.13\text{ }(^{\circ}\text{C})^2$ and $0.22\text{ }(^{\circ}\text{C})^2$ at all the buoy stations. Variances of tidal bottom temperature range between $0.03\text{ }(^{\circ}\text{C})^2$ and $0.46\text{ }(^{\circ}\text{C})^2$. Tidal surface salinity variance is less than 0.5 (psu)^2 at stations other than Bullock Reach and Conimicut Point. These two stations have relatively high tidal surface salinity variances of 1.50 (psu)^2 and 0.97 (psu)^2 , respectively. At Bullock Reach the tidal bottom salinity variance is 0.53 (psu)^2 , and at all other stations it is lower than 0.30 (psu)^2 .

Skill assessments show that tidal tracer variability is generally modeled with less accuracy than subtidal tracer variability. The model is nonetheless good at reproducing tidal tracers at several stations within Narragansett Bay. The tidal surface temperature is modeled with relative accuracy ($WS \geq 0.65$) at all stations, but North Prudence. Tidal bottom salinity, temperature, and density are all modeled with WS values of 0.70 or greater at Conimicut Point, Mount View, and Poppasquash Point. Tidal surface salinity, temperature, and density are all modeled with WS values exceeding 0.69 at Mount View and Quonset Point. The model reproduces all tidal surface and bottom tracers at Conimicut Point and Mount View with relative accuracy ($WS \geq 0.65$), as well as all tidal surface tracers at Quonset Point and all tidal bottom tracers at Poppasquash Point and T-Wharf ($WS \geq 0.65$). The tidal density skills resemble the skills of tidal salinity, which indicates that tidal salinity predominates tidal density. Similar tracer skill values have been found in other model validation studies [16]. Several questions regarding tidal tracer skills present themselves, namely:

- Why does the model generally have diminished capability in reproducing tidal tracers relative to other properties like tidal velocity and sea surface elevation?
- Why is tidal surface temperature modeled with relative accuracy ($WS \geq$

0.65) everywhere except at the North Prudence station (where tidal bottom temperature is also modeled inadequately)?

- Why does the model inadequately capture the tidal bottom temperature in the north at the Bullock Reach, Greenwich Bay Marina, North Prudence and Mount Hope Bay stations (even though it is modeled acceptably at the Conimicut Point station)?
- Why is the tidal bottom salinity modeled with relative high accuracy at the Poppasquash Point station, but not the tidal surface salinity?
- Why does the model exhibit low skill in modeling tidal salinity at the Greenwich Bay Marina and Bullock Reach stations?

The relatively low tidal surface temperature skill produced by the model at the North Prudence station is first considered. Some possible causes for the low skill are disregarded after tests revealed that their effect on the skill value is negligible. These include the 1-hour sampling interval and the choice of model grid point that represents the North Prudence station. The latter may be an issue if the modeled horizontal tracer gradients have subtle offsets from their observed locations. The tidal surface temperature output at grid cells surrounding the cell selected for North Prudence skill analyses gives similar or relatively diminished agreement with the data. The tidal tracer skill values change negligibly when records of 15-minute sampling intervals are compared.

The difference between modeled and observed tidal surface temperature at the North Prudence site is notable when compared with time series at other stations (Figure 1.45). It is expected that the tidal signal is reproduced well in the surface water at North Prudence, since the tidal surface salinity skill is relatively high at this location. Spectral energy density of the observed tidal surface temperature

Figure 1.45. Comparisons of observed (black) and modeled (grey) tidal surface temperature perturbations at the North Prudence (NP), Mount View (MV), Quonset Point (QP), and Poppasquash Point (PP) stations for a portion of the time series. The time series represents the entire period for analysis at the North Prudence site.

records (Figure 1.46) show two peaks. One peak is centered on the semidiurnal period and the other on the diurnal period. The model predicts a semidiurnal peak of similar magnitude than is observed, but the magnitude of the diurnal peak is overestimated by the model. The observed diurnal energy may have a tidal origin. The tide displays notable amplitude differences between successive high water and low water peaks (Figure 1.19). This tidal diurnal inequality causes diurnal inequality in tidal velocity (Figure 1.25) and may contribute to the diurnal signal of the tidal surface temperature. The diurnal spectral energy density peak of tidal surface temperature may also be due to surface processes, namely diurnal heating and cooling, or sea breeze fluctuations.

Figure 1.46. Band averaged spectral energy density of observed (left panels) and modeled (right panels) tidal surface temperature at the North Prudence (NP) station. Spectra represent the period from day 188 to day 210. The dashed lines demarcate the 95% confidence interval.

The origin of the diurnal energy peak of tidal surface temperature at North Prudence can be assessed. Diurnal heating cycles and diurnal tidal inequality cycles are delineated on time series of tidal surface temperature during spring and neap tides (Figure 1.47). Note that a positive increase in the tidal temperature perturbation is consistent with the occurrence of ebb tide, due to the outflow of relatively warm surface water from the head of the estuary. At the beginning of the spring tide the larger diurnal temperature perturbations coincide with both the maximum ebb current before higher low water and 1 pm Eastern Standard

Time (EDT) when surface heating is expected to be high. Towards the end of the spring tide and for most of the neap tide the larger positive diurnal temperature perturbations are still observed near 1 pm EDT, but are not associated with the maximum ebb current before higher low water. The negative perturbations around 1 am EDT are underestimated by the model during this time. It follows that the observed diurnal spectral density peak of tidal surface temperature at North Prudence more strongly indicates heating and cooling of the surface than the effect of the diurnal tidal inequality.

The disparity between observed and modeled tidal surface temperature is due to overestimated heating and underestimated cooling in the model. The North Prudence tidal surface temperature time series show that the model predicts the same diurnal tidal temperature cycle than the data, but it overestimates the magnitude of diurnal fluctuations (Figure 1.47). It is worth recalling that the M2 tidal amplitude is slightly underestimated by the model (Figure 1.22). Although sea breezes have a diurnal signal, no clear correlation can be found between the tidal surface temperature perturbations and sea breeze fluctuations.

The difference in diurnal spectral energy density of observed and modeled tidal surface temperature is not a local phenomenon. The energy is overestimated by the model at eight of the nine buoy stations. The correspondence between the diurnal tidal surface temperature peaks with the daily cycle of heating and cooling is more pronounced during time periods when the cycle does not overlap with the cycle of the diurnal tidal inequality. In addition, the overestimated modeled diurnal peaks are seen to coincide with the daily cycle of heating and cooling during those periods. This occurs at stations with longer data records than the North Prudence station. The larger mismatch at North Prudence as compared to the other stations is due to the fact that the cycles of diurnal heating and

Figure 1.47. Observed and modeled tidal surface temperature perturbation at North Prudence with relation to the observed tidal sea surface elevation at Conimicut Light. The spring tide (top panel) and neap tide (bottom panel) are shown. Vertical orange lines indicate 1 pm EDT when the sea surface temperature is expected to be relatively high. Vertical red lines indicate maximum ebb associated with diurnal higher low water.

cooling and diurnal tidal inequality overlap for most of the short data record. It follows that the difference between observed and modeled diurnal surface heating and cooling contributes to the somewhat imperfect tidal surface temperature skills at all buoy stations. The model bulk forcing specifications of diurnally varying air temperature, outgoing longwave radiation flux, and incoming shortwave radiation flux should be reassessed to improve model accuracy.

Tidal bottom temperature skills are markedly lower at Bullock Reach, North Prudence, and Mount Hope Bay than at Conimicut Point. A mechanism for the data-model disagreement cannot clearly be discerned from time series plots (Figure 1.48). Spectral energy density of the observed and modeled tidal bottom temperature at these stations indicate some differences (Figure 1.49). The majority of the observed tidal bottom temperature variance at Bullock Reach occurs at the semidiurnal and diurnal periods. The model predicts less energy at these periods and relatively high energy at periods that coincide with the M4 and M6 overtides. Variance of tidal bottom temperature is relatively low at North Prudence and Mount Hope Bay as compared to Bullock Reach and Conimicut Point. The amplitude and phase of measured tidal constituents are very similar at Providence and Conimicut Light (Figures 1.22 and 1.23), which indicates that the spectral energy density of the tide is similar at Bullock Reach, North Prudence, and Conimicut Point. Relatively low semidiurnal variance at North Prudence may be due to low bottom temperature gradients. The same may be true for Mount Hope Bay. At North Prudence the model underestimates a diurnal signal, and at Mount Hope Bay the model overestimates energy at the period of the M4 overtide.

The occurrence of tidal bottom temperature variance at overtide periods at Bullock Reach and Mount Hope Bay indicates that the bottom friction at these stations may be overestimated by the model. These findings are consistent with

Figure 1.48. Comparisons of observed (black) and modeled (grey) tidal bottom temperature perturbations at the Bullock Reach (BR), Conimicut Point (CP), North Prudence (NP), and Mount Hope Bay (MH) stations for a portion of the time series. The time series represents the entire period for analysis at the North Prudence site.

results from sea surface elevation harmonics (Figure 1.22). Sea surface elevation harmonics show that the model underestimates the amplitude of the M2 tidal constituent and it overestimates the amplitudes of the M4 and M6 overtones. The discrepancy between measured and modeled overtones decreases southward from the Providence River tide gauge. It appears that the slight inaccuracies in the predictions of tidal constituents cause reduced skills in tidal bottom temperature at Bullock Reach, Conimicut Point, and Mount Hope Bay.

At stations other than North Prudence and Greenwich Bay Marina the spectral energy density of tidal bottom salinity closely resembles that of tidal bottom temperature, both in the data and the model. Energy spectra of tidal bottom salinity at Bullock Reach and Mount Hope Bay show that the same arguments that explain relatively low tidal bottom temperature skills can be applied to tidal bottom salinity. Note that diurnal variance in tidal bottom salinity and temperature is generally attributed to diurnal inequality of the tidal current.

It is likely that the unique observed tidal surface salinity spectrum at Poppasquash Point is due to distinctive surface salinity fronts or freshwater runoff processes that are not represented by the model, leading to relatively low skill compared to the tidal bottom salinity at this station (Figure 1.50). The observed tidal bottom salinity perturbations at Poppasquash Point follow a regular semidiurnal oscillation, which is captured very well by the model. The observed tidal surface salinity perturbations exhibit generally low variance, but short periods of increased variance are seen to coincide with large freshwater runoff events. The model miscalculates the timing and amplitude of the increased variance and fails to capture the characteristics of the low variance. Spectral energy density of the observed and modeled tidal surface and bottom salinity at Poppasquash Point shows these differences (Figure 1.51). The observed tidal surface salinity exhibits the

Figure 1.49. Band averaged spectral energy density of observed (left panels) and modeled (right panels) tidal bottom temperature at Bullock Reach (BR), Conimicut Point (CP), North Prudence (NP), and Mount Hope Bay (MH). Spectra represent the time periods used in the model validation. The dashed lines demarcate the 95% confidence interval.

Figure 1.50. Comparisons of observed (black) and modeled (grey) tidal surface and bottom salinity perturbations at the Poppasquash Point (PP) station for a portion of the time series.

majority of variance over a broad spectrum between the semidiurnal and diurnal periods. In contrast, the majority of energy of the modeled tidal surface salinity is at the semidiurnal period and relatively low variance can be discerned at the diurnal peak. Note that the agreement between spectra of modeled and observed tidal bottom salinity is high at Poppasquash Point. Both the data and the model show a discrete peak at the semidiurnal period, which contains almost all of the energy. The modeled peak is higher than that of the data and thus contributes to a slight model-data discrepancy. Skills of tidal surface and bottom temperature are relatively high at Poppasquash Point.

Surface and bottom tidal salinity are modeled with very low accuracy at the Greenwich Bay Marina station. Skill values for surface and bottom tidal salinity are 0.22 and 0.20, respectively. When time series are compared it is clear that the model computes much weaker magnitudes of tidal salinity perturbations than are observed (Figure 1.52). Spectral energy density of observed tidal surface and bottom salinity exhibits peaks at the semidiurnal and diurnal periods, but no energy is detectable in any spectral band of modeled tidal surface and bottom

Figure 1.51. Band averaged spectral energy density of observed (left panels) and modeled (right panels) tidal surface and bottom salinity at the Poppasquash Point (PP) station. Spectra represent the period from day 150 to day 227. The dashed lines demarcate the 95% confidence interval.

salinity. For reference it can be noted that tidal surface temperature is modeled with relatively high accuracy with a skill of 0.76, and tidal bottom temperature has a skill value of 0.50.

The data-model discrepancy in tidal variance is seen in scatter plots of tracers at Greenwich Bay Marina (Figure 1.53). The standard deviations for tidal surface salinity of the data and the model are 0.6 psu and 0.1 psu, respectively. The standard deviations for tidal bottom salinity of the data and the model are 0.4 psu and 0.1 psu, respectively.

The fact that the model underestimates the tidal salinity variance can either indicate that the predicted tidal current magnitude of the subestuary is too low, or that the tidal salinity gradient along the longitudinal (lengthwise) axis of Greenwich Bay is underestimated. The latter may be due to the fact that the model lacks freshwater sources in Greenwich Bay.

Figure 1.52. Comparisons of observed (black) and modeled (grey) tidal surface and bottom salinity and temperature perturbations at the Greenwich Bay Marina (GB) station for a portion of the time series.

Figure 1.53. Scatterplots of modeled versus observed surface (top row) and bottom (bottom row) tidal tracers at the Greenwich Bay Marina station.

The strength of the tidal current in the region of the Greenwich Bay Marina monitoring site can be assessed by investigating current meter data in conjunction with the tidal sea surface elevation. A SeaHorse Tilt Current Meter (TCM) measured current flow in the bottom 1 m of the water column near the Greenwich Bay Marina buoy site in the summer of 2010 [82]. The near-bottom tidal current magnitude measured by the TCM has a mean of 0.02 m/s with a standard deviation of 0.01 m/s (Figure 1.54 (c)). The modeled near-bottom current magnitude at Greenwich Bay Marina in the summer of 2006 also has a mean of 0.02 m/s and a standard deviation of 0.01 m/s (Figure 1.54 (d)). The observed tidal sea surface elevation at Quonset Point during the period of TCM measurements are compared with modeled tidal sea surface elevation at Greenwich Bay Marina in 2006 (Figure 1.54 (a) and (b)). The comparison shows that the tidal ranges during the summer periods in 2006 and 2010 are similar. The tidal current as computed by the model is thus of the appropriate magnitude in western Greenwich Bay. It can be deduced that the low modeled salinity variance at the Greenwich Bay Marina site is due to an underestimated longitudinal salinity gradient resulting from the lack of freshwater entering the Greenwich Bay system in the model.

Two time periods of most prominent data-model disagreement in tidal surface salinity at the Greenwich Bay Marina station occur between day 157 (start of the time series) and day 162, and between day 175 and day 179. During these periods the observed salinity perturbations are large compared to the rest of the time series. These two periods correspond to very high freshwater runoff events (discussed in more detail in the next section). During those times the model completely underestimates the tidal surface salinity perturbation. The freshwater runoff events cause relatively strong longitudinal salinity gradients at the Greenwich Bay Marina location, but these gradients are not captured by the model due to the lack of

Figure 1.54. Comparison of western Greenwich Bay tidal current magnitude between tilt current meter data and NB-ROMS model output. (a) The observed sea surface elevation at Quonset Point during a month in 2010; (b) observed near-bottom current magnitude from a tilt current meter near the Greenwich Bay Marina during the same month in 2010; (c) NB-ROMS model output during 2006 of sea surface level at the Greenwich Bay Marina station; and (d) NB-ROMS modeled near-bottom current magnitude during 2006 at the Greenwich Bay Marina station.

freshwater input at the head of Greenwich Bay.

The cause of low tidal surface salinity skill at the Bullock Reach station seem to relate to inaccuracies seen at some other stations. It appears though that Bullock Reach is an extreme case. Spectral energy density of observed tidal surface salinity at Bullock Reach show high variance at the diurnal and semidiurnal periods (Figure 1.55 (b)). Comparisons of the variance of observed tidal surface salinity at all stations for the time period of the relatively short Bullock Reach data record show that Bullock Reach and Conimicut Point variance is nearly twice as high as the variance of the next highest station and four times higher than the average variance of the other stations. Observed tidal surface salinity variance roughly decreases with distance from the head of the estuary. The model exhibits energy at the same spectral bands as the Bullock Reach observations, but it underestimates the magnitude of the tidal surface salinity variance (Figure 1.55 (c)). Relatively high observed variance at the diurnal period is attributed to the diurnal inequality of the tidal current (Figure 1.55 (a)).

The difference between observed and modeled tidal surface salinity at Mount Hope Bay and Mount View is similar to that found at Bullock Reach, but it is not discernible at the Conimicut Point and North Prudence stations. The discrepancy between the model and the data at the former stations may be ascribed to station positions relative to freshwater sources. The data record of Bullock Reach corresponds to a period of very high river runoff during a flood. The Bullock Reach station is located on the western edge of the Providence River shipping channel. It is situated 2300 m south of the Pawtuxet River. Previous observations [76] and estuarine theory indicate that the Pawtuxet River and other major freshwater sources to the north have a dominant influence on the shallower area directly west of the shipping channel of the Providence River. The modeled longitudinal

subtidal surface salinity gradient between Bullock Reach and Conimicut Point is lower than observed for a large portion of the Bullock Reach data record, particularly during an extreme flood event that occurred between day 170 and day 180 (Figure 1.55 (d)). The larger observed gradient is consistent with larger observed tidal surface salinity variance. Due to their proximity to freshwater sources the Bullock Reach, Mount Hope Bay, and Mount View stations may be located within large surface salinity fronts that are not adequately predicted by the model. The following section describes that the model overestimates freshwater runoff during the flood. This is consistent with an underestimation of surface salinity gradient at these stations for the flood period.

The tidal tracer results aid in explaining questions posed at the beginning of this section. The following is a recapitulation of why tidal surface temperature is modeled poorly at North Prudence, why tidal bottom temperature is inadequately represented at the northern stations, why tidal surface salinity is misrepresented by the model at Poppasquash Point, and why tidal surface and bottom salinity is inaccurately modeled at Greenwich Bay Marina and Bullock Reach. The tidal tracer skills are generally lower than skills for the other variables due to the fact that the observed tidal tracer variances are small.

Tidal surface temperature is slightly miscalculated by the model due to overestimation (/underestimation) of the diurnal surface heating (/cooling). The disparity between modeled and observed tidal surface temperature is relatively large at the North Prudence site where the data record is short.

Very low observed tidal bottom temperature and salinity variance at North Prudence and Mount Hope Bay indicates weak bottom tracer gradients, which are miscalculated by the model. At Bullock Reach the model predicts the majority of variance to be located at overtide bands, whereas observed variance occurs at

Figure 1.55. (a) Observed and modeled tidal surface salinity perturbation at Bullock Reach with relation to the observed tidal sea surface elevation at Providence tide gauge. Vertical red lines indicate maximum ebb associated with diurnal higher low water. Band averaged spectral energy density of (b) observed and (c) modeled tidal surface salinity at Bullock Reach (BR). Spectra represent the period from day 152 to day 193. The dashed lines demarcate the 95% confidence interval. (d) Observed and modeled subtidal surface salinity gradient between Bullock Reach (BR) and Conimicut Point (CP).

the semidiurnal and diurnal bands. Some tidal bottom temperature and salinity energy also occurs at overtide periods at the Mount Hope Bay station. This is consistent with the differences in the amplitudes of observed and modeled tidal constituents.

Tidal surface salinity is misrepresented by the model at the Poppasquash Point station, although the other tidal tracer variables are predicted with relatively high accuracy at this site. It appears that the observed tidal surface salinity has a unique character, which may indicate local freshwater runoff or salinity gradients that are not represented by the model.

Observed tidal surface salinity variance is relatively high at stations in the northern regions of Narragansett Bay, particularly Bullock Reach where the variance is underestimated by the model. The model underestimates longitudinal surface salinity gradients at stations that have a dominant freshwater influence, notably during the flood when the model specifies an excess of freshwater transport in the north. The relatively low predicted surface salinity gradients lead to insufficient tidal surface salinity variance in the model at Bullock Reach, Mount Hope Bay, and Mount View.

Both tidal surface and bottom salinity is underestimated by the model at Greenwich Bay Marina. The model predicts that very low variance exists in these variables at this site, but significant variance is observed. The lack of variance in the model is attributable to the lack of freshwater runoff sources in the Greenwich Bay system that leads to low longitudinal surface and bottom salinity gradients.

Tracer variability at subtidal time scales

Observed tracers exhibit some subtidal variability during the model validation period. Mean subtidal surface temperatures range between 20 °C and 24 °C at all buoy stations. The mean subtidal bottom temperature range between 14 °C

and 22 °C with cooler temperatures at the southernmost stations and highest temperatures at Greenwich Bay Marina. Variance in subtidal surface and bottom temperature ranges from 3.5 (°C)² to 12.0 (°C)². Mean temperature stratification varies from 1.5 °C at Greenwich Bay Marina to 5 °C at Quonset Point. Subtidal surface salinity varies between 22 psu and 26.5 psu at stations other than Bullock Reach. At Bullock Reach the mean surface salinity is 14.5 psu. Note that the Bullock Reach data record is shorter than the other records and coincides with large runoff events. Mean subtidal bottom salinity varies between 28 psu and 30 psu at the buoy stations. Variance in subtidal surface salinity is 2.0 (psu)² at Quonset Point and 13 (psu)² – 19 (psu)² at Bullock Reach, Conimicut Point, and Mount Hope Bay. Subtidal bottom salinity variance is relatively low and ranges between 0.33 (psu)² and 3.9 (psu)² at all stations. Mean salinity stratification ranges from 2.5 psu at Greenwich Bay Marina to 6.5 psu and 13 psu at Conimicut Point and Bullock Reach, respectively.

Spatially averaged skill values show that subtidal tracers are computed with high accuracy by the NB-ROMS model. Subtidal surface temperature is modeled with highest skill. A few representative data-model comparisons show the high accuracy of modeled subtidal surface temperature (Figure 1.56). The predicted subtidal surface temperature exhibits high correlation and low root mean square errors at all the buoy sites. Skill assessments at individual stations indicate that the ability of the model to predict subtidal tracers varies spatially. Subtidal surface temperature skills exceed 0.88 at all stations, and the subtidal surface salinity skills exceed 0.88 at stations other than Quonset Point, T-Wharf, and Greenwich Bay. Higher spatial skill variability is calculated for subtidal bottom tracers. Subtidal bottom temperature skills are relatively low at the North Prudence, Quonset Point, and Poppasquash Point stations. Subtidal bottom salinity skills are relatively low

at the Quonset Point, Poppasquash Point, and T-Wharf stations.

Figure 1.56. Time series of observed (black) and modeled (grey) subtidal surface temperature at Conimicut Point, Mount View, Greenwich Bay Marina, and Poppasquash Point.

The most consistent discrepancy between the modeled and observed subtidal tracers is seen for bottom temperature (Figures 1.57 and 1.58). At individual stations the subtidal bottom temperature is modeled with lower accuracy than the subtidal surface temperature. Modeled subtidal bottom temperature correlates very well with the observed data set at eight stations, but a mean offset occurs for the majority of the time series record (Table 1.3). The magnitude of the offset varies between stations. The temporal mean offset varies between 1.3 °C and 3.3 °C at the southern stations (Figure 1.57), and between 1.6 °C and 2.3 °C at the northern stations when the entire length of each time series is considered. Note that at the Greenwich Bay Marina site the model overestimates the tidal bottom temperature. The relatively low skills for subtidal bottom temperature at

Conimicut Point, North Prudence, and Poppasquash Point can in large part be attributed to the fact that the mean offset is largest at these sites. The root mean square error at these stations ranges between 2.25 °C and 3.56 °C (Table 1.3). The skill value is lowest at Quonset Point where both a low correlation and high root mean square error is calculated.

Figure 1.57. Time series of observed (black) and modeled (grey) subtidal bottom temperature at the southern stations, namely Mount View, Quonset Point, Poppasquash Point, and T-Wharf.

The mean offset in subtidal bottom temperature is seen only for a portion of the time series. Modeled subtidal bottom temperature matches well with the observations near the beginning of the time series at Bullock Reach, Mount View, Quonset Point, and T-Wharf. Mean offset values are relatively low during the early portion of the time series at Mount Hope Bay and Poppasquash Point. Modeled and observed temperature values diverge roughly from days 170 – 175 onward. The same may be true at the Conimicut Point and North Prudence sites, but

Table 1.3. Skill values for subtidal bottom temperature at each of the nine fixed-site stations. *WS* is the Willmott skill, *CC* is the correlation coefficient (statistically significant at the 95% confidence level), and *RMSE* is the root mean square error.

Station	BR	CP	NP	MV	QP	PP	TW	GB	MH
<i>WS</i>	0.89	0.78	0.68	0.90	0.59	0.69	0.87	0.90	0.88
<i>CC</i>	0.97	0.96	0.98	0.95	0.73	0.92	0.97	0.91	0.97
<i>RMSE</i> ($^{\circ}\text{C}$)	1.96	2.38	2.25	1.76	2.30	3.56	1.90	1.70	2.32

these data records start later than the aforementioned stations. A large fraction of the relatively short Quonset Point data record coincides with the period of largest mean offset seen at other stations. This explains why the subtidal bottom temperature skill is lower at Quonset Point than at other stations.

Figure 1.58. Time series of observed (black) and modeled (grey) subtidal bottom temperature at the southern stations, namely Bullock Reach, Conimicut Point, North Prudence, and Mount Hope Bay.

The misrepresentation of subtidal bottom temperature by the model is caused

by a nonlocal process, since the mean offset is found at the majority of stations. The model boundary forcing is a likely source for the inaccuracy. Snapshots of sea surface temperature from MODIS [83] give the measured surface temperature near the location of the model boundary where image pixels are not contaminated by land effects. The MODIS data is used to assess the validity of surface temperature boundary forcing, which in turn allows for inferences regarding the bottom temperature. A lateral mean temperature value is calculated along a line between The Narrows and Brenton Point, across the West and East Passages, for each day that all pixels along the line are usable. The total number of pixels averaged vary between four and eight, depending on the angle of the image. An average is similarly calculated from hourly temperature output at the model boundary across the West and East Passages. The comparison of MODIS and model sea surface temperature reveals that the model underestimates surface temperature on most imagery days that follow day 175 (Figure 1.59). During that time period the temperature mismatch between MODIS and the model is not due to inaccuracies at tidal time scales, since the temperature differences are larger than the modeled tidal tracer amplitudes. Prior to day 175 the surface temperature difference is relatively small and may be attributed to differences in surface temperature at tidal time scales.

The MODIS and NB-ROMS surface temperature comparison allows for inferences regarding the bottom temperature. The trend of the modeled instantaneous boundary temperature indicates that the MODIS versus NB-ROMS discrepancy exists at subtidal time scales for imagery generated after day 175 (Figure 1.59). Previous studies of temperature near the mouth of Narragansett Bay in Rhode Island Sound showed a relatively constant subtidal vertical temperature gradient during the summer months [84]. The observed temporal mean difference between the surface and bottom temperature at this location was about 5 °C during June to

Figure 1.59. Comparison of MODIS (red) and NB-ROMS (black) instantaneous surface temperature at the mouth of Narragansett Bay. Temperature is a lateral average across the mouths of the West Passage and East Passage. Black dots indicate the modeled surface temperature corresponding to the time of the satellite sea surface temperature imagery. The black line is the modeled hourly surface temperature. The blue line is the modeled hourly bottom temperature.

mid-August. The model predicts a relatively constant surface to bottom difference of 6 °C during those months in 2006 (Figure 1.59). If a constant temperature gradient is assumed, it is probable that the offset between observed and modeled surface temperature also exists for bottom temperature. It follows that the modeled subtidal bottom temperature is low compared to real conditions. The inaccuracy is translated to the northern parts of the NB-ROMS grid by the inflowing bottom water. The temporal mean difference between observed and modeled surface temperature at the boundary is 1.6 °C for the time period following day 175, which is consistent with data-model differences in subtidal bottom temperature at tracer stations after that day. Agreement between MODIS and the model prior to day 175 corresponds with accurately modeled subtidal bottom temperature at buoy stations before days 170 – 175.

The model predicts subtidal surface salinity with high accuracy, but a small disparity between the model and observations can be found. The model underestimates the subtidal surface salinity during a freshening event for a period of 25 to 30 days. The mismatch occurs between day 160 and day 185 at Conimicut Point, Mount View, Quonset Point, Poppasquash Point, and T-Wharf (Figures 1.60 and 1.61). At the Mount Hope Bay station the freshening event and data-model mismatch occurs between day 167 and day 185. Despite the mean offset during the freshening event the modeled subtidal surface salinity is highly correlated with observations (Table 1.4).

The fact that the mismatch between observed and modeled subtidal surface salinity occurs during a freshening event points to a deficiency in the freshwater runoff specifications. Two high river runoff events occurred in short succession during the validation time series (Figure 1.21). Model river forcing shows the peak of the first event on day 158 in the Blackstone and Pawtuxet rivers and on day

Figure 1.60. Time series of observed (black) and modeled (grey) subtidal surface salinity at the northern stations, namely Bullock Reach, Conimicut Point, Greenwich Bay Marina, and Mount Hope Bay.

Table 1.4. Skill values for subtidal surface salinity at each of the nine fixed-site stations. *WS* is the Willmott skill, *CC* is the correlation coefficient (statistically significant at the 95% confidence level), and *RMSE* is the root mean square error.

Station	BR	CP	NP	MV	QP	PP	TW	GB	MH
<i>WS</i>	0.88	0.96	0.95	0.88	0.75	0.93	0.83	0.79	0.96
<i>CC</i>	0.79	0.94	0.95	0.87	0.64	0.87	0.77	0.81	0.93
<i>RMSE</i> (psu)	2.27	1.64	0.85	1.51	1.53	1.69	1.70	2.30	1.80

Figure 1.61. Time series of modeled (grey) and observed (black) subtidal surface salinity at the southern stations, namely Mount View, Quonset Point, Poppasquash Point, and T-Wharf.

159 in the Taunton River (Figure 1.62). The second event peaked on day 176 in the Blackstone and Pawtuxet rivers and a day later in the Taunton River. After the two runoff events the river transport returns to mean annual flow rates near day 190 in rivers other than the Taunton River. The runoff peaks are seen to be wider and to persist several days longer in the Taunton River compared to other rivers.

Figure 1.62. River transport with estimated groundwater and ungauged surface runoff as forced in the NB-ROMS model. Rivers include the Mosshassuck and Woonasquatucket (red), Blackstone and Ten Mile (blue), Pawtuxet (magenta), Taunton (black), Hunt (purple), and Palmer (green).

The time period during which the model underestimates subtidal surface salinity corresponds with the occurrence of the two high runoff events. During times of high runoff the modeled subtidal surface salinity is lower than observed in the West Passage (excluding North Prudence where no data are available during this time) and East Passage, as well as at the Mount Hope Bay site. At the Mount

Hope Bay station the mismatch is seen to appear later in the time series. The model correctly captures the freshening associated with the first runoff event at this station. During times of mean or low river transport rates the model provides accurate predictions of the subtidal surface salinity at these stations.

The modeled subtidal surface salinity at Greenwich Bay Marina and Bullock Reach exhibits discrepancies that are not explained in the same way than the other stations. At the Greenwich Bay Marina station the model overestimates salinity for the majority of the time series, causing a larger mean square error than at the other stations (Table 1.4). Similar to the stations discussed thus far, the data show a freshening during the first part of the time series, followed by a gradual increase in salinity. The model predicts this trend, but it overestimates the mean subtidal surface salinity. Furthermore, the observations indicate a freshening event around day 212 that is not measured at any of the other stations; this freshening event is not captured by the model. The overestimated modeled subtidal surface salinity is consistent with findings regarding tidal salinity at Greenwich Bay Marina. The lack of river inflow to Greenwich Bay, as specified in the model, leads to higher surface salinity than is observed. The freshening event around day 212 may be due to runoff characteristics specific to Greenwich Bay, which is not represented by the model.

The Bullock Reach subtidal surface salinity observations show a dramatic decrease as a result of the high river runoff events. A similarly notable subtidal surface salinity decrease occurs at Conimicut Point during peak runoff. At Bullock Reach the minimum salinity due to the first event is 7.4 psu and the minimum salinity due to the second event is 6.8 psu. This is markedly lower than the proximate Conimicut Point station that has a minimum salinity of 13.2 psu and 14 psu during the first and second events, respectively. The large observed differ-

ence between these stations during the second runoff event is consistent with the relatively large surface salinity gradient shown in the previous section to explain tidal surface salinity discrepancies at Bullock Reach. The modeled subtidal surface salinity at Bullock Reach roughly correlates with the observations, but the root mean square error is relatively large (Table 1.4). As discussed in the previous section, the Bullock Reach station is subject to large volumes of freshwater and strong surface salinity gradients. The disagreement between modeled and observed subtidal surface salinity may be due to a misrepresentation of large surface salinity fronts in the region of this station during the flood.

Subtidal bottom salinity is modeled with high accuracy at a number of stations during the second half of the time series (Figure 1.63). The stations include Bullock Reach, Conimicut Point, and Mount Hope Bay. The model slightly underestimates the subtidal bottom salinity during the first 30 days of the time series at these stations. This period corresponds to the duration of high freshwater runoff events and is consistent with downward mixing of underestimated subtidal surface salinity. The model boundary assignment of bottom salinity could also contribute to the disparity. However, the mismatch is more pronounced at the northern and western stations than at stations in the East Passage (Figure 1.64). If the model boundary were the cause of the data-model difference, the effect would be largest at the southern East Passage stations where the inflowing bottom water most resemble the boundary conditions.

Modeled subtidal bottom salinity is not well correlated with the observations at Mount View, Quonset Point, Poppasquash Point, and T-Wharf (Table 1.5). The model predicts lower variance than is observed at these stations. At the Mount View and T-Wharf stations the model underestimates subtidal bottom salinity during the first 30 days of the time series in a similar way than stations in the

Figure 1.63. Time series of observed (black) and modeled (grey) subtidal bottom salinity at the northern stations, namely Bullock Reach, Conimicut Point, Greenwich Bay Marina, and Mount Hope Bay.

Table 1.5. Skill values for subtidal bottom salinity at each of the nine fixed-site stations. *WS* is the Willmott skill, *CC* is the correlation coefficient (statistically significant at the 95% confidence level), and *RMSE* is the root mean square error.

Station	BR	CP	NP	MV	QP	PP	TW	GB	MH
<i>WS</i>	0.88	0.78	0.76	0.70	0.54	0.60	0.68	0.93	0.92
<i>CC</i>	0.82	0.66	0.70	0.59	0.43	0.59	0.46	0.88	0.87
<i>RMSE</i> (psu)	0.97	0.81	0.64	1.01	0.67	0.95	0.75	1.01	0.82

north, but further data-model disagreement is seen for the remainder of the time series. The model does not predict an observed freshening of the bottom water at Poppasquash Point between day 177 and day 210 and thus overestimates subtidal bottom salinity during that time period.

The results of this section show that subtidal tracers are predicted with high accuracy by the model, but causes for minor model-data discrepancies are identified. Correlation between observed and modeled subtidal bottom temperature is high, but discrepancies are manifested as a mean offset for the majority of the time series. The relatively cool inflowing bottom water specified at the boundary translates to underestimated subtidal bottom temperatures at the buoy sites. The model underestimates the subtidal surface salinity during periods of relatively extreme river runoff, but it performs well in predicting subtidal surface salinity for times outside the peak runoff events. At the Greenwich Bay Marina site the model overestimates subtidal surface salinity due to a lack of freshwater sources at the head of Greenwich Bay. The disagreement between modeled and observed subtidal surface salinity at Bullock Reach may be due to a misrepresentation of large surface salinity fronts at this station during the freshwater flood. At the majority of the stations the subtidal bottom salinity is underestimated during the large runoff events, which indicates the downward mixing of underestimated subtidal surface salinity in the model.

Figure 1.64. Time series of observed (black) and modeled (grey) subtidal bottom salinity at the northern stations, namely Mount View, Quonset Point, Poppasquash Point, and T-Wharf.

1.3.3 Turbulence closure scheme comparison

The skill assessment of Section 1.3.1 indicates the relative accuracy of models that implement different turbulence closure schemes. The skill analysis reveals minor differences in the accuracy of the GLS and MY25 schemes. The model that implements the LMD scheme produces markedly different skills than the other schemes and generally shows poor agreement with the observations. The largest difference in the performance of closure models occurs for properties at subtidal time scales (Figure 1.7). The differences in skill values of the LMD model relative to the other models are generally much larger for subtidal properties than tidal properties. The causes for these differences are assessed in this section.

Tracers

Instantaneous profiles of temperature and salinity at shallow locations of Narragansett Bay clearly show differences in the mixing characteristics of turbulence closure models. Observed profiles of temperature and salinity were obtained at Rumstick South and Warwick Neck from day 177 to day 194. The start of this time period coincides with the maximum river transport of the second large runoff event. The river influence is clearly discernible at Rumstick South, which is located approximately 1.70 km directly south of the Palmer/Warren River mouth (Figures 1.65). A relatively fresh surface and weakening halocline is observed at the start of the time series, followed by a brief restratification period. A period of freshening and relatively strong mixing is observed around day 190. The salinity profiles produced by the $k - \varepsilon$ G88 and MY25 models closely resemble the data, except for a slightly more saline water column towards the end of the time series. The LMD model exhibits similar mixing and restratification patterns than observed, but the surface salinity is markedly lower and more uniform in time, and the halocline is stronger than observed. During periods of restratification the near-bottom water

has salinity values that resemble the observations.

A similar data-model discrepancy can be seen in comparisons of the temperature profiles (Figures 1.65). A pulse of relatively cool and saline bottom water between day 185 and 190 is predicted by the $k - \varepsilon$ G88 and MY25 models to be somewhat cooler and more pervasive in the water column than observed. The models generally predict lower bottom temperature, which is consistent with the finding that boundary forced bottom temperature is underestimated. The LMD model predicts higher surface temperatures and lower bottom temperatures than observed, leading to a stronger and persisting thermocline.

Comparisons of modeled and observed hydrographic variables at Warwick Neck reveal similar discrepancies than at Rumstick South (Figure 1.66). At this location the overestimated diurnal surface heating caused by model specifications of the surface boundary is apparent. The LMD surface heating and bottom cooling is extreme, causing a relatively strong thermocline with the duration of the time series. Temperature records clearly show that the observed water column is more frequently vertically well-mixed than predicted.

Observed profiles of instantaneous temperature and salinity from Rumstick North and Bullock Reach show variability forced by environmental conditions, which is predicted with different degrees of success by the models. The period of data collection coincides with a few low rainfall events, decreasing atmospheric heating, and relatively sustained wind events. Two southwesterly wind events, one shortly before day 220 and the other shortly after day 220, cause efficient vertical mixing throughout the water column at Rumstick North (Figure 1.67). Observed salinity profiles suggest that the water column is weakly restratified during a northerly wind event that separates the two southwesterly events. A sustained northwesterly toward the end of the time series causes relatively strong

Figure 1.65. Data-model comparisons of Rumstick South (RS) instantaneous salinity and temperature profiles; models are $k - \varepsilon$ G88, MY25, and LMD.

Figure 1.66. Data-model comparisons of Warwick Neck (WN) instantaneous salinity and temperature profiles; models are $k - \varepsilon$ G88, MY25, and LMD.

vertical mixing of relatively saline and cool water. The $k - \varepsilon$ G88 and MY25 models slightly overestimate the salinity during periods of vertically well mixed water at Rumstick North. These models underestimate the cool bottom water flux associated with the northerly wind event between day 222 and 225. The LMD scheme predicts the patterns of mixing and restratification that are observed, but the model underestimates the surface salinity and bottom temperature. A strong halocline and thermocline is predicted during periods of stratified water. Observed profiles at the shallow Bullock Reach site show a strong tidal vertical mixing and restratification signal. All models capture the tidal vertical mixing characteristics (Figure 1.68), although a stronger pycnocline is predicted by the LMD scheme. The LMD model underestimates the surface salinity and bottom temperature.

Figure 1.67. Data-model comparisons of Rumstick North (RN) instantaneous salinity and temperature profiles; models are $k - \varepsilon$ G88, MY25, and LMD.

The differences in vertical mixing of the three turbulence closure models at the profiler stations can be inferred from profiles of vertical turbulent diffusion

Figure 1.68. Data-model comparisons of Bullock Reach (BR) instantaneous salinity and temperature profiles; models are $k - \varepsilon$ G88, MY25, and LMD.

coefficients for tracers (Figure 1.69). At the Bullock Reach, Rumstick South, and Warwick Neck stations the $k - \varepsilon$ G88 and MY25 models produce fairly uniform vertical mixing in the interior of the water column. The $k - \varepsilon$ G88 gives marginally higher mixing at Rumstick South and Warwick Neck. At Rumstick North these models predict somewhat higher mixing in the surface boundary layer than the bottom boundary layer with $k - \varepsilon$ G88 predicting higher bottom mixing than MY25, and MY25 producing higher surface mixing than $k - \varepsilon$ G88. The LMD model predicts that the vertical mixing of tracers is much larger in the surface boundary layer than in the rest of the water column. The surface boundary layer mixing predicted by the LMD model is very large compared to the other models, and the bottom boundary layer mixing is slightly larger than other models at Rumstick South and Warwick Neck.

Observed temporal mean profiles of salinity and temperature are more ac-

curately represented by the $k - \varepsilon$ G88 and MY25 models than the LMD model (Figure 1.69). The LMD model predicts near-bottom salinity that is similar to the observations, but it completely underestimates salinity of the upper water column. Temporal mean salinity stratification is thus much higher in the LMD model than observed. The observed mean temperature profiles show a relatively uniform water column (Figure 1.69). The $k - \varepsilon$ G88 and MY25 models predict slightly higher mean temperature stratification than observed, due to lower near-bottom temperatures. The LMD model predicts relatively high surface temperatures and low bottom temperatures, such that the temperature stratification is much higher than observed.

Figure 1.69. Temporal mean profiles of the vertical turbulent diffusion coefficient for tracers (top), salinity (middle), and temperature (bottom) at the four profiler stations. The models shown are $k - \varepsilon$ G88 (black), MY25 (red), and LMD (blue). Averages correspond to observation periods.

Inaccurate specifications of the NB-ROMS model forcing, which lead to mis-

matches between modeled and observed hydrography, do not relate to inadequacies regarding the vertical mixing (Section 1.3.2) and are therefore expected to cause discrepancies in all the models, regardless of the applied turbulence closure scheme. This is confirmed by the comparisons of observed profiler time series with output from the $k - \varepsilon$ G88 and MY25 models. The lower than observed bottom temperature specified at the model boundary leads to underestimated bottom temperatures at the profiler stations by both of these models. Agreement of the $k - \varepsilon$ G88 and MY25 models with observed surface temperature, together with the accurate representation of the vertical salinity gradient, show that vertical mixing is fairly accurately modeled. The slightly higher than observed vertical temperature stratification produced by $k - \varepsilon$ G88 and MY25 is thus due to vertical mixing of bottom water that is cooler than observed and not due to an inaccuracy in the representation of vertical turbulent diffusion. It can be assumed that the vertical diffusion coefficients of the $k - \varepsilon$ G88 and MY25 models (Figure 1.69) resemble those of the observations.

The LMD model exhibits problems in reproducing observed profiler hydrography that relate to both the model forcing specifications and the vertical mixing characteristics. Combined with forcing that specifies an excess of freshwater during flood events and underestimated bottom temperature (Section 1.3.2), the vertical turbulent diffusion coefficient of the LMD scheme indicates high mixing in the surface boundary layer, low mixing across the pycnocline, and high mixing in the bottom boundary layer relative to the other schemes. These combined factors produce excessively low surface salinity and bottom temperature. The vertical mixing structure causes freshwater (warm water) to mix efficiently in the surface boundary layer and saline water (cold water) to mix efficiently in the bottom boundary layer, but mixing across the pycnocline is suppressed. The difference between tracer ob-

servations and LMD output is relatively large, which suggests that the mismatch between observed and modeled vertical diffusion coefficients is large.

The parameters that may contribute to the artificial separation of the surface boundary and interior, as seen in output of the vertical turbulent diffusion coefficients and tracer profiles, can be identified by considering the formulation of the LMD scheme. The local vertical diffusion coefficients of the surface boundary layer are calculated as the product of the surface boundary layer depth, a turbulent velocity scale, and a cubic vertical shape function (equation A.25). The cubic shape function is designed such that vertical diffusion coefficients of the surface boundary layer fit smoothly with those of the interior. Vertical diffusion coefficients of the interior are dependent on the gradient Richardson number (see Appendix A). The vertical structure of diffusion coefficients of the surface boundary are largely determined by the shape function for the following reasons [30]: only the local component of the surface tracer flux is relevant, since the forcing conditions are generally stable; the local vertical diffusivity depends on a depth-independent turbulent velocity scale, since the boundary layer is generally stable or neutral; and the boundary layer depth is determined from the bulk Richardson number. Note that the bulk Richardson number may generally be higher in the LMD scenario run than observed, due to overestimated freshwater runoff – this would lead to an underestimation of the boundary layer depth.

The fact that the LMD scheme employs a separate treatment of the surface boundary layer and the interior when calculating vertical turbulent diffusivities may cause its inaccuracy in reproducing profiler time series. This treatment is inappropriate for the partially- to well-mixed, shallow water columns of Narragansett Bay. The relatively high maximum mean vertical diffusivity in the surface boundary layer of the LMD model may be due to an excessively high turbulent

velocity scale, resulting from an overestimated friction velocity, or an underestimated stability parameter. Alternatively, it may be due to the fitting procedure of the cubic shape function where polynomial coefficients are chosen to match the interior diffusion coefficients at the bottom of the surface boundary layer and the Monin-Obukhov similarity theory at the surface.

The performance of models can further be assessed by considering the predictions of subtidal surface and bottom tracers at the buoy stations. In general, the largest discrepancies between observations and models occur when subtidal tracers are modeled by the LMD closure scheme. The LMD scheme performs well in calculating subtidal surface temperature, but it inaccurately predicts subtidal surface and bottom salinity, and bottom temperature. The difference between LMD subtidal output and observations is evident in scatter diagrams that compare output of all stations (Figure 1.70). The most prominent model errors are large mean offsets in subtidal surface salinity and bottom temperature. The LMD model underestimates the subtidal surface salinity by 5 psu – 12 psu at average and low salinity values. In addition, the model underestimates bottom temperature by 4 °C – 7 °C. Better agreement is predicted at the lowest temperatures.

Time series of subtidal bottom temperature at a number of representative stations show good agreement between the observations and models near the beginning of the time series, but the accuracy of models degrade after day 170 when predicted temperatures become lower than observed (Figure 1.71). This is consistent with results of Section 1.3.2 showing that the forced boundary bottom temperature is underestimated after day 170. The LMD model predicts markedly lower subtidal bottom temperatures than the other models. Similar to conclusions drawn from the profiler comparisons, the relatively low bottom temperature of the LMD model is caused by inadequate mixing of surface and bottom water across

Figure 1.70. Comparisons of modeled versus observed surface (top row) and bottom (bottom row) subtidal tracers for all station output. The model implements the LMD turbulence closure scheme.

the pycnocline. At Greenwich Bay Marina the $k - \varepsilon$ G88 and MY25 models provide accurate predictions of subtidal bottom temperature, but the LMD model underestimates the subtidal bottom temperature. This indicates that the effect of low bottom temperature forcing is diminished at the Greenwich Bay Marina site when $k - \varepsilon$ G88 and MY25 are implemented due to mixing of bottom water with surface water. Since mixing across the pycnocline is suppressed in the LMD model, the effect of forced low bottom temperatures persists even after it has been advected to the head of Greenwich Bay (Figure 1.74). The root mean square error of subtidal bottom temperature improves by 52% between Poppasquash Point and Greenwich Bay Marina in the $k - \varepsilon$ G88 model, but it improves only by 20% between these stations in the LMD model.

Figure 1.71. Time series of observed (black) subtidal bottom temperature as compared to the $k - \varepsilon$ G88 model (grey), MY25 model (red), and the LMD model (blue) at the Bullock Reach, Mount View, Greenwich bay Marina, and Mount Hope Bay stations.

Time series of subtidal surface salinity at a few representative buoy stations show the mismatch between models and observations (Figure 1.72). The $k - \varepsilon$ G88 and MY25 models produce accurate subtidal surface salinity during the early and later parts of the time period. These models underestimate subtidal surface salinity during the high freshwater runoff events due to model forcing of excessive freshwater during that time (see Section 1.3.2). The LMD model generally produces accurate subtidal surface salinity for a short duration at the beginning of the time series. For the majority of the data record the LMD model predicts subtidal surface salinity that correlates with the observations, but is severely underestimated. A slight improvement can be discerned toward the end of the record. The inaccurate surface salinity of the LMD model after the onset of high river runoff is consistent with overestimated vertical mixing of the surface boundary and suppression of mixing across the pycnocline.

Representative time series show the differences in subtidal bottom salinity between the models and observations (Figure 1.73). The $k - \varepsilon$ G88 and MY25 models slightly underestimate the subtidal bottom salinity at the majority of stations during the time of high freshwater runoff conditions, as discussed in Section 1.3.2. The underestimated surface salinity is vertically mixed to produce relatively low bottom salinity in these models. The LMD scheme predicts inhibited mixing across the pycnocline, such that the underestimated surface salinity generally does not cause lower than observed bottom subtidal salinity. In general, the LMD model predicts slightly higher subtidal bottom salinity than observed, indicating that vertical mixing is weaker than observed.

The predicted vertical subtidal salinity structure at the Greenwich Bay Marina site is different from the other stations. The $k - \varepsilon$ G88 and MY25 models slightly overestimate subtidal surface salinity due to the lack of freshwater sources

Figure 1.72. Time series of observed (black) subtidal surface salinity as compared to the $k - \varepsilon$ G88 model (grey), MY25 model (red), and the LMD model (blue) at the Bullock Reach, Mount View, Greenwich Bay Marina, and Mount Hope Bay stations.

Figure 1.73. Time series of observed (black) subtidal bottom salinity as compared to the $k - \varepsilon$ G88 model (grey), MY25 model (red), and the LMD model (blue) at the Bullock Reach, Mount View, Greenwich Bay Marina, and Mount Hope Bay stations.

in Greenwich Bay (see Section 1.3.2), but the LMD model underestimates subtidal surface salinity (Figure 1.72). It shows that the excess in forced freshwater is very weakly mixed across the pycnocline in the LMD model, such that the low surface salinity advects to the head of Greenwich Bay where no freshwater sources are specified (Figure 1.74). The discussion on the subtidal bottom salinity as predicted by the $k - \varepsilon$ G88 and MY25 models at other stations also applies at the Greenwich Bay Marina site, but the LMD output of subtidal bottom salinity at this location shows different trends from the other sites. The latter model predicts a good match during the time of high freshwater runoff conditions prior to day 170, but it underestimates subtidal bottom salinity for the remainder of the time period. Underestimated subtidal bottom salinity is consistent with the LMD prediction of a relatively deep surface boundary layer for salinity in Greenwich Bay, such that the bottom conductivity sensor at the Greenwich Bay Marina site, at the head of the bay, is modeled to be within the surface boundary layer (Figure 1.74).

Figure 1.74. Temporally and laterally averaged sections of temperature and salinity from the head to the mouth of Greenwich Bay, as predicted by the $k - \varepsilon$ G88 (top panels) and LMD (bottom panels) models.

Temporally averaged density sections at various locations along the longitudinal axis of Narragansett Bay show key differences between the $k - \varepsilon$ G88 and LMD

models (Figure 1.75). The general tendency of the LMD model to predict well-mixed surface and bottom water columns and a strong pycnocline can be discerned in all sections. The LMD density sections show that high bottom density and low surface density are of similar magnitudes in the deep East Passage Channel and the shallower West Passage Channel. The strong vertical density gradient persists into the Providence River. In the $k - \varepsilon$ G88 model the relatively higher bottom density of the East Passage Channel compared to the West Passage Channel is well represented. The $k - \varepsilon$ G88 model predicts the gradual mixing of surface freshwater from the head of the Bay southward.

Figure 1.75. West to east sections of density, averaged over the model validation period, as predicted by the $k - \varepsilon$ G88 (left) and LMD (right) models. Sections are along the West and East Passages north of Hope Island (top), along Ohio Ledge at the mouth of the Warren/Palmer River (middle), and south of the Pawtuxet River mouth in the Providence River (bottom).

Velocity

The performance of models in capturing subtidal velocity vary based on the velocity component and station in question. The LMD model offers improvements over the other models for the subtidal surface east-west component at the West Passage and East Passage Shoal sites and the subtidal surface north-south velocity at the East Passage Channel station. The two-equation models perform similarly in calculating velocity components, except in the case of subtidal surface east-west velocity at the East Passage Shoal station.

Figure 1.76. Time series of observed (black) subtidal surface east-west velocity as compared to the GLS and MY25 models (grey), and the LMD model (blue) at the three ADCP stations.

The models exhibit varying success in representing subtidal surface east-west velocity at the ADCP stations. Time series of subtidal surface east-west velocity (Figure 1.76) and statistical measures of model performance (Table 1.6) show that the relative high skills of the LMD model at the West Passage and East Passage Shoal stations are attributable to higher correlations with the observations than

Table 1.6. Skill values for subtidal surface east-west velocity at the West Pasasage (WP), East Passage Channel (EPc), and East Passage Shoal (EPs) stations. *WS* is the Willmott skill, *CC* is the correlation coefficient (statistically significant at the 95% confidence level), and *RMSE* is the root mean square error.

Model skill	Station		
<i>WS:</i>	WP	EPc	EPs
k-e G88	0.63	0.55	0.50
MY25	0.65	0.52	0.50
LMD	0.72	0.47	0.83
<i>CC:</i>			
k-e G88	0.53	0.29	0.30
MY25	0.52	0.24	0.26
LMD	0.64	0.12	0.75
<i>RMSE (m/s):</i>			
k-e G88	0.03	0.01	0.02
MY25	0.03	0.01	0.02
LMD	0.03	0.01	0.01

the other models. At the East Passage Shoal station the LMD output also produces a lower root mean square error when compared with data. The difference in performance between the GLS and MY25 models in calculating subtidal surface east-west velocity at the East Passage Shoal station (Figure 1.17) is due to a disparate representation of the magnitude of one westward peak centered on day 219 of the relatively short observational period. This peak is the result of a shift in wind conditions from southwesterly to predominantly northerly (Figure 1.20). Each model represents the peak at slightly different magnitudes. The discrepancy in the modeled peak magnitude is 0.03 m/s between the GLS and MY25 models.

The subtidal surface north-south velocity is generally predicted with high accuracy ($WS \geq 0.65$) by all the models. Subtidal surface north-south velocity time series (Figure 1.77) and statistics (Table 1.7) show that at the East Passage Channel station the LMD model provides a better correlation with observations and produces a lower root mean square error than the other models. In contrast, sub-

Figure 1.77. Time series of observed (black) subtidal surface north-south velocity as compared to the GLS and MY25 models (grey), and the LMD model (blue) at the three ADCP stations.

tidal surface north-south velocity at the East Passage Shoal station is inadequately represented ($WS < 0.65$) by the LMD model both as a result of low correlation and a high root mean square error with the data.

The GLS and MY25 models provide accurate predictions ($WS > 0.65$) of the subtidal bottom north-south velocity, but the LMD model performs poorly ($WS < 0.65$) in predicting this variable. Time series comparisons (Figure 1.78) and statistical values (Table 1.8) show that the low Willmott skill of the LMD model at the East Passage Channel station is due to a high root mean square error, despite a high correlation with the observations. The observations show a predominantly northward surface flow with isolated events of flow reversal, but the LMD model predicts predominantly southward flow, particularly during the first half of the time series. At the West Passage and East Passage Shoal stations the LMD model produces the same root mean square errors than the other models,

Table 1.7. Skill values for subtidal surface north-south velocity at the West Pasasage (WP), East Passage Channel (EPc), and East Passage Shoal (EPs) stations. *WS* is the Willmott skill, *CC* is the correlation coefficient (statistically significant at the 95% confidence level), and *RMSE* is the root mean square error.

Model skill	Station		
WS:	WP	EPc	EPs
k-e G88	0.74	0.65	0.85
MY25	0.72	0.65	0.86
LMD	0.72	0.77	0.54
CC:			
k-e G88	0.59	0.62	0.76
MY25	0.57	0.64	0.77
LMD	0.60	0.86	0.46
RMSE m/s:			
k-e G88	0.02	0.06	0.02
MY25	0.02	0.06	0.02
LMD	0.03	0.05	0.04

Figure 1.78. Time series of observed (black) subtidal bottom north-south velocity as compared to the GLS and MY25 models (grey), and the LMD model (blue) at the three ADCP stations.

Table 1.8. Skill values for subtidal bottom north-south velocity at the West Pasasage (WP), East Passage Channel (EPc), and East Passage Shoal (EPs) stations. *WS* is the Willmott skill, *CC* is the correlation coefficient (statistically significant at the 95% confidence level), and *RMSE* is the root mean square error.

Model skill	Station		
<i>WS:</i>	WP	EPc	EPs
k-e G88	0.70	0.91	0.74
MY25	0.70	0.90	0.78
LMD	0.46	0.57	0.63
<i>CC:</i>			
k-e G88	0.60	0.88	0.74
MY25	0.63	0.87	0.77
LMD	0.21	0.79	0.52
<i>RMSE</i> m/s:			
k-e G88	0.03	0.03	0.03
MY25	0.03	0.03	0.03
LMD	0.03	0.08	0.03

but the correlations between LMD output and the observations are low at both of these stations.

The observed temporal mean vertical profiles of subtidal velocity components at the West Passage station are not captured well by the models (Figure 1.79). Observations exhibit westward flow near the surface that decreases in magnitude with depth, reverses very slightly near the middle of the water column, and increases again westward to the near-bottom. The $k - \varepsilon$ G88 and MY25 models predict two-layered flow that is westward in the upper water column and eastward in the lower water column. The LMD model overestimates the magnitude of the westward surface flow and the mid-column reversal. The LMD model underestimates the magnitude of the bottom flow. The north-south velocity component is observed to be two-layered at the West Passage station with northward flow in the upper third of the water column and southward flow that increases in magnitude with depth below that. None of the models correctly predict the vertical

structure of the flow, but the LMD model approximately captures the magnitude of the surface and bottom flow. The LMD model produces a maximum northward flow in the middle of the water column. The $k - \varepsilon$ G88 and MY25 models predict three-layered flow with the surface flowing southward.

At the East Passage Channel station the models capture portions of the temporal mean vertical profiles of subtidal velocity (Figure 1.79). The observed mean east-west velocity component at the East Passage Channel station exhibits relatively low magnitude vertically. The data reveal a weak eastward near-surface flow and an even weaker westward flow in the majority of the water column below the surface layer. The models overestimate the magnitude of the flow in the bulk of the water column. The LMD model predicts weak eastward flow through the entire depth. The $k - \varepsilon$ G88 and MY25 models predict three-layered east-west flow at the East Passage Channel station with the middle layer resembling the observed flow. The observed north-south velocity component at the East Passage Channel station is southward in a thin layer near the surface and relatively strongly northward throughout the remainder of the water column. All three models predict northward flow maxima near the middle of the water column and decreased magnitudes toward the bottom, whereas the observed flow is vertically uniform from the middle of the water column to the bottom. The LMD model predicts a flow reversal near the bottom.

Output of temporal mean vertical flow structure at the East Passage Shoal station from the $k - \varepsilon$ G88 and MY25 models is qualitatively similar to the observations, but output from the LMD model differs from the data (Figure 1.79). The east-west velocity component predicted by the $k - \varepsilon$ G88 and MY25 models show offsets in magnitude. The observed north-south velocity component is southward near the surface and in the middle of the water column, whereas the near-bottom

is northward. The $k - \varepsilon$ G88 and MY25 models roughly predict the vertical flow structure, but estimate flow reversal near the middle of the water column. The LMD model predicts a relatively weak, vertically uniform northward flow.

Figure 1.79. Temporal mean (day 150 - day 227) vertical profiles of the subtidal east-west velocity (top) and north-south velocity (bottom) at the three ADCP stations. Positive velocity indicates eastward and northward flow. Observations (grey) are compared with the following models: $k - \varepsilon$ G88 (black), MY25 (red), and LMD (blue).

Temporally averaged north-south velocity sections at various locations along the longitudinal axis of Narragansett Bay show differences in lateral flow structure and magnitude between the $k - \varepsilon$ G88 and LMD models (Figure 1.80). The LMD model produces higher lateral variance in the deep East Passage Channel and exhibits relatively strong southward flow along the western wall of the channel. Southward flow in the western portion of the West Passage is diminished in the LMD model. On Ohio Ledge the LMD model predicts lower northward channel flow at depth, deeper southward near-surface flow, and a higher near-surface

recirculation to the north. Northward channel flow and southward near-surface flow in the Providence River are relatively weak in the LMD model.

Figure 1.80. West to east sections of north-south velocity, averaged over the model validation period, as predicted by the $k - \varepsilon$ G88 (left) and LMD (right) models. Positive velocity is northward. Sections are along the West and East Passages north of Hope Island (top), along Ohio Ledge at the mouth of the Warren/Palmer River (middle), and south of the Pawtuxet River mouth in the Providence River (bottom).

1.4 Conclusion

Our validation work shows that the NB-ROMS model provides a good representation of circulation and hydrography in Narragansett Bay. Vertical mixing is appropriately represented by the $k - \varepsilon$ G88 statistical two-equation turbulence closure scheme. Comparisons of turbulence closure schemes demonstrate that the GLS implementation of two-equation closure models and the original implementation of MY25 in ROMS afford analogous predictions of Narragansett Bay properties. This is consistent with studies elsewhere that found only minor differences

between these schemes [85, 32, 15]. The LMD empirical turbulence closure scheme is unsuitable for predictions of vertical turbulent mixing in Narragansett Bay.

Misrepresentation of vertical turbulent mixing in Narragansett Bay by the LMD model leads to large errors in its predictions of tracers and velocity at all time scales. Vertical mixing produced by the LMD scheme exhibits excessively mixed surface and bottom layers and a strong pycnocline that inhibits mixing between these layers through the majority of the model domain. The vertical mixing characteristics augment some minor inaccuracies in the specified model forcing, such that subtidal upper water column salinity and lower water column temperature are greatly underestimated. The LMD model predicts subtidal vertical and lateral velocity structures that do not match the observations. These inaccuracies are expected to negatively affect constituent transport in the model. The fact that the LMD scheme employs different strategies to calculate the vertical turbulent coefficients of the surface boundary layer and the interior seems to cause the erroneous representation of properties in the partially- to well-mixed Narragansett Bay.

Comparisons of observations and output from the NB-ROMS model, which uses the $k - \varepsilon$ G88 turbulence closure scheme, show that the model accurately predicts variables that are essential to correctly represent transport and exchange characteristics. The model performs very well in predicting sea surface elevation and velocity at tidal time scales. At subtidal time scales the model accurately reproduce velocity, temperature, and salinity.

Small discrepancies between the NB-ROMS model and observations are related to forcing specifications and are not attributed to the $k - \varepsilon$ G88 vertical mixing parameterization. The discrepancies are identified in the validation analysis to enable model optimization. For the purpose of model optimization, the tidal

temperature, salinity, and subtidal sea surface elevation require improvement.

Inaccuracies in modeled tidal temperature and salinity relate to the NB-ROMS model representation of tidal constituents, in addition to diurnal atmospheric heating and cooling in the case of temperature, and horizontal gradients in the case of salinity. Small miscalculations in the model predictions of tides and diurnal atmospheric processes cause errors in tidal salinity and temperature, since these perturbations are small. A slight overestimation of the fluctuations of atmospheric heating and cooling in the model causes a mismatch with observed tidal surface temperature at diurnal time scales. The model lacks freshwater sources in Greenwich Bay, which leads to low subestuarine longitudinal salinity gradients and severely underestimated variance in tidal surface and bottom salinity at this location. Sites that are proximate to applied freshwater runoff sources exhibit high observed tidal surface salinity variance. The model underestimates this variance during high runoff events when an excess of freshwater causes reduced horizontal salinity gradients. Inadequacies in the model prediction of tidal constituents include an underestimation of the M2 tidal constituent amplitude by 10% and a 70% difference between the observed and modeled phase of the M4 overtide. Amplification of the M6 overtide from the mouth to the head of the estuary is twice as high in the model than the data. The modeled phase transition of the M6 overtide, as it propagates along the estuary, is not consistent with the observations.

The NB-ROMS model predictions of tidal temperature and salinity can be improved in a number of ways. Diurnal surface heating and cooling can be improved by modifying shortwave and longwave radiation fluxes used in the atmospheric bulk flux parameterization. Specifying freshwater sources in Greenwich Bay and reduced runoff during flood events would improve horizontal salinity gradients on tidal time scales. Tidal constituent predictions may be improved by adjusting

the overtide forcing at the model boundary, specifying a spatially varying bottom roughness length, and reducing smoothness of the channel bathymetry.

The NB-ROMS model underestimates subtidal sea surface elevation variance throughout the model domain, as well as the subtidal sea surface elevation gradient along the longitudinal axis of Narragansett Bay. The latter leads to somewhat lower subtidal barotropic current variance than the observations. These issues may be related to the specification of subtidal sea surface elevation and velocity at the model boundary that coincides with the mouth of Narragansett Bay. The NB-ROMS boundary specifications are obtained from a Narragansett Bay model whose domain extends onto the continental shelf [7]. The latter model is driven by spatially uniform wind derived from observations along the coast of Narragansett Bay. A previous study showed that wind magnitudes on the shelf are twice as high as those inside Narragansett Bay [75]. Inadequate wind stress on the shelf of the large domain model could cause an underestimation of the Ekman processes on the shelf, which would lower the variance of subtidal sea surface elevation at the mouth and the estuarine longitudinal sea surface elevation gradient. Applying spatially varying wind that corresponds to observations across the large domain model may increase the accuracy of subtidal sea surface elevation and the longitudinal subtidal sea surface elevation gradient in the NB-ROMS model. The lack in forcing of atmospheric pressure in the model could contribute to the low variance of subtidal sea surface elevation throughout the bay.

Despite accurate predictions of subtidal temperature and salinity by the NB-ROMS model, a few modifications would allow further optimization. Increasing the temperature at the model boundary would improve the subtidal bottom temperature match throughout the model domain. Reducing freshwater runoff during flood events would increase the accuracy of subtidal salinity and the horizontal

salinity gradients during those events. Introducing freshwater runoff in Greenwich Bay would correct subtidal surface salinity at the head of the subestuary.

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MANUSCRIPT 2

Observations of the physical processes that influence water quality of
Greenwich Bay and upper Narragansett Bay

by

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is in preparation for submission to *Estuaries and Coasts*

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Abstract

The physical characteristics of subsystems in Narragansett Bay are assessed using velocity, temperature, and salinity observations with high spatial and temporal resolution. Greenwich Bay is a subestuary of Narragansett Bay that is subject to severe water quality degradation during the summer months. Prolonged dissolved oxygen deficiency caused by bacterial decomposition of phytoplankton blooms is particularly damaging to its ecosystem. Observations are used to develop relationships between environmental forcing, circulation and flushing, and water quality in Greenwich Bay. Physical processes can cause a reduction in dissolved oxygen concentration by isolating the lower water column from the atmosphere, or by inhibiting exchange with ventilated water. Results suggest that prevailing wind can lead to flow structures that explain oxygen resupply via advection, or oxygen depletion due to water retention. Dominant summer wind conditions cause a reduction in exchange between the inner and outer basins of Greenwich Bay. Generally low vertical turbulent mixing, owing to low tidal energy, and low subtidal currents suggest prolonged isolation of bottom water from the atmosphere. Diurnal wind produces enhanced vertical turbulent mixing, which can alleviate low oxygen conditions. The interaction of the diurnal wind with the semidiurnal tide produces intratidal residuals. Results show the importance of coupling between Greenwich Bay internal flow and currents in the Warwick Neck channel, and highlight the role of this channel as artery for exchange between pollution sources from the north and Greenwich Bay.

2.1 Introduction

Greenwich Bay is an urban subestuary of Narragansett Bay that is particularly susceptible to water quality degradation during the summer months. The decline of water quality in the highly populated northern regions of Narragansett Bay has been ascribed to bacterial contamination, metal pollution, and eutrophication resulting from excessive nutrient loading [1, 2]. The most damaging to Narragansett Bay ecosystems is prolonged dissolved oxygen deficiency caused by bacterial decomposition of organic matter produced by phytoplankton blooms [3, 4, 5]. A high incidence of hypoxia has been observed in Greenwich Bay [6, 7, 8, 9, 10]. A study found that Greenwich Bay was more severely hypoxic than other regions of Narragansett Bay between 2004 and 2006 [6]. It has also been suggested that Greenwich Bay is one of the dominant sources of hypoxic water in Narragansett Bay [11]. A general trend of decreasing dissolved oxygen from east to west in Greenwich Bay has been documented [12, 9] suggesting spatial heterogeneity in processes relating to water quality. Recurrent fish and benthic invertebrate mortalities in Greenwich Bay are attributed to hypoxia and anoxia [3, 13]. Negative effects of long term critically low dissolved oxygen in this subsystem include reduced ecosystem diversity and degraded benthic communities [3, 7]. In addition to hypoxia, summer beach closures resulting from high levels of bacteria is a consequence of poor water quality in Greenwich Bay.

Physical forcing functions can contribute to water quality in a number of different ways. There is a range of processes leading to hypoxia in different regions of Narragansett Bay [6]. Depletion of dissolved oxygen can occur by physical mechanisms, such as high water temperatures that lower the capacity of the water to retain dissolved oxygen during the summer [3, 7] and high vertical density stratification that can lead to an isolation of bottom water and reduce the potential for

replenishment by atmospheric oxygen [14, 15, 16, 17, 18]. An increase in stratification during weak neap tides contributes to bottom water hypoxia in the vicinity of Ohio Ledge 2.1 in Narragansett Bay [16]. Stratification caused by summer freshwater runoff correlates well with the incidence of hypoxia in the same region [6]. In addition to hypoxia, water quality impairment can constitute bacterial contamination. A physical process that causes bacterial contamination is storm-related freshwater runoff that introduces large quantities of untreated water into urban systems. Two physical mechanisms relating to water quality involves horizontal advection. Firstly, a subsystem can experience reductions in dissolved oxygen or increases in bacteria when exchange with adjacent oxygen-depleted or bacteria-rich systems occurs. A subsystem can become more susceptible to hypoxia when water with high concentrations of nutrients are introduced from adjacent systems. Secondly, oxygen replenishment or the removal of bacteria or nutrients are hindered when exchange with adjacent ventilated or healthy water is prevented. The latter can occur due to low current speeds or due to retentive flow structures, such as gyres. Such flow structures explain limited exchange between Greenwich Bay and Narragansett Bay during periods of northward wind [19].

Physical forces operate at different spatial and temporal scales to determine the dispersion of water and its constituents. Dispersion is here defined as the spreading of constituents through the water by all physical mechanisms [20]. These mechanisms occur both at intratidal and subtidal time scales and include vertical turbulent mixing, advection, and shear dispersion. Intratidal time scales are defined to occur at periods of 3 – 25 hours and subtidal time scales include all periods exceeding 25 hours.

Intratidal estuarine variability is largely a function of circulation that is forced by the tidal constituents at diurnal and lower periods. The interaction of these

tidal currents with density stratification and topography can lead to complex vertical and lateral variability. Nonlinear tidal variability can be categorized as both intratidal and subtidal. Tidal inequality occurs as the superposition of the semidiurnal and diurnal tidal constituents [21, 22, 23, 24]. Wind variability can occur at time scales of a day or less, such as during cycles of sea- and land breezes [25, 17]. This diurnal wind variability, here referred to simply as diurnal wind, will be shown to have an important influence on Narragansett Bay mixing and advection. Scalar properties vary at intratidal time scales as a result of tidal advection of scalar gradients and periodic mixing of the tide. Stratification may reflect intratidal variability due to the tide or as a result of surface heating and cooling cycles. Dissolved oxygen concentration varies on diurnal time scales due to the relative dominance of photosynthesis over respiration during the daytime and of respiration over photosynthesis during the nighttime.

Turbulent mixing acts to disperse constituents locally at intratidal time scales and contributes to horizontal dispersion through interaction with larger time scales via the process of shear dispersion [26]. Vertical turbulent mixing can be generated through shear layers formed between (1) water masses of differing density, (2) the horizontal tidal current and the bottom, or (3) the atmosphere and water surface when wind stress is applied.

Vertical turbulent mixing by the tide strongly depends on vertical density stratification, which can have a stabilizing or destabilizing effect [27]. The gradient Richardson number (Ri) is a scaling factor that indicates the relative importance of localized buoyancy and vertical shear [28] and can be written according to [24] as

$$Ri = \frac{N^2}{S^2} = \frac{-\frac{g}{\rho_0} \frac{\partial \rho}{\partial z}}{\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2} \simeq \frac{-\frac{g}{\rho_0} \frac{\Delta \rho}{\Delta z}}{\left(\frac{\Delta u}{\Delta z}\right)^2 + \left(\frac{\Delta v}{\Delta z}\right)^2}, \quad (2.1)$$

where N^2 is the Brunt-Väisälä frequency, S^2 is the shear-squared, g is the gravi-

tational constant, ρ_0 is a reference density for seawater, ρ is density, u and v are the horizontal components of velocity, and Δ indicates the surface to bottom difference between variables. Linear stability theory shows that instability occurs at $Ri \leq 0.25$ [28], but field studies have shown that the critical value of Ri may be higher [29]. Ri of 0.25 or less is the theoretical criteria for turbulence to intensify with time. Ri higher than 0.25 cause a suppression or decay of turbulence [30].

Tidal vertical mixing and water quality can depend on strain-induced periodic stratification, which is the occurrence of relatively high stratification during the ebb tide as compared to the flood tide [31, 32]. Flood currents are stronger near the surface than the bottom of an estuary. In a stratified, positive estuary this vertical shear introduces more oceanic water at the surface relative to the bottom, which causes isopycnals to become nearly vertical during the flood and reduce stratification [17]. Ebb currents are stronger near the surface than the bottom and this vertical shear introduces more riverine water at the surface relative to the bottom, which leads to an increase in vertical stratification [17]. The intratidal variability of stratification can have different influences on water quality depending on individual systems. For example, models have shown that tidally periodic stratification does not increase the likelihood of bloom formation, which is seen with persistent stratification [33]. However, the growth of phytoplankton may be affected when periodic stratification causes cell retention in light-limited bottom water and periodic mixing returns phytoplankton to the upper water column where cells can photosynthesize [34].

The discussion of the tidal current effect on mixing and water quality can be extended to a spring-neap cycle. Relatively weak neap tidal currents can cause low vertical mixing as a result of weaker bottom stress applied as compared with the spring tide. This can lead to higher stratification during the neap tide and

vertically mixed water columns during the spring tide [35]. Neap tide stratification has been associated with bottom water hypoxia in Narragansett Bay [16].

Dispersion of scalar constituents can be accomplished by horizontal turbulent mixing due to eddies of spatial scales comparable to the local depth [36]. This mixing it is likely caused by shoreline irregularities and secondary circulation, but it is not completely understood [36]. Mixing by turbulent eddies in the longitudinal direction (i.e. along the length of the estuary) is considered unimportant, since the effect of shear flow on horizontal dispersion is much larger [26]. Shear flow dispersion is discussed below.

Subtidal estuarine variability occurs at time scales exceeding a day. Subtidal flow is forced mainly by tidal fluctuations at the scale of the spring-neap tide and longer, wind that is associated with synoptic weather systems, rapid changes in freshwater runoff, and seasonal changes in river flow and surface processes. The wind exerts the strongest control on Narragansett Bay subtidal circulation [37]. Tidal residual flow is an intratidal and subtidal characteristic of nonlinear tidal processes. Tidal residuals can occur from the modification by Coriolis acceleration of the tidal current, the interaction of the tidal current with bathymetry, known as tidal rectification [38, 39, 26, 40, 41, 42, 43, 44, 45], and from the synoptic wind field [44].

The dispersion of constituents at subtidal time scales occurs via advection and shear flow dispersion. Shear flow dispersion is the spreading of a concentration gradient by velocity shear and turbulent mixing that acts perpendicular to the shear [46, 26]. Both the mean current shear and tidally-oscillatory shear can lead to dispersion [26]. It follows that vertical and lateral shear dispersion may result from tidal currents, including tidal residual currents, wind-driven flow, and gravitational flow. Shear dispersion by the tidal current shear leads to dispersion at subtidal

time scales [36]. Dispersion by shear currents is larger than that produced by small-scale turbulence [26]. Horizontal velocity gradients generally lead to much higher dispersion than vertical velocity gradients due to the large width to depth aspect ratio of estuaries [26]. Tidal pumping is the net advective flux of constituents that occur due to tidal residuals [26]. In estuaries where concentration gradients and tidal excursions are large the tidal pumping component of dispersion can dominate other mechanisms [47, 48, 49].

The theoretical description of wind-driven flow in estuaries is as follows. Wind stress acting along the relatively deep longitudinal axis of the estuary forces surface transport downwind and increases the sea surface gradient in the direction of the wind. The pressure gradient that results from the sea surface slope forces bottom flow in the opposite direction to the wind [50]. Relatively shallow areas are expected to flow in the direction of the wind throughout the water column, since the center of mass of the water is displaced to the deeper side of the basin [26]. Surveys of velocity in Narragansett Bay showed high correlation between the direction that the wind is blowing toward and the direction of the surface current [51, 52, 53, 19]. The surface current lags the wind force by approximately 2 hours in Narragansett Bay [19].

Physical controls on Narragansett Bay water quality is hypothesized to be an essential component of the water quality problem, but it is not well understood. The present study aims to describe the dominant physical processes in Greenwich Bay and the Warwick Neck channel using Eulerian observations of velocity, sea surface elevation, and tracers. The velocity observations provide unprecedented spatial resolution of flow in a subestuary. These observations allow for inferences to be made on the role of physical processes in water quality.

Greenwich Bay is a northwest-southeast aligned subestuary of the north-south

aligned Narragansett Bay (Figure 2.1). It has an average depth of 2.4 m at low mean water and a mean tidal range of 0.5 m [8]. The bathymetry of Greenwich Bay is characterized by a channel with maximum depth of 11 m entering at the mouth between shallow shoals to the north and south, and a gradual decrease in depth westward from the channel. The Greenwich Bay mouth can be defined as the line connecting Sandy Point and the flagpole on Warwick Neck. The length of the bay along the channel axis is 5 km, measured from the mouth to the west coast at the entrance of Apponaug Cove. A coastline feature in the south, named Sally Rock Point, causes a narrowing of the subestuary width, producing two distinct basins. The western and eastern basins are referred to as the inner basin and the outer basin, respectively. A number of small brooks enter at each of the Greenwich Bay coves. The largest of the tributaries are Hardig Brook discharging into Apponaug Cove, and the Maskerchugg discharging into Greenwich Cove [54].

The channel between Warwick Neck and Patience Island is here referred to as the Warwick Neck channel (Figure 2.1). In the vicinity of the study area the channel reaches a maximal depth of 18 m. The Warwick Neck channel and the Greenwich Bay channel form the bifurcation of a channel system to the south. The location where the Greenwich Bay and Warwick Neck channels connect is characterized by high bathymetry gradients. Depths change from 17 m in the Warwick Neck channel to 10 m in the Greenwich Bay channel within a distance of 350 m. Considering the difference in the depths of the two channels at their connection point, the Greenwich Bay channel may be viewed as a branch off the main channel that curve around Patience Island.

This paper has the following structure. Field observations and data analysis are discussed in Section 2.2. The results are subdivided into a number of different themes. Environment forcing conditions during the observational periods are

shown in Section 2.3.1. Tidal current characteristics and vertical turbulent mixing are discussed in Section 2.3.2. Findings regarding the subtidal variability is considered in Section 2.3.3 with emphasis on wind-driven characteristics. This is followed by a depiction of intratidal net flow, forced by the interaction between the diurnal wind and the semidiurnal tide, in Section 2.3.4. Results on the subtidal flow of the Warwick Neck channel in 2009 aid in interpretation of 2006 velocity data [19], which is discussed in Section 2.3.5.

2.2 Method

2.2.1 Observations

Profiles of velocity were obtained during the summer and early fall of 2009 using Teledyne RD Instruments bottom-mounted Acoustic Doppler Current Profilers (ADCP's). The work was funded by the NOAA Coastal Hypoxia Research Program grant NA05NOS4781201 and by Rhode Island Sea Grant numbers NA10OAR4170076 and NA10OAR4170076. Six-minute intervals were selected for ensemble averaging. Two ADCP's were located on either side of the Warwick Neck channel and one ADCP in the center of the Greenwich Bay channel (Figure 2.1). The depth resolution for both the western ADCP (WNW09) and the eastern ADCP (WNE09) was 1 m. The Greenwich Bay channel ADCP (GBC09) had a depth resolution of 0.5 m. Acoustic backscattering caused contamination of the top layers and these data were omitted; deepest layers are not sampled due to the physical configuration of the instrument. Details regarding the field deployments are provided in Table 2.1.

A total of fifteen SeaHorse Tilt Current Meters (TCM's) were deployed in Greenwich Bay to obtain high spatial resolution of the near-bottom flow (Figure 2.1). The work was funded by the NOAA Coastal Hypoxia Research Program grant NA05NOS4781201 and by Rhode Island Sea Grant numbers NA10OAR4170076

Table 2.1. ADCP deployment information during 2009.

Station	Period (GMT)	Location (°)	Depth (m)	Bins
WNE09	07 Jul – 30 Sep 2009	41.6640, -71.3706	11.95	12
WNW09	07 Jul – 30 Sep 2009	41.6656, -71.3731	14.63	12
GBC09	07 Jul – 30 Sep 2009	41.6699, -71.3929	8.49	8

and NA10OAR4170076. The instruments measured a weighted average of velocity over a depth interval of 10 cm to 1 m from the bottom. Sampling intervals were set at 5-minutes. The time period of reliable data ranged from 24 July 2010 to 23 August 2010. A TCM (GB15) was located in the Greenwich Bay channel approximately 530 m further landward from GBC09. Aspects of the flow at GB15 were compared with near-bottom flow at GBC09 for validation purposes. Each TCM consisted of a buoyant PVC pipe of 1 m length that was attached to a heavy base by a flexible tether [55]. A HOBO ®Pendant G Data Logger (UA-004-64) measured acceleration at the top of the PVC pipe. The measured data were processed by HOBOWare (Lite) software to obtain raw acceleration data. Raw acceleration was converted to tilt angles and finally to velocity components through calibration with acoustic current profilers [55]. The conversion is based on the principle that near-bottom velocity can be obtained from the known theoretical dependence of the inclination angle of a buoyant straight cylindrical pipe on the drag produced by flowing water [56, 57, 55]. The velocity resolution of each instrument is 0.95 cm/s, and accuracy can be estimated to be maximally 2.5 cm/s (V. Sheremet pers. comm.). Velocity data were averaged over 1-hour periods to reduce the standard error. The ratio of the standard error of the estimate to the mean absolute velocity could then be calculated for each period according to

$$\sqrt{\frac{SSE}{(N-2)}} / |\bar{V}| \quad (2.2)$$

where SSE is the sum of the squared errors, $N - 2$ is the number of degrees of

freedom, and $\overline{V}$ is the mean velocity [58]. The ensemble mean of equation 2.2 was less than 0.4 at each station. The standard error of the mean (SEM) was calculated to be 0.72 cm/s.

Temperature, salinity, and dissolved oxygen data were obtained from the Greenwich Bay Marina (GB) and Sally Rock (SR) Fixed-Site Monitoring Network buoys (Figure 2.1). The buoys are maintained through collaborations between the Rhode Island Department of Environmental Management Water Resources Division, the Narragansett Bay National Estuarine Research Reserve, the Narragansett Bay Commission, the University of Rhode Island Graduate School of Oceanography, Roger Williams University, and the Bay Window Program [16, 6, 59]. The temperature and salinity are measured by a surface and bottom YSI 6000-series sonde at fifteen-minute intervals. The surface and bottom sondes are situated within the upper one meter and the lower one meter of the water column, respectively. Snapshots of tracer profiles were obtained from an SBE-19 Conductivity-Temperature-Depth (CTD) instrument. *In situ* density was calculated using the Gibbs Seawater (GSW) Oceanographic Toolbox [60] that employs the international thermodynamic equation of seawater (TEOS-2010) [61]. Density stratification at the buoy sites was approximated by the surface to bottom density difference.

Tide and wind gauge data at Providence Harbor (P), Quonset Point (Q), and Newport (N) were acquired from the National Oceanic and Atmospheric Administration’s Physical Oceanographic Real Time System (NOAA PORTS) [62] (Figure 2.1). We use the convention of referring to the wind as the direction that it is blowing toward, unless otherwise stated. Observations of river transport were accessed from the national water information system of the U.S. Geological Survey (USGS) [63]. Rainfall data at T.F. Green Airport were retrieved from the Weather Underground [64].

Figure 2.1. Narragansett Bay bathymetry map showing important elements, including the locations of sea surface elevation and wind gauges (squares), and the Fixed-Site Monitoring Network buoys (diamonds). Tide and wind gauge stations are at Providence Harbor (P), Quonset Point (Q), and Newport (N). Buoy stations are at Greenwich Bay Marina (GB), Sally Rock (SR), and Mount View (MV). An enlarged map of Greenwich Bay (Box A) contains the locations of the ADCP's (stars) and the TCM instruments (circles). TCM stations are referred to as GB##, such that GB01 is location 1 on the map. (Bathymetry data from <http://www.geomapapp.org>)

2.2.2 Data analysis

Data analysis methods included filtering, harmonic and spectral analyses, and Empirical Orthogonal Function (EOF) analyses. Time series were filtered to compare the intratidal and subtidal properties. A 2.5–33 hour bandpass filter was applied to instantaneous velocity time series to assess the tidal (or intratidal) flow. In an attempt to isolate the response to diurnal wind the instantaneous velocity time series were processed by an 18–33 hour bandpass filter [65], as well as a 17.5–25.5 hour bandpass filter [66]. The diurnal wind response was not clearly represented by these bandpass-filtered time series. Dynamics related to diurnal wind variability and the subtidal flow were evaluated using lowpass-filtered time series with cutoff periods of 20 hours and 33 hours, respectively. The MATLAB [67, 68] functions `FILTFILT` and `BUTTER` were used to construct filters. Harmonic analyses of properties were conducted using the MATLAB toolbox `T-TIDE` [69]. Discrete energy spectra were calculated by implementing the method proposed by [70], using the MATLAB functions `DETREND` and `FFT` [67, 68].

MATLAB routines [71] were used to conduct EOF analyses on velocity data [58]. EOF analysis is the decomposition of a spatial time series, such as the vertically varying velocity data collected by an ADCP, into a linear combination of orthogonal spatial modes. It is a method for partitioning the variance of spatial time series. The EOFs are ordered by decreasing eigenvalue so that the first few modes with largest eigenvalues contain the majority of the variance of the data. By extracting the patterns of these functions the dynamical processes involved may become apparent. The orthogonal functions are defined by the covariance structure of the spatial time series [58].

For observations collected in time, $t = t_i$ ($1 \leq i \leq N$), and space, $\vec{x}_m$ ($1 \leq m \leq M$), the goal of the EOF procedure is to write a data series $\psi_m(t)$ at a location

$\vec{x}_m$ as the sum of M orthogonal spatial functions $\phi_i(\vec{x}_m) = \phi_{im}$ such that

$$\psi(\vec{x}_m, t) = \psi_m(t) = \sum_{i=1}^M [a_i(t)\phi_{im}], \quad (2.3)$$

where $a_i(t)$ is the amplitude of the i th orthogonal mode [58]. The time variation of a dependent scalar variable at each location is the linear combination of M spatial functions, ϕ_i , whose amplitudes are weighted by M time-dependent coefficients, $a_i(t)$, which characterizes the temporal variability of the modes [58]. The orthogonality requirement is

$$\sum_{m=1}^M [\phi_{im}\phi_{jm}] = \delta_{ij}, \quad (2.4)$$

where δ_{ij} is the Kronecker delta. Another constraint is that the time amplitudes $a_i(t)$ are uncorrelated over the sample data such that

$$\overline{a_i(t)a_j(t)} = \lambda_i\delta_{ij} = \overline{a_i(t)^2}\delta_{ij}, \quad (2.5)$$

where overbar denotes a time averaged value and $\lambda_i = \overline{a_i(t)^2}$ is the variance in each orthogonal mode [58]. Forming a covariance matrix with the time series as $\psi_m(t)\psi_k(t)$ the canonical form of the eigenvalue problem is derived for the i th mode at the m th location [58]:

$$\sum_{k=1}^M \overline{\psi_m(t)\psi_k(t)}\phi_{ik} = \lambda_i\phi_{im}. \quad (2.6)$$

In basic terms the procedure of solving for the EOFs of a single field is to (i) form a matrix of the observations and remove the temporal mean; (ii) construct the covariance matrix; (iii) find the eigenvalues and eigenvectors of the covariance matrix; (iv) find the highest eigenvalues and the corresponding eigenvectors (EOFs); (v) find the time-dependent amplitudes of each mode of the EOF [71, 58].

The method followed to calculate gradient Richardson number (equation 2.1) at the GBC09 ADCP station involved observations of surface and bottom current

velocity, temperature and salinity. The tidal current shear was estimated as the difference between 2.5–33 hour bandpass-filtered velocity time series measured in the surface and bottom ADCP sampling bins. The Brunt-Väisälä frequency was obtained from the Gibbs Seawater (GSW) Oceanographic Toolbox [60] by providing input of measured temperature and salinity. The only tracer measured as a continuous time series at GBC09 was bottom temperature. Surface temperature and salinity, as well as bottom salinity was obtained from the SR station inside Greenwich Bay. A correction was applied to the salinity data from SR based on differences observed between snapshot CTD profiles made during the ADCP sampling period in 2009, which showed salinity at GBC09 generally exceeded that of SR by approximately 1 psu. The CTD profiles at GBC09 and SR were made within 20 minutes of each other. Continuous surface and bottom temperature data were collected using a HOBO Pro v2 Data Logger (U22-001) at a location approximately one kilometer further landward up the channel from GBC09 between 13 July 2009 and 7 August 2009. Bottom temperature at this site agreed well with the ADCP bottom temperature record and the surface temperature at the same site agreed with surface temperature at SR.

The Ri was also approximated for the interior of Greenwich Bay. In this calculation the 2010 surface and bottom temperature and salinity time series from the SR buoy were used to calculate the Brunt-Väisälä frequency. The 2.5–33 hour bandpass-filtered bottom velocity observations from a 2010 TCM (GB08) were available for the calculation of shear, due to its proximity to the SR monitoring buoy. The only unknown in the calculation was surface tidal velocity, which was approximated by applying a correction factor to the bottom velocity record. The surface velocity was assumed to be 2.08 times stronger than the bottom velocity, which was observed to closely describe the relationship between surface and bottom

tidal currents at GBC09.

2.3 Results and discussion

2.3.1 Environmental forcing conditions

The dominant forcing mechanisms in the transport of water and its constituents include the astronomical tides and their overtides that are generated by nonlinear processes, the wind, and river transport. The characteristics of these forces during the field programs are considered in this section.

Sea surface elevation data for the duration of the sampling periods in 2009 and 2010 are decomposed into tidal constituents using the T-TIDE toolbox [69] and the seven largest constituents are discussed (Figure 2.2). The semidiurnal amplitudes are seen to dominate, particularly the M2 tidal constituent. The amplitude of the M4 overtide exceeds that of the diurnal O1 and K1 constituents. The M6 overtide exhibits a relatively low amplitude.

The wind field is characterized by a number of distinctive events during 2009 (Figure 2.3). Prior to day 240 such events include two southwestward wind pulses that are associated with relatively high precipitation and large freshwater runoff between day 201.6 and 205.9 (Figure 2.5). Another southwestward event is observed between day 224.0 and 226.0. Before day 240 the wind exhibits prolonged periods of diurnal wind variations. Diurnal wind variability is generally characterized by one of two patterns. This first pattern is sea breeze-land breeze alternations, and the second pattern is a fluctuation between sea breezes and wind relaxations. Diurnal wind variations relating to sea breezes occur on 24 days of the 51 days prior to day 240.

Sea breezes are identified based on the definition given in a study on the synoptic-scale controls of New England sea breezes [25]. The authors state that a sea breeze occurs when surface wind toward the northeast, the southeast, or

Figure 2.2. Instantaneous sea surface elevation characteristics at Quonset Point (Q) during the field programs in 2009 and 2010. (a) Amplitude and (b) phase for 2010 are shown, but it is not significantly different from the tidal harmonics calculated for 2009. (c) and (d) are sea surface elevation time series for 2009 and 2010, respectively.

the southwest shifts around midday to be toward the northwest as a result of land surface heating. As the land cools into the night this wind shifts back to its original direction. Alternatively, it shifts seaward and can be identified as a land breeze. The definition was somewhat modified for Narragansett Bay where the sea breeze wind caused by land heating is frequently toward the north, the north-northeast, or the north-northwest. Synoptic surface weather charts from the National Centers for Environmental Prediction [72] confirm these occurrences as sea breeze events when the local flow is not consistent with the geostrophic wind. The spectral energy density distribution of the north-south wind component indicates significant energy centered on the diurnal band (Figure 2.3). Diurnal wind variations are nonstationary and may be misrepresented by a distribution that assumes stationarity [73, 74]. Although some diurnal wind variations coincide with sudden shifts in the geostrophic wind field, the majority of the diurnal energy is attributed to the relatively abundant sea breeze days.

After day 240 the wind regime changes from being dominated by diurnal variability to longer time scale events. During this period the wind is less frequently directed toward the north as compared to the period prior to day 240. Only five sea breeze days are observed between day 240 and day 272. Sustained wind events are mostly toward the southwest, and a relatively high north-northwestward event occurs near the end of the time series. The change in wind regimes indicates a seasonal shift. A similar shift in wind regimes was observed during a Narragansett Bay circulation study in 2006 [19]. The implication of these regimes on water properties is discussed in Section 2.3.3.

Few sustained wind events are observed during the 2010 field study (Figure 2.4). Wind variability at time scales of a day or less is prevalent during the sampling period. Sixteen days of the 31-day field program are sea breeze days. The spectral

Figure 2.3. Time series of wind velocity components (starts in (a) and continues in (b)) and (c) the spectral energy density of the north-south wind component at Quonset Point (Q) in 2009. The north-south wind components are shown in black and the east-west components in grey. Sea breeze (SB) days are indicated in blue (north-south component) and red (east-west component). The dashed lines on the spectral energy density plot demarcate the 95% confidence interval.

energy density of the north-south component of the wind shows significant energy at the diurnal period.

Figure 2.4. (a) Wind velocity components and (b) the spectral energy density of the north-south wind component at Quonset Point (Q) in 2010. The north-south wind components are shown in black and the east-west components in grey. Sea breeze days are indicated in blue (north-south component) and red (east-west component). The dashed lines on the spectral energy density plot demarcate the 95% confidence interval.

The river transport during the summer of 2009 exhibited a few flow peaks following high rainfall events (Figure 2.5). The most prominent of these peaks is centered on day 205.7 when the Pawtuxet River transport reaches $58.05 \text{ m}^3/\text{s}$. This is notably higher than the mean annual flow rate of the Pawtuxet River of $9.87 \text{ m}^3/\text{s}$ [75].

Spectral energy density of the instantaneous current shows the relative importance of different forcing time scales on the response at ADCP and TCM stations. Spectra of the instantaneous near-surface currents exhibit large peaks at the semidiurnal period (Figure 2.6). The semidiurnal energy at GBC09 is low

Figure 2.5. (a) T.F. Green precipitation (black) and Pawtuxet River transport (grey) during field programs in 2009 and (b) T.F. Green precipitation (black) and Hunt River transport (grey) during field observations in 2010.

compared to the other stations. Significant peaks centered on 4.1 hours and 6.2 hours are observed in the near-surface currents at WNE09 and WNW09. These peaks correspond to the M6 and M4 overtide constituents. At GBC09 the peak at 6.2 hours is found to be significant. Relatively small peaks can be discerned at the diurnal period for near-surface currents at WNW09 and GBC09.

The diurnal spectral energy peaks are due to a combination of interacting processes. Diurnal variability includes the diurnal tidal current constituents and diurnal wind cycles. Diurnal tidal constituents can be approximated as stationary phenomena, but not the diurnal wind. The response produced by diurnal wind may be misrepresented by the spectral analysis that assumes stationarity [73, 74].

Spectral energy density of the instantaneous near-bottom currents at ADCP stations generally have similar characteristics to the near-surface currents. A difference is detected between the near-surface and near-bottom spectra of GBC09.

Figure 2.6. Spectral energy density of the ADCP near-surface currents along the principal axes at (a) GBC09, (b) WNE09, and (c) WNW09.

The near-bottom flow exhibits a significant peak at the M6 overtide period of 4.1 hours, but not the near-surface flow (Figure 2.7). The flow at GB15 has similar spectral energy density characteristics than the near-bottom current at GBC09 (Figure 2.7). There is good agreement between the magnitudes of energy peaks centered on the 4.1-hour, 6.2-hour, and 12.4-hour periods at GB15 and GBC09.

Figure 2.7. Spectral energy density of the near-bottom currents along principal axes at (a) GBC09 and (b) GB15 in the Greenwich Bay channel.

2.3.2 Intratidal variability and mixing

Tidal current characteristics

The tidal current ellipses of constituents at the ADCP and TCM stations reflect the spatial variability of the tidal current (Figure 2.8). Tidal constituents with the highest velocity at the ADCP stations include the N2, M2, S2, and M4 tides. All four near-surface tidal constituents are rectilinear and aligned with the channel axis at WNE09. Near-surface constituents at WNW09 are relatively rotary. N2, M2, and M4 have channel axis alignments, and S2 has an east-west alignment.

Near-bottom constituents at WNE09 are rotary, whereas those at WNW09 are rectilinear. The semimajor velocity of the M2 near-surface constituent is 0.35 m/s and 0.37 m/s at WNE09 and WNW09, respectively. The near-bottom velocity is 0.24 m/s and 0.27 m/s at WNE09 and WNW09, respectively. The alignment of near-surface tidal constituents at GBC09 are similar, with flood currents directed toward the north-northwest. The near-bottom tidal constituents are aligned such that flood currents are directed toward the northwest. In general the currents are rectilinear, but the M2 near-surface current is relatively rotary. Tidal constituent magnitudes are lower in the Greenwich Bay channel (GBC09) than in the Warwick Neck channel. The semimajor velocity of the M2 near-surface constituent is 0.17 m/s and the near-bottom velocity is 0.10 m/s.

Tidal ellipses at TCM locations show a gradual decrease in the tidal constituent magnitude for the lower water column from the mouth of Greenwich Bay to the inner basin (Figure 2.9). The tidal current is generally aligned with the coastline in both the inner and outer basin. The N2 and M2 constituents have rectilinear currents at the mouth, relatively rotary currents in the outer basin, and nearly rectilinear currents in the inner basin. The semimajor M2 velocity at GB15 is 0.10 m/s, which is the same as the M2 near-bottom velocity at GBC09.

The barotropic tidal (2.5 – 33 hour bandpass-filtered) currents along principal axes at the ADCP stations provide insight to the general tidal current characteristics over a spring-neap tidal cycle (Figure 2.10). The period shown corresponds with conditions of low runoff and wind that generally vary at time scales longer than a day. A diurnal tidal inequality leads to slight differences in the speed of successive flood currents at all the ADCP stations. The diurnal inequality of the tidal current can cause the net advection of water constituents like bacteria and dissolved oxygen, which is explained in Section 2.3.4. The maximum barotropic

Figure 2.8. Near-surface (black) and near-bottom (blue) current ellipses of the (a) N2, (b) M2, (c) S2, and (d) M4 tidal constituents at ADCP stations.

Figure 2.9. Current ellipses of the (a) N2, (b) M2, (c) S2, and (d) M4 tidal constituents at the TCM's.

flood and ebb current speeds for the sampling duration are highest on the western side of the Warwick Neck channel and lowest in the Greenwich Bay channel. Maximum barotropic flood magnitudes are 0.58 m/s, 0.71 m/s, and 0.30 m/s at the WNE09, WNW09, and GBC09 stations, respectively. Maximum barotropic ebb magnitudes are 0.55 m/s, 0.64 m/s, and 0.32 m/s at the WNE09, WNW09, and GBC09 stations, respectively. The tidal current speed is thus higher during the flood tide than the ebb tide at WNW09. Maximum flood and ebb speeds are of similar magnitudes at each of the other two stations. Maximum flood and maximum ebb occur approximately two hours before high and low tide, respectively. The mean duration of the flood is higher than the ebb at all the ADCP stations. The flood duration is 7.1 hours at WNE09, and 7.2 hours at WNW09 and GBC09.

A double-peaked flood current and single-peaked ebb current are evident during each tidal cycle. This asymmetry is characteristic of Narragansett Bay tidal currents [37, 76, 77, 78, 19] and is due to the large amplitude and phase changes of the M4 and M6 overtides relative to M2 with distance up the Bay [79]. Double-peaked floods and single-peaked ebbs are not expected to affect the transport of water constituents like bacteria and dissolved oxygen [75].

The tidal (2.5 – 33 hour bandpass-filtered) current measured by Greenwich Bay TCM's reveal similar characteristics to the ADCP data (Figure 2.11). TCM data provide a significant improvement on spatial coverage. Note that the periods shown correspond to times of low runoff and wind that varies at time scales longer than a day. Station GB15 near the mouth exhibits double-peaked flood and single-peaked ebb characteristics during both the spring and neap tides. At GB15 the maximum near-bottom tidal current speed along the principal axis of the channel is 0.30 m/s for the duration of the 2010 sampling period. For comparison, the maximum near-bottom current speed measured at GBC09 is 0.34 m/s during the

Figure 2.10. Tidal barotropic ADCP velocity along the principal channel axes at (a) WNE09, (b) WNW09, and (c) GBC09 over a spring-neap tidal cycle (black). Positive velocity is landward and negative velocity is seaward. The sea surface elevation at Quonset Point (Q) is shown in grey.

sampling period in 2009. The near-bottom tidal current is relatively weak at GB01 near the head of Greenwich Bay where a maximum tidal speed of 0.08 m/s is measured.

Figure 2.11. Tidal near-bottom north-south (solid black line) and east-west (dashed black line) velocity components at Greenwich Bay TCM stations, GB01 and GB15, during a spring tide (a – b) and neap tide (c – d). Positive velocity is northward and eastward. The sea surface elevation at Quonset Point (Q) is shown in grey.

Hydrography at intratidal time scales and mixing

Hydrography determines water column stratification, which plays a role in vertical turbulent mixing characteristics and water quality. Vertical turbulent mixing is an important physical process that contributes to the dispersion of water and its constituents in estuaries, and is a function of the vertical stratification, tidal current shear, and wind conditions. Low vertical mixing during relatively high vertical stratification and low current shear may cause poor water quality by reducing atmospheric replenishment of oxygen in the bottom water [14, 15, 16, 17, 18].

Periodic stratification and mixing resulting from the tidal flow can influence phytoplankton growth [34].

A comparison of the tidal current vertical shear at the Greenwich Bay channel ADCP during two periods, with similar tidal ranges of approximately average magnitudes, shows differences relating to wind forcing (Figure 2.12 (a) – (d)). A higher temporal averaged vertical shear occurs during a time when continuous cycles of diurnal wind prevails compared to periods when sustained wind dominates. The increased average vertical shear during diurnal wind is due to relatively high peaks that vary at time scale of about three to seven hours. Some of these peaks exceed 0.002 1/s^2 in magnitude. In contrast, the vertical shear during the period of sustained wind varies with lower magnitudes and higher frequency. The majority of peaks have magnitudes of less than 0.001 1/s^2 . Similar increases in shear at tidal time scales associated with diurnal wind are observed at the Warwick Neck ADCP's. Diurnal wind variability is nonstationary and exhibits changes at similar time scales than the tidal variability. The increased tidal current shear during diurnal wind conditions suggests an interaction between the diurnal wind and semidiurnal tide.

Results suggest that vertical tidal shear during periods of average tidal ranges, average sustained wind, and moderate stratification does not lead to vertical turbulent mixing in the channel of Greenwich Bay, but when these conditions are coincident with diurnal wind cycles the tendency for vertical turbulent mixing at intratidal time scales can be increased. Time series of Ri show an increased tendency for vertical turbulent mixing, resulting from increased vertical shear, by the diurnal wind compared to mixing tendency produced by the tidal shear alone. The shear peaks observed under moderately stratified conditions during cycles of diurnal wind correspond with decreases in Ri (Figure 2.12 (e) and (g)) that are lower

in magnitude and longer in duration than those produced under moderately stratified conditions during sustained wind between days 256 and 257.5 (Figure 2.12 (f) and (h)). Note that high sustained wind leads to an unstable water column between days 255 and 256, and slight increases in shear after day 257.5 could be related to wind variability at intratidal time scales.

Figure 2.12. Contrasting vertical turbulent mixing characteristics at GBC09 during a period of continuous diurnal wind cycles (left panels) and a period when wind varies at time scales longer than a day (right panels). The contrasted time series are taken at similar stages of the spring-neap tidal cycle. (a) – (b) North-south (solid grey) and east-west (dashed grey) wind components and normalized sea surface elevation (black) at Quonset Point (Q); (c) – (d) the vertical shear squared of the tidal current; (e) – (f) the gradient Richardson number (only values between zero and seven are shown); and (g) – (h) the Brunt-Väisälä frequency.

Salinity, temperature, and current meter data from 2010 are assessed to reveal hydrography and vertical tidal mixing characteristics in the interior of Greenwich Bay. An approximate Ri is calculated via the method explained in Section 2.2. The approximation assumes that the unknown surface tidal current is a constant factor of the bottom current. This Ri is not useful for inferring about vertical

mixing effects by diurnal wind, since the influence of diurnally varying wind on the surface current is not known at the TCM locations. Despite the unknowns the results of Ri provide a rough estimate of the tendency for vertical turbulent mixing in Greenwich Bay.

Intratidal characteristics of temperature, salinity, and stratification are determined from data at the Sally Rock (SR) buoy in Greenwich Bay. Surface salinity and vertical density stratification exhibit weak semidiurnal variation (Figure 2.13 (a) and (b)). The surface salinity decreases from slack before maximum ebb when it reaches its lowest value. Relatively fresh surface water is advected from the head of Greenwich Bay during the ebb current. Surface salinity increases after maximum ebb to its highest value at the maximum flood current when more saline surface water is advected into Greenwich Bay from the mouth. Generally these tidal salinity anomalies are less than 1 psu. The tidal signal in bottom salinity is smaller than at the surface. Vertical temporal variability of density suggests strain-induced periodic stratification at the buoy location (Figure 2.13 (b)). Vertical density stratification is highest during the ebb and lowest during the flood. The tidal straining influence on stratification is superimposed on the longer term stratification characteristics, which will be discussed in Section 2.3.3. Temperature at the surface and bottom varies diurnally by less than one degree Celsius as a result of surface heating and cooling (Figure 2.13 (a)).

The Ri estimated for the interior of Greenwich Bay indicates that the water column has a low tendency to mix during periods when weak stratification is observed. Highest values of Ri occur during the ebb current and lowest values during the flood current, consistent with tidal variations in stratification. The spring-neap tidal period between day 205 and 218 exhibits a minimum Ri of 0.42 and a temporal mean of 150 (Figure 2.13 (c)) with vertical density stratification

varying between 0.1 kg/m^3 and 1.5 kg/m^3 . The average value is high compared to that of a number of estuaries and sounds where Ri ranges from 0.8 to 19.9, but within the cited range compared to surface water in Long Island Sound that has an average Ri of 447.7 [80]. The stability of the water column is attributed to the low vertical shear of weak tidal currents in Greenwich Bay.

Figure 2.13. Hydrography and vertical turbulent mixing at Sally Rock (SR) for the duration of the measured time series shown together with the east-west component of velocity from GB07 (grey). Positive velocity is toward the east and corresponds to the ebb current. (a) The surface (red) and bottom (blue) instantaneous salinity time series, and the surface (magenta) and bottom (cyan) instantaneous temperature time series. (b) The vertical density gradient. (c) The gradient Richardson number – horizontal dashed line indicates $Ri = 0.25$, which represents the transition between stability and instability.

The weakly stratified water columns in the channel and interior of Greenwich Bay seem to be prone to prolonged periods of stability during low to moderate tidal ranges due to low tidal energy available for vertical mixing. This would contribute to hypoxia by isolation of the bottom water from the atmosphere. Diurnal wind can improve vertical turbulent mixing by increasing shear at intratidal time scales.

2.3.3 Subtidal variability and advection

Subtidal characteristics indicate the potential for dispersion by advection and is therefore important for water quality considerations. Subtidal flow is defined as the response of water to forcing that varies at time scales exceeding a day. Subtidal flow, sea surface elevation, and hydrography are represented by 33-hour lowpass-filtered time series. The wind forcing discussion in this section focuses on the subtidal response to sustained wind events, which vary at time scales longer than a day. The response to wind variability at time scales of a day or less is discussed in Section 2.3.4.

The present study uses ADCP observations on both sides of the narrow Warwick Neck channel to describe the lateral structure of the subtidal axial flow in 2009. Two ADCP instruments are critical for understanding the complicated dynamics through the channel. Previous ADCP data from the center of the Warwick Neck channel showed a dramatic difference in the subtidal flow between summer and autumn conditions during 2006 [19]. The autumn flow regime exhibited low current variance compared to the summer regime. It was hypothesized that a lateral change in the axial flow of the channel may have caused the observed flow transition between the summer and autumn. A new interpretation for the discrepancy in the summer and autumn flow in 2006 relating to diurnal wind variations is derived from the 2009 observations and discussed in Section 2.3.5.

Sea surface elevation and wind

Wind that varies at time scales exceeding a day is known to dominate the subtidal current variability in Narragansett Bay and it determines the subtidal sea surface gradient [37]. The longitudinal (axial) sea surface gradient is obtained by subtracting sea surface elevation time series, given relative to the NAVD88 datum [62], at Newport (N) from those at Providence Harbor (P). The NAVD88

datum is a fixed reference for elevations determined by geodetic leveling [62]. Positive sea surface gradient is defined to point in the direction of increasing height. The subtidal longitudinal surface gradient corresponds closely to the north-south wind component during the 2009 sampling period (Figure 2.14 (a)). A correlation coefficient of 0.91 is calculated when the 33-hour lowpass-filtered time series of north-south wind (direction toward) and surface gradient are compared.

Wind variability that occurs at time scales of a day or less contributes to variations in the longitudinal sea surface gradient, suggesting an influence of the wind on the flow at these time scales. This influence is discussed in more detail in Section 2.3.4. Wind variability at these scales can be identified in 20-hour lowpass-filtered time series (Figure 2.14 (b)), which are considered here to show that diurnal wind cycles and their effect on the sea surface gradient are not resolved in the 33-hour lowpass filter representations. A comparison of the 20-hour lowpass-filtered time series of wind and sea surface gradient produces a correlation coefficient of 0.88. The time series of the surface gradient contains a small amount of energy from the semidiurnal band as a result of filter edge effects. This semidiurnal energy can be identified as amplitudes that vary by less than 0.01 m about the mean signal. Variability in the surface gradient that correlates with diurnal wind variability is characterized by larger diurnal fluctuations. The correlation is most clearly evident for four successive events centered on day 210.0, four successive events between day 226.0 and 230, and four successive events centered on day 265.0. The 20-hour lowpass-filtered representation of wind shows the abundance of diurnal wind variability.

The high correlation coefficients suggest that the north-south wind is dominant in determining the longitudinal subtidal sea surface elevation gradient. Wind blowing toward the north increases the up-estuary gradient and wind blowing to-

ward the south causes a reduction or reversal of the up-estuary gradient. These findings agree with previous observations in the West Passage of Narragansett Bay [37]. This is also consistent with other findings regarding local wind effects on the surface of an estuary [81].

The subtidal sea surface gradient may additionally be forced remotely through shelf Ekman dynamics. The northward wind often occurs together with an eastward wind component. This northeastward and north-northeastward wind may lead to upwelling on the shelf outside the estuary mouth, which would promote an increased up-estuary subtidal sea surface gradient by decreasing the water level at the mouth [82, 83, 84, 85]. In contrast, a southwestward and south-southwestward wind event may cause downwelling on the shelf, an associated increase in the water level at the mouth, and a reduced or reversed up-estuary gradient.

Figure 2.14. (a) The 33-hour lowpass-filtered time series of longitudinal sea surface elevation gradient (grey) between Providence Harbor (P) and Newport (N) and the north-south wind component at Quonset Point (Q) (black). (b) The 20-hour lowpass-filtered time series of longitudinal sea surface elevation gradient (grey) between Providence Harbor and Newport and the north-south wind component at Quonset Point (black). Positive wind values indicate northward-blowing wind.

Two distinct forcing regimes can be identified in time series of sea surface elevation gradient and wind during the ADCP sampling period (Figure 2.14), similar to those observed during the Narragansett Bay observational study in 2006 [19].

The transition between the two regimes in 2009 can be approximately defined, such that the first regime extends from the start of the time series to day 240, and the second regime extends from day 240 to the end of the sampling period. The temporal mean of the up-estuary subtidal sea surface elevation gradient is higher during the first period than during the second period. The difference in mean surface gradient may be due to dissimilarity in the wind characteristics between the two periods, and/or differences in observed density distributions. The first period is generally characterized by wind with a dominant northward component. A large portion of the observed northeastward and north-northeastward wind are attributed to diurnal summer sea breezes, but in some instances these conditions are associated with the geostrophic surface wind field (compare Figures 2.3 and 2.14). A few intermittent wind events with southward components are seen to temporarily reduce the up-estuary sea surface elevation gradient during the first period. The second period is characterized by wind with a dominant southward component. Wind events are generally of longer duration during the second period than during the first period.

Hydrography

The differences in sea surface gradient and wind between the first time period (days 188 – 240) and the second time period (days 240 – 271) are concurrent with differences in hydrography (Figure 2.15). Two separate hydrographic regimes were similarly observed in Narragansett Bay in 2006 [19]. The first time period of the 2009 study corresponds to relatively high vertical density gradients at Mount View (MV) in the West Passage and at Sally Rock (SR) in Greenwich Bay (Figure 2.1). Stratification intensifies following the freshwater runoff event that reaches a peak on day 205.7. During the second period in 2009 the water column generally resembles mixed or weakly stratified conditions. A gradual decrease in surface and

bottom water temperature and a gradual increase in surface and bottom salinity occur during this time. Short restratification events correspond with relatively small peaks in freshwater runoff, whereas mixing events coincide with sustained wind. The changes in wind and hydrography between the first and second part of the sampling period reflect seasonal characteristics. The first period corresponds to summer forcing conditions, and the second period marks a transition to autumn forcing conditions.

Figure 2.15. The 33-hour lowpass-filtered surface (red) and bottom (blue) salinity, and surface (magenta) and bottom (cyan) temperature at (a) Mount View (MV) and (c) Sally Rock (SR); 33-hour lowpass-filtered stratification at (b) Mount View and (d) Sally Rock is also shown.

Long term flow structure

The long term axial circulation in the Warwick Neck channel is laterally sheared as a result of the gravitational flow that is modified by Coriolis acceleration and the frictional influence of the channel as noted in other channel systems [86, 87, 88]. The vertical structure of the current obtained by averaging over the entire sampling period exhibits two-layered flow on the eastern side of the channel (WNE09). A surface flow of 1.5 m thickness is directed toward the estuary mouth and a flow below it of 5.5 m thickness is directed toward the estuary head (Figure 2.16 (a)). The maximum mean flow speed in the upper and lower layers are 0.015 m/s and 0.014 m/s, respectively. On the western side of the channel (WNW09) the whole water column flows in the direction of the estuary mouth (Figure 2.16 (b)). The maximum mean flow speed at this location is 0.035 m/s in the lowest layer. The long term flow structure at these two stations is consistent with net exchange observed in channels of other estuaries [88].

Long term channel exchange at the mouth of Greenwich (GBC09) is consistent with two-layered flow forced by gravitational circulation. The vertical structure of the current obtained by averaging over the entire sampling period is characterized by a down-estuary (outflowing) surface current of 0.5 m thickness and an up-estuary (inflowing) current below the surface layer of 5 m thickness. The maximum mean flow speed in the upper and lower layers are 0.004 m/s and 0.032 m/s, respectively.

Wind-driven circulation from ADCP's (2009)

The response of the Warwick Neck and Greenwich Bay channels to sustained wind events exhibits lateral and vertical variability that is characterized by the subtidal (33-hour lowpass-filtered) flow at WNE09, WNW09, and GBC09. Subtidal axial flow in the Warwick Neck channel responds to wind stress that can be

Figure 2.16. Temporal mean vertical velocity along the principal axes at (a) WNE09, (b) WNW09, and (c) GBC09. The average is for the entire sampling period from 7 July – 30 September. Positive velocity indicates flow in the direction of the estuary/subestuary head.

transferred through the depth of the water column and dominates over the gravitational flow. Subtidal flow in the Greenwich Bay channel determines the efficiency of exchange between the subestuary and Narragansett Bay and is strongly dependent on the wind. The wind response at GBC09 is consistent with the theory of wind-driven flow in estuaries.

A conceptualization of the flow in the Warwick Neck channel summarizes its response to southwestward, northward, and north-northwestward wind events in 2009 relative to the long term mean flow structure (Figure 2.17 (a)). Typical southwestward wind events lead to an increase in the down-estuary surface flow magnitude and bottom flow that resembles the long term mean with down-estuary flow in the west and up-estuary flow in the east (Figure 2.17 (b)). Southwestward wind that persists for an extended period of time causes a further increase in down-estuary surface flow and a reversal of the bottom flow in the east toward the estuary mouth, such that the entire channel flow is in a down-estuary direction (Figure 2.17 (c)). During a northward wind event the surface flow in the east is reversed relative to the long term mean, such that it is directed toward the estuary head (Figure 2.17 (d)). Northward wind causes a reduction in the magnitude of the surface down-estuary flow in the west. The bottom flow resembles the long term mean during the northward wind event. A north-northwestward wind event

forces the entire channel flow in a direction toward the estuary head (Figure 2.17 (e)). It reverses the surface flow and the western channel bottom flow relative to the long term mean.

Figure 2.17. Conceptualization of wind response in the Warwick Neck channel from the perspective of looking along the channel axis toward the estuary head. Surface and bottom currents are indicated by circles with dots for down-estuary flow and pluses for up-estuary flow. The sizes of the circles represent relative magnitudes not to scale. The response to generalized sustained wind events are shown relative to (a) the long term temporal mean, including (b) typical southwestward wind, (c) persisting southwestward wind, (d) typical northward wind, and (e) north-northwestward wind.

The effect of wind on the subtidal flow is now discussed in more detail with reference to the sustained events observed in 2009. A statistical treatment of the sustained wind events is not feasible due to the low occurrence of similar events. Events include: a predominantly northward wind event centered on day 207.5 (N1); a north-northwestward wind event centered on day 270.5 (NNW1);

an east-southeastward wind event centered on day 196.0 (ESE1); a southeastward event centered on day 220.0 (SE1); a predominantly eastward event centered on day 258.0 (E1); a predominantly southward event centered on day 244.0 (S1); south-southwestward events centered on day 190.2 (SSW1) and 202.8 (SSW2); a southwestward and south-southwestward event centered on day 225.0 (SW1); a southwestward, west-southwestward, and south-southwestward event centered on day 254 (SW2); and a southwestward event centered on day 260 (SW3). Note that the relatively high wind event that is centered on day 205.4 shifts from south-southeastward to south-southwestward and back to south-southeastward at 10 – 12 hour periods. This wind event and all others not listed above as sustained events vary at time scales of a day or less and contribute to intratidal residual flow, which is discussed in Section 2.3.4

Subtidal exchange on both sides of the Warwick Neck channel varies strongly with depth as a function of southward and southwestward wind. The effect of wind blowing toward the south, the south-southwest and the southwest on the flow on the eastern side of the channel (WNE09) is fairly consistent (Figure 2.18 (a) and (d)). During the events labeled SSW1, SSW2, SW1, S1, and SW3 the wind causes an increase in the magnitude of the down-estuary surface flow relative to the mean flow, particularly on the western side of the channel where a maximum increase of a factor of 6.8 occurs. During the extensive SW2 event the wind not only forces an increase of the surface current speed by about an order of magnitude, but also causes a reversal of the deep layers. The wind overcomes the gravitational force observed in the mean flow on the eastern side of the channel, such that the subsurface water column flows toward the estuary mouth. The SW2 event dramatically increases the mean whole-water column down-estuary flow and the vertical shear on the western side of the channel (WNW09) with maximum current

speed occurring in the upper water column (Figure 2.18 (b)). At both stations the flow structures suggest that momentum from applied wind stress during SW2 is translated from the water surface to the deeper layers. The event labeled SW3 has a similar effect than SW2 on the flow at WNW09. The SSW1, SSW2, SW1, and S1 events cause an increase in the mean surface down-estuary flow on the western side of the channel.

Wind events that exhibit eastward wind components seem to have a relatively low influence on the Warwick Neck axial subtidal flow. A southeastward wind labeled SE1 produces flow structures on the eastern side of the Warwick Neck channel that are similar to those during southwestward events. The western side of the channel exhibits a weak reversal of the surface flow during SE1. During the east-southeastward and eastward events (ESE1 and E1) the flow structures at WNE09 and WNW09 resemble those of the temporal mean flow.

The two-layered mean flow in the Greenwich Bay channel is intensified by wind of dominant eastward and southeastward velocity, which is observed during the ESE1, SE1, and E1 events (Figure 2.18 (c) and (d)). These events cause dramatic increases in the maximum upper layer outflow relative to the mean flow by factors of 5.5, 13.8, and 9, respectively. The maximum lower layer flow is increased by factors of 3, 4, and 2.6, respectively. Results suggest that flow intensification is due to the fact that the eastward wind stress acts in the same direction as the surface layer gravitational flow. The associated increase in bottom flow suggests that the wind produces a surface slope and consequent pressure gradient that forces the lower water column in the same direction as the bottom layer gravitational flow.

An intensification of two-layered flow in Greenwich Bay is observed toward the end of the southwestward event, SW1, but not during the SW2 or SW3 events. The intensification during SW1 is attributed to a sudden shift from southwestward

Figure 2.18. The 33-hour lowpass-filtered subtidal velocity along principal axes at the (a) WNE09, (b) WNW09, and (c) GBC09 ADCP stations. Each line of the velocity plots represents a depth bin; red lines indicate upper water column layers and blue lines indicate lower water column layers. The thick red and blue lines show the top and bottom layers, respectively. Positive velocity indicates up-estuary flow (e.g into Greenwich Bay for GBC09). (d) Instantaneous wind velocity with color lines showing sea breeze (SB) days (positive indicates wind blowing toward the north and the east). Vertical grey bars delineate wind events named for the direction that the wind is blowing toward.

to south-southwestward wind that interacts with the ebbing current (the wind-tide interaction mechanism is discussed in Section 2.3.4). The SW3 event causes relatively weak and depth-uniform inflow at GBC09. During SW2 the whole water column flows into the subestuary and is vertically sheared with increased speed near the bottom, except when a shift in the wind to a more westward direction occurs at the scale of the ebbing current. A sheared, inflowing water column with increased bottom velocity is also observed during SSW2. The observations show that south-southwestward and southwestward wind force the surface current in an opposing direction to the gravitational force. The increased bottom velocity observed during SW2 and SSW2 is ascribed to wind shifts that occur at the scale of coinciding flooding currents, rather than to wind stress translated to the bottom layers and acting together with the gravitational force.

Sustained wind with dominant northward components can lead to Warwick Neck channel flow that opposes the gravitational circulation. During event N1 a reversal of the surface flow occurs on the eastern side of the Warwick Neck channel at WNE09 (Figure 2.18). The whole water column moves nearly uniformly in the direction of the estuary head. On the western side of the channel, at WNW09, the water column flows weakly toward the estuary mouth. The NNW1 event forces a relatively strong, depth-uniform up-estuary flow at WNE09 with speed of 0.08 m/s and an up-estuary flow at WNW09 with surface speed of 0.07 m/s. Flow structures suggest that the effect of NNW1 wind stress is to oppose the surface mean flow at WNE09 and WNW09. At the WNW09 station it is evident that momentum from the wind is translated throughout the depth, causing flow reversals also near the bottom.

A disparity exists in the response to predominantly northward wind events in the Greenwich Bay channel. The N1 event causes a weakening of the two-layered

flow at GBC09 and the NNW1 event leads to a depth-uniform outflowing water column with maximum speed of 0.06 m/s. The response to NNW1 at GBC09 can be explained if it is assumed that a very thin, unmeasured surface layer exists in the channel, and across the shallows on either side of the channel, flow in the direction of the wind. The wind sets up a surface slope and an associated pressure gradient, which forces the majority of the channel flow toward the estuary mouth, in opposition to the gravitational flow.

Water quality inferred from 2009 circulation

We test the hypothesis that circulation plays a major role in the water quality of Greenwich Bay and since subtidal circulation is dominated by the wind, it is expected that increased or decreased wind-driven advective exchange would strongly influence water quality. Water quality problems in Greenwich Bay have been attributed to high nutrient loads and consequent eutrophication, which leads to recurrent hypoxia. The Greenwich Bay Special Area Management report states that Greenwich Bay is affected both by impaired water from subsystems to the north in Narragansett Bay, and by internal nutrient pressures introduced mainly by wastewater input into Greenwich Cove [89]. The water quality of the Seekonk and Providence River subestuaries in the north are often degraded in the summer [9, 10]. Two important questions in the context of Greenwich Bay water quality is 1) what is the likelihood that water from the north contaminates this subsystem, and 2) how does water internal to the system become retained?

Greenwich Bay resupply varies significantly with environmental forcing (Figure 2.19). The water quality of Greenwich Bay may be compromised when flow in the lower water column on the western side of the Warwick Neck channel tends westward, and the bottom water of the Greenwich Bay channel flows into the subestuary. This flow configuration is a favorable state for inflow of high nutrient

or hypoxic water from the north into Greenwich Bay. The flow regime is clearly observed when the wind is directed toward the south and the southwest. When the wind is eastward and southeastward the inflow of bottom water is relatively strong in the Greenwich Bay channel, but the bottom water along the west of the Warwick Neck channel does not have a high westward component. This water could impact Greenwich Bay through the channel connections further south, but it is not clear if inflowing bottom water would originate from the West Passage in the south, the Warwick Neck channel in the north, or both. Water from the north is least likely to impact Greenwich Bay when the wind has a northward component. During predominantly northward wind the bottom flow along the west of Warwick Neck channel is weak and tends in an up-estuary direction.

Figure 2.19. Spatial structure of the surface (red), the mid-water column (magenta), and the bottom (blue) subtidal current at the three ADCP locations. Flow vectors represent the maximum response to the sustained wind events named (a) NNW1, (b) N1, (c) E1, (d) SW3, (e) S1, and (f) SE1.

Flow along the Greenwich Bay channel as a function of wind indicates the

conditions for retention or flushing of its interior (Figure 2.19). Conditions that are conducive to efficient exchange include east-southeastward, southeastward, and eastward wind events. During these events the vertical two-layered structure of surface outflow and bottom inflow is strong. The channel volume flux can be estimated by multiplying the depth-averaged velocity of a portion of the water column with the cross-sectional area of the channel for that portion. The channel was assumed to be triangular in the lower 2.5 meters with a width of 380 m and rectangular in the upper 6 meters with a width of 380 m. Estimated channel volume flux during the eastward wind event (E1) is $140 \text{ m}^3/\text{s}$ into the subestuary and $40 \text{ m}^3/\text{s}$ out of the subestuary. During the prominent north-northwestward wind event (NNW1) the majority of the water column flows strongly out of the estuary. This scenario may be favorable to exchange if it is assumed that a thin, unmeasured surface layer flows into the estuary in the channel and along the shallows on either side of the channel. An estimate of volume flux during this event is $210 \text{ m}^3/\text{s}$ out of the subestuary. This volume flux is consistent with the outflux of the entire volume of Greenwich Bay over the 22-hour duration of the event. Northward wind causes relatively weak flow, which is not directed along the Greenwich Bay channel axis. This condition reduces the potential for exchange between Greenwich Bay and Narragansett Bay. Estimated channel volume flux during the northward wind event (N1) is $50 \text{ m}^3/\text{s}$ into the subestuary and less than $10 \text{ m}^3/\text{s}$ out of the subestuary.

The prevalence of wind-driven influx from northern sources into Greenwich Bay and retention of water inside Greenwich Bay must be determined to understand the impact of these occurrences on water quality. The direction of the summer wind generally varies between 330° and 0° (direction toward), and between 0° and 70° on a compass rose, i.e. the northward wind component dominates (Fig-

ure 2.20). This suggests that conditions limiting exchange in the Greenwich Bay channel may occur frequently. However, a large portion of the northward, north-northeastward, and northeastward summer wind falls in the category of diurnal wind variability as opposed to sustained wind. The water response to diurnal wind is discussed in Section 2.3.4. A low occurrence of southwestward wind events is observed during the summer, suggesting that contamination of Greenwich Bay water from northern sources is rare.

Figure 2.20. Relative distribution of wind direction between 1 June and 31 July of different years. The direction that the wind is blowing toward is plotted. Northward (0°) points to the top of the page and eastward (90°) points to the right of the page.

Wind-driven circulation from TCM's (2010)

The effect of wind on the subtidal circulation throughout Greenwich Bay is determined from a spatial distribution of TCM data. A strong relationship is observed between the wind and the subtidal current in Greenwich Bay in 2010. The wind forces two-layered flow, which is modified by topography and Coriolis acceleration. An EOF analysis of the TCM velocity components demonstrates the

wind-driven subtidal variability (Figure 2.21). The first EOF mode of the east-west velocity describes 62% of the variance. The principal component of mode 1 produces a correlation coefficient of -0.75 with the east-west wind when considering the direction that the wind is blowing toward (Figure 2.21 (a)). The first empirical orthogonal function (EOF1) of the subtidal east-west velocity exhibits strong magnitudes near the mouth of Greenwich Bay and along the longitudinal (northwest-southeast aligned) axis in the outer basin (Figure 2.21 (c)). The northwest-southeast axis is a continuation of the channel axis. Speeds are of intermediate value in the northwestern corner of the bay along the longitudinal axis. Relatively weak flow occur at TCM's along the coast. The flow is in the opposite direction from the other stations in the southwest corner of the inner basin, along the southern coastline immediately east of Sally Rock, and on the shallow shoal south of the channel (GB03, GB07, and GB13 locations shown in Figure 2.1). From the EOF analysis it can be deduced that the east-west component of Greenwich Bay bottom water generally flows in the opposite direction to the east-west wind component. This is consistent with the theory of wind-driven flow in estuaries. Theory predicts that flow over shallows is in the direction of the wind. The EOF analysis suggests that this applies to the above-mentioned stations.

Flow patterns show lateral variability suggesting that Coriolis acceleration is non-negligible when considering the wind-induced flow in Greenwich Bay. This is supported by an estimate of the Rossby number, which is calculated from a representative subtidal velocity $U = 0.02$ m/s, a length scale $L = 2000$ m, and the Coriolis parameter $f = 0.967 * 10^{-4}$ 1/s to be $R_o = U/Lf = 0.1$. The effect of the Coriolis term is to deflect the surface current to the right of the direction of the wind and the bottom current to the left of the wind [90, 91].

The north-south subtidal current illustrates the effect of the wind and Coriolis

Figure 2.21. EOF analysis of Greenwich Bay subtidal TCM velocity with comparison to the wind (grey lines). First principal components (PC1) are indicated by black lines. (a) PC1 of the subtidal east-west TCM velocity with east-west wind velocity (wind direction FROM); (b) PC1 of the subtidal north-south TCM velocity with east-west wind velocity (wind direction TOWARD); (c) the first EOF of the subtidal east-west TCM velocity, where colors indicate magnitude and arrows indicate direction; (d) the first EOF of the subtidal north-south TCM velocity.

acceleration on the lateral (cross-axial/north-south) flow of Greenwich Bay. The first EOF mode of the subtidal north-south TCM velocity explains 55% of the variance (Figure 2.21 (d)). The principal component of this mode has a correlation coefficient of 0.61 with the east-west wind when considering the direction that the wind is blowing toward (Figure 2.21 (b)). The magnitudes of EOF1 for subtidal north-south velocity is highest along the longitudinal axis in the outer basin, and intermediate amplitudes are calculated on the western edge of the inner basin. Stations located in the northwest corner of the inner basin, and on the shallow shoals north and south of the channel (GB01, GB12, and GB13 locations shown in Figure 2.1) are predicted to flow in the opposite direction to the others and therefore the wind. The depth of the northwest corner station is similar to adjacent stations. The subtidal current at this station may be dominated by topographic steering.

EOF findings are generally consistent with wind-driven current theory that includes the effect of Coriolis acceleration. When the east-west wind is positive (eastward) the axial near-bottom flow is negative (westward) and the cross-axial near-bottom flow is positive (northward). On the other hand, when the east-west wind is negative (westward) the axial near-bottom flow is positive (eastward) and the cross-axial near-bottom flow is negative (southward). The analysis indicates that wind forcing modified by Coriolis acceleration is a major determinant of Greenwich Bay subtidal flow. The portion of the subtidal flow that is not explained by the wind may be attributable to topographic steering and nonlinear tidal processes. Note that the wind may explain EOF findings without including the Coriolis effect, since westward wind components often occur together with northward wind components, which could cause southeastward bottom flow, and eastward wind components sometimes occur together with southward wind com-

ponents, which could lead to northwestward bottom flow.

The temporal mean horizontal flow characteristics of the Greenwich Bay subtidal near-bottom current is assessed for periods of sustained wind, i.e. events persisting for longer than a day, using spatially distributed TCM's. Three such events are identified during the 2010 TCM sampling period, namely an east-southeastward event centered on day 207.7, a north-northeastward event centered on day 216.2, and a north-northwestward event centered on day 228.2 (Figure 2.22).

The east-southeastward wind event causes the near-bottom flow to be relatively high compared to other wind events, particularly near the mouth and along the longitudinal axis (Figure 2.22 (a)). The longitudinal flow in the inner and outer basins is northwestward, directed approximately opposite to the wind. The flow in the north appears to be steered by the coastline. The flow in the northwestern corner of the inner basin is a small, local counterclockwise gyre. The gyre is consistent with a scenario where the relatively strong westward flow is steered by the northwestern alcove between Apponaug Cove and the cusp of land at Chepwanoxet Point. The western coastline causes steering of the inflowing water, such that the two stations along the southern coast in the southwestern corner of the inner basin (GB03 and GB04 locations shown in Figure 2.1) experience predominantly southward flow. Maximum current speed during the east-southeastward event occurs in the Greenwich Bay channel where the flow is into the subestuary with a magnitude of 0.10 m/s. This indicates good general agreement between the TCM response and the GBC09 response during similar wind events (compare bottom response of GBC09 for event ESE1 in Figure 2.18).

Observations show that flushing and exchange is enabled when the wind blows along the subestuary longitudinal axis (east-southeastward) in a similar direction to the surface gravitational flow. The relatively high near-bottom flow magnitudes,

reaching 0.10 m/s near the mouth and 0.04 m/s in the northwest corner of the inner basin, suggest that wind stress sets up a surface slope and an associated pressure gradient that forces bottom water in the opposite direction to the wind and in the same direction as the gravitational force. Under these circumstances the near-bottom inflow spans the outer and the inner basins and replenishment of the inner basin is enabled. This TCM result is consistent with findings at GBC09 where high two-layered exchange between the outer basin and Narragansett Bay is observed during an east-southeastward wind event. Volume flux between the outer and inner basins of Greenwich Bay is estimated by multiplying the velocity measured at each of GB07, GB08, and GB09 with a third of a cross-sectional area computed as the product of the lower half of the water column and the width of the estuary along the line of the instruments. The volume flux from the outer basin to the inner basin during the east-southeastward event shows that approximately half of the volume of Greenwich Bay, and more than the total volume of the inner basin, is exchanged in the lower half of the water column during the 2.3-day event. This suggests improved system health due to efficient replenishment of the inner basin.

Enhanced exchange promoting system health during east-southeastward wind events can be offset by an upwelling mechanism driven by these events, which is detrimental to system health [7]. By conservation of mass principles the down-estuary surface flow and up-estuary bottom flow forced by east-southeastward events cause upwelling along the western and northwestern coastline of the inner basin. A sudden upwelled pulse of hypoxic bottom water deprives upper water column organisms from oxygen, leading to mass mortality [92, 93, 94, 7].

Evidence of upwelling during an east-southeastward wind event is seen in time series of subtidal stratification and dissolved oxygen at Greenwich Bay Marina (GB) (Figure 2.23). Data show a period between day 206.7 and 211.1 when

Figure 2.22. Subtidal near-bottom flow patterns resulting from sustained (≥ 24 hours) wind events, namely (a) east-southeastward wind, (b) north-northeastward wind, and (c) north-northwestward wind. Vectors indicate the average flow for the duration of each event.

surface dissolved oxygen is reduced and bottom dissolved oxygen is increased to concentrations that exceed the surface values. This time period coincides with the east-southeastward wind event represented in Figure 2.22, which later shifts toward the north-northeast. The decrease in surface dissolved oxygen occurs before the bottom increase, indicating upwelled hypoxic bottom water with the onset of the east-southeastward wind. The density is the same at the surface and the bottom, which is consistent with horizontal pycnoclines that are upwelled. Bottom dissolved oxygen concentration starts to increase about twenty hours after the surface reduction, probably as a result of bottom water replenishment from the outer basin.

Figure 2.23. 33-hour lowpass-filtered time series of (a) subtidal stratification at Greenwich Bay Marina (GB), (b) surface (black) and bottom (grey) dissolved oxygen concentration at GB, and (c) the east-west (grey) and north-south (black) wind components at Quonset Point (Q).

The north-northeastward wind event causes a reduction in the magnitude of

Greenwich Bay subtidal flow relative to other events and leads to currents that are generally not oriented along the longitudinal axis (Figure 2.22 (b)). A local counterclockwise gyre in the northwestern corner of the inner basin is weak. The outer basin exhibits a counterclockwise flow that causes a reduction in bottom exchange with the inner basin. After entering Greenwich Bay the near-bottom water from the mouth tends southwestward (in the opposite direction to the wind) before it passes Sally Rock. Near-bottom water in the southwestern corner of the inner basin responds relatively strongly in the opposite direction to the surface flow.

Subtidal flow during the north-northwestward event tends in a direction perpendicular to the longitudinal axis of the bay at the majority of stations (Figure 2.22 (c)). An outer basin counterclockwise gyre leads to southward flow north of Sally Rock Point, causing less exchange between the outer and the inner basins compared to the north-northeastward wind event. The flow in the inner basin is southwestward in the north and southward in the south, tending opposite to the wind with modification by topography. In addition to the limited exchange between the inner and outer basins, the bottom exchange is inhibited in the channel. The north-northwestward summer event in 2010 is weaker than that observed during the autumn in 2009 when exchange in the channel is strong.

Low exchange between the inner and outer basin and low bottom flow during the north-northeastward and north-northwestward wind events suggest that these forcing conditions may be detrimental for water quality in Greenwich Bay. The magnitude of bottom flow may be important for inner basin water quality when vertical mixing and replenishment from the outer basin are inhibited. The only way to replenish bottom dissolved oxygen under these circumstances is from northward-flowing surface water within the inner basin that is downwelled as it reaches the

northern coast. With an average inner basin subtidal velocity of 0.02 m/s during a sustained north-northeastward wind, it would take a bottom parcel more than 28 hours to flow from the north to the south of the inner basin. This could lead to deterioration of water quality of the inner basin. Limited exchange between the outer basin and Narragansett Bay could similarly cause a degradation of outer basin bottom water quality.

As previously shown, the predominant wind component is northward during the summer months (Figure 2.20), which implies that conditions of low exchange in Greenwich Bay may be prevalent and could explain water quality degradation. Estimates of the time required to lower bottom water dissolved oxygen concentrations from 7.5 mg/l (not hypoxic) to 2.0 mg/l (hypoxic) by summertime respiration and benthic uptake are 5.8 hours and 1.1 days in the Apponaug and Greenwich coves, respectively [8]. Estimates are 2.9 days and 3.4 days in the mid- and outer bay, respectively [8]. Rates of dissolved oxygen consumption is thus estimated to increase from the outer to the inner basin. Inner basin consumption rates on the order of 1 to 2 days could reasonably exceed the rate of horizontal exchange.

2.3.4 Intratidal residual flow

Residual flow that varies at time scales less than a day is observed at the ADCP's in 2009 and is caused by the interaction between diurnal wind and the semidiurnal tide. The interaction appears to be linear and the water response depends on the relative stages of the wind and the tide. Net flow at time scales less than a day that is forced by the wind is important to understand for the following reasons. Firstly, it is sometimes assumed that tidal advection of water and its constituents is negligible due to the relatively symmetric nature of periodic tidal flow in estuaries with simple geometries. Secondly, vertical turbulent mixing and horizontal shear dispersion depend on vertical current variability at intratidal time

scales. Thirdly, the wind-driven response is traditionally assessed using lowpass-filtered time series with filter cutoff periods of 30 to 36 hours. It will be shown that the flow response to diurnal wind forcing is removed by lowpass filters using these traditional cutoff periods in a similar way that the 33-hour lowpass filter removes the diurnal wind and diurnal sea surface gradient signals in Figure 2.14 (a). These diurnal signals are preserved by a lowpass filter that uses a cutoff period of 20 hours (Figure 2.14 (b)), which is used in this section to identify the flow response to diurnal wind.

A number of physical mechanisms for the generation of tidal residuals are known. Tidal residuals produced via the modification of the tidal current by Coriolis acceleration, and those caused by the interaction between the tidal current and bathymetry have been extensively studied [26, 40, 41, 42, 43, 44, 45]. The effect of sustained wind on the tidal flow has been investigated to a lesser extent [44]. Diurnal wind variations may occur in estuaries that produce either negligible or non-negligible tidal residuals by these processes, but the dynamics of diurnal wind-induced residuals at intratidal time scales are not well known. Quonset Point wind data show that diurnal wind cycles can be numerous and persistent (Figures 2.3 and 2.4), indicating that their effect on the intratidal flow may be important. For example, a cycle of diurnally varying wind persisted for ten continuous days during the summers of 2009 and 2013.

The depth-dependent intratidal flow response to the diurnal wind is evident in time series of the 20-hour lowpass-filtered velocity at ADCP stations (Figure 2.24). This filtered signal contains all the energy and variability that is present in the 33-hour lowpass-filtered time series, but also includes energy from the diurnal band. Some leaked energy from the semidiurnal band appears in the signal. The response to the diurnal wind can be clearly distinguished from the purely semidiur-

nal response, which is uniform with depth and of approximately equal magnitudes during the flood and ebb. The response to sustained wind between day 220 and 226 manifests as signals with low semidiurnal velocity superimposed onto velocity that is nearly unvarying over periods exceeding a day. Successions of diurnal wind cycles between day 210 and 220, and between day 226 and 230 are observed in 2009. The start of each period of diurnally varying wind coincides with neap tidal currents. The flow response to diurnal wind can be compared to sustained wind conditions using Figure 2.24. This representation clearly shows the net flow of water on time scales less than a day.

The effect of the diurnal wind on Warwick Neck channel flow can be described as a step by step progression in time (Figure 2.24). During the first half of sea breeze day 1, following a weak southeastward wind event, the flow at WNE09 is two-layered with the upper layers flowing down the estuary and the lower layers flowing up the estuary. At WNW09 the flow is very weakly down the estuary. When the sea breeze turns on toward the north-northeast, the surface flow at WNE09 and WNW09 reverses. The rapid shift between the north-northeastward sea breeze and calm conditions toward the end of sea breeze day 1 coincides with an ebb current, which causes outflow of lower layers at WNW09 and, to a lesser extent, at WNE09. The maximum outflow corresponds to the timing of the maximum ebb current.

The northward wind corresponding to a sea breeze on day 3 coincides with two flood current peaks. At WNE09 this wind causes a weakening of the first ebb current on the day between the two floods, such that the whole water column flows in an up-estuary direction at the time of the ebb. Along the western channel the flow is two-layered with surface flow in the direction of the head and bottom flow in the direction of the mouth. At these time scales the tidal current is overcome by

Figure 2.24. Characteristics of the upper layer (solid lines) and lower layer (dashed lines) flow, due to the interaction between diurnal wind and the semidiurnal tide, are shown as 20-hour lowpass-filtered time series along the principal axes of WNE09 (c), WNW09 (d), and GBC09 (e). Time series of the near-surface tidal current at WNE09 (a) and the wind (b) are shown as reference. North-south wind components (black) and east-west wind components (grey) are shown, with positive values indicating northward and eastward wind. The orange vertical lines delineate days of diurnal wind cycles.

the wind influence at WNW09, such that the effect of the ebb current is negligible. Toward the end of day 3 the surface flow at WNW09 weakens and the bottom flow becomes strongly intensified due to a diminishing of the northward wind at the same time that an ebb current occurs. The maximum bottom down-estuary flow coincides with the timing of the maximum ebb current. Very similar wind-tidal interactions to those on day 3 occur on days 4 and 5, causing the flow structures at WNE09 and WNW09 to be similar on all three days.

Note that the flow structures resulting from wind-tide interactions on day 1 and those between day 3 and 5 manifest as high, two-layered flow at WNW09 in the 33-hour lowpass-filtered time series (Figure 2.18). The relatively high surface up-estuary flow is due to the northward wind interacting in a constructive way with the flood current, and the very high bottom down-estuary flow is due to a constructive interaction between the seizing northward wind and the ebb current. This down-estuary intensification at WNW09 can be explained as the combined effect of the gravitational force that causes down-estuary bottom flow and the ebb tide that also forces down-estuary flow.

On sea breeze days 7, 8, and 9 the relative phases of the wind and tide are different than observed during the first set of sea breeze days. Tidal ranges are similar between the first and second set of sea breeze days, but the diurnal wind characteristics are different. Periods between the northward-blowing sea breezes exhibit higher wind magnitudes of predominantly northeastward direction, as opposed to wind speeds that are near zero. The water column is also less stratified during the second set of sea breeze days. The semidiurnal tidal flow at WNE09 is offset to be predominantly positive, meaning that some ebbs cause only a weakening of the water column flow at these time scales, as opposed to reversals. The vertical water column is less sheared at WNE09 on days 7, 8, and 9 than during

diurnal wind conditions between days 1 and 6. Toward the end of sea breeze days 7 and 8 the diminishing northward wind combined with the maximum ebb current causes strong down-estuary bottom flow at WNW09.

The interaction between diurnal wind and the semidiurnal tide causes three short, repeating flow structures at GBC09 between sea breeze days 1 and 6 (Figure 2.24). The duration of each flow structure varies between ten and nineteen hours. The first structure is characterized by upper water column inflow and lower water column outflow. This structure is opposite to the gravitational flow. On sea breeze days 1, 3, and 6 the two-layered flow occurs as a response to the interaction of the maximum northward wind and a flooding current. The flood tide alone would cause the whole water column to flow into the subestuary. At these time scales the lower layers flow out of the estuary probably in response to the wind-induced pressure gradient.

The second structure at GBC09 manifests as a thin outflowing surface layer and a thick inflowing layer. Instances of this flow occurs at the end of day 1 and early on day 4. It is formed when the wind is weakly eastward or near zero between sea breezes at the same time as a flooding current. This is consistent with the modification of a flood current by eastward wind stress, but can also be explained either by gravitational circulation, or eastward wind influence alone.

A third flow structure at GBC09 has a relatively thick and strong outflowing upper water column layer and a thinner weak inflowing layer. This occurs when the wind is northeastward between sea breezes at the same time as the ebbing current. It is observed at the end of days 2, 4, and 5. Comparing the second and third flow structures the effect of the tide becomes apparent. The wind contribution to the flow is similar, but the stages of the tide are different. The wind influence establishes the two-layered surface outflow and bottom inflow for both the second

and third structures, whereas the tide determines the strength and thickness of the inflowing and outflowing currents. The flood current causes a relatively weak and thin surface layer outflow, whereas the ebb current causes a thick and relatively strong upper water column outflow. Near the start of day 5 the spring flooding current and weak northeastward wind causes the whole measured water column to flow into the subestuary, but the wind influence is still evident in the high vertical shear.

2.3.5 Interannual subtidal characteristics

The intensified water response at WNW09 resulting from the interaction between diurnal wind and the semidiurnal tide can explain the disparity in subtidal current speeds at the ADCP in the center of the Warwick Neck channel during 2006, here referred to as WNC06 [19]. The 36-hour lowpass-filtered time series exhibited relatively strong subtidal flow during the first part of the 2006 sampling period up to day 220, followed by a sudden decrease in flow intensity during the second period. The 33-hour lowpass-filtered time series at WNC06 and WNW09 are compared (Figure 2.25). The WNW09 time series shows a similar decrease in current speeds roughly from day 230 onward. Mean flow speeds decrease from 0.037 m/s and 0.034 m/s during the first periods to 0.023 m/s and 0.027 m/s during the second periods at WNC06 and WNW09, respectively. Maximum flow speeds decrease from 0.26 m/s and 0.27 m/s during the first periods to 0.14 m/s and 0.12 m/s during the second periods at WNC06 and WNW09, respectively. The flow response to sustained wind events during the second periods, and the characteristics of flows during the first periods are qualitatively similar between WNC06 and WNW09. During both the 2006 and 2009 sampling periods a two-layered flow intensification during the first period occurs together with diurnal wind variability.

The relatively high current speeds at WNW09 prior to day 230 are attributed

Figure 2.25. Comparison of the 33-hour lowpass-filtered time series along the principal axes at (b) WNW09 and (d) WNC06. Red indicates the flow in the uppermost bin and blue indicates the flow in the lowermost bin. The wind at Quonset Point (Q) corresponding to the sampling periods in (a) 2009 and (c) 2006 is shown with positive values representing northward and eastward wind. Black is the north-south wind component and grey is the east-west wind component. The orange blocks delineate periods of increased flow speeds during sea breeze days.

to constructive wind-tide forcing, as discussed in Section 2.3.4. The 20-hour lowpass-filtered time series at WNC09 show that the same mechanism can explain high current speeds prior to day 220 in 2006 (Figure 2.26). The highest down-estuary bottom flow speeds are observed at the beginning of sea breeze days 1, 2, and 3. These strong currents occur when a relaxation in the northward sea breeze coincides with the maximum ebb current. The response can be described by the combined effect of the gravitational force, which causes down-estuary bottom flow following a relaxation of the northward wind, and the maximum ebb current that also forces down-estuary flow. This is the same mechanism that causes the increased down-estuary bottom flow speeds at WNW09. A less dramatic response during sea breeze days 7, 8, and 9 may be due to different diurnal wind characteristics and relative wind and tide phases acting in a less stratified water column. The sea breeze days are shown to facilitate the discussion. Flow intensification could also result from synoptic wind shifts when they occur at time scales less than a day.

2.4 Conclusion

Greenwich Bay exhibits a low tendency for vertical turbulent mixing due to low tidal energy. Wind with diurnal variability produces an increase in the vertical shear of tidal currents. The increase in vertical tidal current shear leads to an increased tendency for vertical turbulent mixing in the Warwick Neck and Greenwich Bay channels.

The wind and geometry strongly determine subtidal flow characteristics in Greenwich Bay. Predominantly northward wind causes isolation of the inner basin and reduced exchange between the outer basin and Narragansett Bay. Wind with a dominant eastward component promotes exchange between the Greenwich Bay basins and Narragansett Bay.

Figure 2.26. Characteristics of the upper layer (solid lines) and lower layer (dashed lines) flow to the interaction between diurnal wind and the semidiurnal tide are shown as 20-hour lowpass-filtered time series along the principal axis of WNC06 (c). Time series of the near-surface tidal current at WNC06 (a) and the wind (b) are shown as reference. North-south wind components (black) and east-west wind components (grey) are shown with positive wind indicating northward and eastward wind. The orange vertical lines delineate days of diurnal wind cycles.

The north-south wind component determines the subtidal longitudinal sea surface elevation gradient in Narragansett Bay. The sea surface elevation gradient responds both to diurnal wind variability and sustained wind events.

Subtidal flow in the Warwick Neck channel is dominated by wind forcing. A characterization of the complicated lateral and vertical subtidal variability requires observations from two ADCP's despite the narrowness of the channel. Sustained southwestward wind leads to an increase in the thickness and speed of the density-driven down-estuary surface flow on the eastern side of the Warwick Neck channel and an increase in speed of the density-driven down-estuary water column on the western side of the channel. Persistent southwestward events can force a reversal of the mean bottom flow in the east, causing the entire channel flow to be directed toward the estuary mouth. Sustained northward wind forces a low magnitude, depth-uniform response in the Warwick Neck channel with up-estuary flow in the east and down-estuary flow in the west. High north-northwestward wind causes density-driven flow in the west to reverse toward the estuary head, such that the entire channel flow is in an up-estuary direction.

The interaction between diurnal wind and the semidiurnal tide produces vertically sheared intratidal residuals in the Warwick Neck and Greenwich Bay channels. These residuals depend on the relative phases of the wind and tide. Phases that act constructively cause a dramatic increase in the flow magnitude on the western side of the Warwick Neck channel.

Physical processes observed in Greenwich Bay explain the high incidence of water quality impairment during the summer. Low vertical turbulent mixing and weak circulation can lead to prolonged isolation of the bottom water, preventing replenishment of oxygen by the atmosphere.

Subtidal flow responses in the Warwick Neck and Greenwich Bay channels

indicate that only sustained southward and southwestward wind events are likely to cause an influx of contaminated water from northern subsystems into Greenwich Bay. These events occur relatively infrequently during the summer.

Low horizontal exchange in Greenwich Bay that can reduce replenishment with ventilated or healthy water from adjacent systems is forced by dominant summer wind conditions. Occasional eastward summer wind events could improve water quality as a result of enhanced exchange, but could also cause organism mass mortality when hypoxic water is upwelled along the west coast of Greenwich Bay.

The following recommendations can be made. Numerical tracer modeling can substantiate inferences about the effect of observed flow characteristics on tracer dispersion and water quality. The traditional lowpass-filtered time series with cutoff periods of 30 to 36 hours should be used with caution to interpret the wind-driven circulation. Flow forced by the interaction between the diurnal wind and the semidiurnal tide may be misrepresented by these time series. Care should be taken in the interpretation of underway ADCP surveys conducted on diurnal wind days. The tidal average would represent an intratidal residual flow, as opposed to a mean estuarine, or sustained wind-induced flow.

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MANUSCRIPT 3

**Modeling the relationship between Greenwich Bay circulation, flushing
efficiency, and environmental forcing conditions**

by

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Abstract

Model simulations are used to develop relationships between dominant environmental forcing conditions and the spatial and temporal flushing characteristics of Greenwich Bay. We use the Regional Ocean Modeling System (ROMS) with high grid resolution in subsystems of Narragansett Bay that are subject to poor water quality. In Greenwich Bay the horizontal resolution of the model is approximately 100 m and the vertical resolution ranges from 0.1 m to 0.7 m. We employ a semi-idealized strategy that excludes natural high-frequency signals from the secondary forcing specifications and develop idealized or synthetic forcing time series to conduct scenario experiments. Numerical passive dye tracers are used to quantify flushing rates and to study dispersion as a function of wind, tides, and runoff. Results show high variability in dye residence times, particularly as a function of wind direction. This variability is explained by differences in the mean advective characteristics. Relationships between forcing conditions and dispersion indicate several physical predispositions of Greenwich Bay to poor water quality. The tide forces a mean outer basin recirculation feature and an isolated inner basin in the scenario where no wind is applied. The dominant summer wind forces retentive flow structures in both the inner and the outer basin, due to differences in the orientation of the longitudinal axis of the subestuary and the applied wind stress. Typical summertime diurnal wind that has a daily averaged north-northeastward direction improves flushing compared to a sustained northward wind, particularly in the inner basin where residence times are about 15 days and 30 days, respectively. Residence times produced by the diurnal wind are nonetheless high compared to the time for oxygen consumption in Greenwich Bay, which is approximated to be less than five days.

3.1 Introduction

Greenwich Bay is a subestuary of Narragansett Bay that undergoes severe water quality degradation during the summer months. Water quality problems include recurrent hypoxia events and intermittent bacteria impairment [1, 2, 3, 4, 5, 6]. These can lead to organism mortality and beach closures, and have caused ecosystem degradation [7, 4]. Studies generally focus on the biological and chemical aspects of water quality, but the physical controls must also be understood to effectively manage the system. The effect of different environmental forcing conditions on the water quality of Greenwich Bay is the focus of this study.

Several physical mechanisms for water quality impairment exist. Higher water temperatures during the summer increase stratification and reduce the capacity for water to retain dissolved oxygen compared to the winter [4]. A stratified water column can lead to isolation of the bottom water and a consequent reduction in the potential for atmospheric oxygen replenishment [8, 9, 10, 11, 12]. Freshwater runoff can increase stratification and nutrient input [6], and storm-related runoff events can introduce large quantities of untreated water into an urban system, which can lead to bacterial contamination of the system. The water quality of a subestuary can also become compromised if water is not replenished through exchange with the parent estuary. Low replenishment rates are particularly detrimental to urban water bodies such as Greenwich Bay that receive high nutrient loads and are susceptible to high bacteria influx during periods of stormwater runoff [1, 13, 14].

Dispersion characteristics of passive numerical dyes are assessed to determine Greenwich Bay exchange. Dispersion is defined as the spreading of constituents through the water by all physical mechanisms [15]. The method employed in this study is to use a high-resolution, three-dimensional Regional Ocean Modeling System (ROMS) [16, 17] model to conduct forcing scenario experiments. The approach

is semi-idealized in the sense that secondary forcing conditions, i.e. those that remain unchanged between scenarios, are smoothed to remove natural high-frequency signals that may contaminate the response to the primary forces being tested. The primary forces are idealized time series that represent sustained conditions typical of Narragansett Bay. The semi-idealized strategy allows for easy identification of the effect of an individual force on circulation and flushing in Greenwich Bay. Dye residence times enable quantitative comparisons of the flushing characteristics of different scenarios.

The duration of dissolved oxygen consumption in Greenwich Bay is used as a reference for dye residence times to provide a link between modeled flushing rates and water quality. Estimates have been made of the time that it takes to lower dissolved oxygen by water column respiration and benthic oxygen uptake from a value of 7.5 mg/l (not hypoxic) to a value of 2.0 mg/l (hypoxic) [1]. The estimation assumed darkness, constant rates of respiration, and a lack of oxygen input from surface waters or adjacent areas. Estimates for the Apponaug Cove, Greenwich Cove, and the middle and eastern regions of Greenwich Bay (Figure 3.1) were 5.8 hours, 1.1 days, 2.9 days and 3.4 days, respectively. The values show that rates of dissolved oxygen consumption increase from the outer to the inner basin. These time scales provide a qualitative reference for defining conditions when advective resupply of oxygen from offshore waters is efficient versus inefficient relative to *in-situ* oxygen consumption. Inner basin residence times exceeding three days and outer basin residence times exceeding five days are assumed to indicate a high risk for water quality degradation of each respective basin.

The dye residence times are further referenced with previous estimates of flushing in Greenwich Bay. Two-layered box models estimated that it takes 4 – 9 days to flush a third of the volume of Greenwich Bay [18, 1, 13]. Studies that

Figure 3.1. Bathymetry map of Greenwich Bay with important landmarks shown, including the East Greenwich wastewater treatment facility (WWTF). Note that color axis limits are assigned to emphasize variability; values outside the limits are clamped to the first or last color. (Bathymetry data from <http://www.geomapapp.org>)

employed two-dimensional models also estimated that Greenwich Bay flushing was relatively efficient, calculating e-folding times of 3 – 9 days [19, 20]. Another modeling study conducted three-dimensional wind scenario testing and found that specific wind conditions lead to residence times as high as 30 days [21].

Greenwich Bay is characterized by a channel entering at the mouth between very shallow regions in the northeast and southwest corners of the bay (Figure 3.1). The bay depth gradually decreases westward of the channel along the northwest-southeast aligned longitudinal (lengthwise) axis. The maximum channel depth is 11 m and the average depth of Greenwich Bay is 2.4 m [1]. Two natural basins can be distinguished, which are referred to as the inner basin and the outer basin. A number of small freshwater tributaries enter at each of the coves. The largest of these are Hardig Brook that discharges into Apponaug Cove, and the Maskerchugg that flows into Greenwich Cove [22]. The East Greenwich wastewater treatment facility (WWTF) is located in Greenwich Cove.

3.2 Method

The semi-idealized model was built from a validated, realistic Regional Ocean Modeling System (ROMS) [16, 17] model for Narragansett Bay (NB-ROMS model) [23] and forced with smoothed boundary conditions, freshwater runoff, and meteorological properties that are used in bulk formulations to compute surface heat and momentum fluxes. These specifications represented forcing conditions that remained unchanged between the reference run and wind scenario runs. A 31-day (year days 151 – 182) spin-up of the model was done with this forcing using a NB-ROMS initialization file. Boundary salinity and temperature input exhibited only seasonal variability in the semi-idealized model. Boundary tide height and tidal currents were obtained from extracting twelve of the major tidal constituents from the NB-ROMS boundary input and adding them to annual averages. Forcing functions specified for the bulk parameterization of surface heat and momentum fluxes included annual average values for air pressure and humidity, seasonally varying air temperature, zero rainfall, as well as longwave and shortwave radiation fluxes that varied at diurnal and seasonal time scales. In addition to the twelve freshwater point sources in the NB-ROMS model that represented freshwater flux for the greater Narragansett Bay, two point sources were included in Greenwich Bay. The two freshwater sources represented the two rivers located at the head of Greenwich Bay, namely Hardig Brook and the Maskerchugg. Transport of the East Greenwich wastewater treatment facility in Greenwich Cove was added to the Maskerchugg point source in the model. Groundwater estimates for Apponaug Cove and Greenwich Cove [22] were also added to the respective point sources. All rivers were specified to discharge at constant long-term averaged rates.

Semi-idealized model scenarios differed in their specifications regarding the wind and tide. A unique forcing time series was applied after the 31-day spin-up

Table 3.1. List of model scenario runs.

Run ID	Forcing scenario
REF	Reference case (no wind)
NOTIDE	Tidal forcing removed from REF (no wind)
NE-WIND	REF with constant (4 m/s) northeastward wind
E-WIND	REF with constant (4 m/s) eastward wind
SE-WIND	REF with constant (4 m/s) southeastward wind
S-WIND	REF with constant (4 m/s) southward wind
SW-WIND	REF with constant (4 m/s) southwestward wind
W-WIND	REF with constant (4 m/s) westward wind
NW-WIND	REF with constant (4 m/s) northwestward wind
N-WIND	REF with constant (4 m/s) northward wind
SB1-WIND	REF with diurnal wind (avg dir = east-northeastward; avg spd = 4.8 m/s)
SB2-WIND	REF with diurnal wind (avg dir = north-northeastward; avg spd = 3.8 m/s)

for each scenario run (Table 3.1). Cases included a run with zero wind, which was used as a reference for runs that tested the effect of different constant (or sustained) wind directions and diurnally varying wind. Diurnal wind scenarios were included due to the high occurrence of diurnally varying wind associated with sea breezes across Narragansett Bay in the summer. A scenario where the tidal forcing was removed from the reference case was run to effectively interpret flushing and dispersion produced by the reference run.

The reference run was thus a condition that included tidal and density forcing, but excluded wind. The wind direction scenarios contained tidal and density forcing, in addition to constant (4 m/s) wind derived from long-term averaged observed wind. The diurnal wind scenarios included tidal and density forcing, as well as diurnally varying wind. Idealized diurnal wind remained unchanged from one day to the next and was based on observations. Two diurnal wind cases were investigated, which were similar in average speed to the sustained wind cases (3.8

m/s and 4.8 m/s), but varied in their long-term averaged wind direction.

Another spin-up of one day was conducted for each scenario run with the unique forcing condition specified. A one day spin-up for the wind scenarios was sufficient in an area where the lag between wind and water response is less than three hours [21]. After a one day spin-up for the case that excluded the tidal force, less than 1% of background tidal energy remained detectable.

Numerical dye patches were released in different locations of the bay, but all dyes were released at the same time. Dyes were specified in the upper and lower half of the water column, here referred to as surface and bottom dye patches, to assess vertical flushing characteristics. Dyes were located in the four quarters of Greenwich Bay, as well as across the entire basin, to test the spatial variability of dispersion (Figure 3.2). Release occurred after the one day spin-up, on day 183 (3 July), during a neap tidal period at the timing of the maximum ebb current. The neap stage of the tide and the ebb current were chosen for the release of the dye, because studies showed that both the neap and the ebb were associated with increased stratification and consequent reductions in vertical mixing, which could contribute to impaired water quality [8, 9, 10, 11, 12]. Releasing the dye during other stages of the tide caused differences in residence times for each model scenario, which were slight for some cases and large for others. The effect of the timing of dye release on residence time is not discussed in this paper.

Residence time was defined as the time that it took for the concentration of a dye patch to decrease to $1/\exp(1)$ of its original value in the Greenwich Bay basin. The calculation was defined to assess flushing from the entire Greenwich Bay system, regardless of the the initial positions of dyes. Since the rate of oxygen consumption is high in both the inner and the outer basin, it was assumed that impaired water from the inner basin would not be significantly improved when

Figure 3.2. Release locations of numerical dye patches designated northwestern (NW), northeastern (NE), southwestern (SW), and southeastern (SE) quarter patches. A dye patch that fills the entire Greenwich Bay basin, named Full, was also released. At each location a dye patch was released in both the upper water column and the lower water column, such that a total of ten dyes were released in Greenwich Bay for each model run.

dispersed into the outer basin. Residence time calculations employed time series of dye output at 6-hour intervals.

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3.3 Results and discussion

3.3.1 Residence times of dyes

Residence times of dye patches show the effect of different forcing conditions on the flushing of distinct areas of Greenwich Bay (Figure 3.3). One of the notable results obtained from dye residence times is that flushing efficiency is spatially variable, both in the horizontal and vertical directions. Another important finding is that flushing strongly depends on the applied wind direction.

Vertical flushing characteristics depend on the applied forcing conditions. In the majority of scenarios a surface dye patch released in a specific quarter of Green-

Figure 3.3. Residence times of dye patches subject to different forcing functions, namely the scenario without tidal forcing (NOTIDE), the reference scenario (REF), sustained wind scenarios named for the direction that the wind is blowing toward (e.g. NE represents a northeastward wind), and diurnal wind scenarios where SB1 has an average east-northeastward direction and SB2 has an average north-northeastward direction. Surface (solid) and bottom (dashed) colored lines distinguish between dye patches named for the release locations (e.g. NW_s is for the northwestern quarter surface dye patch). The black horizontal dashed lines indicate the estimated times for dissolved oxygen uptake in the (i) inner and (ii) outer basins.

wich Bay flushes before the bottom dye patch that is released in the same quarter. Bottom patches flush prior to their surface counterparts in the case of a sustained westward wind scenario, most notably the bottom dye that is specified across the entire bay, which flushes 1.5 days earlier than the corresponding surface dye. When diurnal wind is applied the surface and bottom dye patches of a specified quarter flush at the same rates.

Results show two distinct horizontal flushing regimes in Greenwich Bay. One regime is characterized by the flushing of the outer basin prior to the inner basin, and the other constitutes flushing of the southern half of the bay prior to the northern half. The NOTIDE run produces the latter regime, whereas the REF run brings about the first regime. In the majority of cases the inner basin flushes later than the outer basin. This is most evident in cases of sustained wind directed toward the north and west, and diurnal wind (SB2-WIND) that cause the surface patch in the northeastern quarter to flush before the surface patch in the southwestern quarter by 12.5 days, 10 days, and 9.3 days, respectively. The northwestern and southwestern quarter patches generally flush at similar rates, but when the wind is in a sustained westward direction there is a discrepancy of about seven days between the residence times of these patches. Sustained wind toward the east, southeast, and south cause the southern half of the bay to flush prior to the northern half, but the differences in residence times of all dye patches are relatively small. The northeast and northwest quarter surface patches both take 2.4 days longer than the southwest quarter surface patch to flush when the wind is eastward.

The bimodal flushing style produced by the wind can be explained by the relative angle between the wind direction and the longitudinal axis of the bay, in addition to Coriolis and tidal influences. Delayed flushing of the northern halve of Greenwich Bay relative to the southern halve indicates a response to Coriolis

acceleration that produces net inflow along the northern coastline and net outflow along the southern coastline. Wind that promotes this flow regime is directed roughly along the longitudinal axis. Although the same regime is expected for the reference run, the REF case causes the outer basin to flush prior to the inner basin due to tidal effects. Scenarios where the wind is directed approximately normal to the longitudinal axis also lead to higher residence times in the inner basin relative to the outer basin.

The residence times produced by the reference case are used for comparison with other forcing scenarios. Only the southeastern quarter surface patch flushes within the time frame expected to promote water quality when the reference case is applied. In the reference scenario the difference in residence times of the southeastern quarter bottom and surface patches is 2.3 days. The northeastern quarter surface patch flushes 3.2 days after the southeastern quarter surface patch. Inner basin residence times of the reference run exceed eight days. This suggests that Greenwich Bay is predisposed to poor flushing and consequent water quality problems, particularly in the inner basin.

Only a small window exists where wind direction is favorable to flushing of Greenwich Bay. The southwestern and southeastern quarter patches flush in under 2.6 days when the sustained wind is oriented toward the southeast and the east. The other dye patches take between 2.6 days and 4.6 days to flush during sustained southeastward and eastward wind. Wind toward the south causes an increase in flushing relative to the reference case and the maximum residence time is 5.6 days for the northeastern quarter bottom patch. The highest residence times for the southeastward, eastward, and southward wind conditions may be acceptable for water quality, considering that the cited times for dissolved oxygen uptake are for upper limit uptake rates. The residence times for sustained wind conditions

suggest another predisposition of Greenwich Bay to poor water quality, namely that the dominant wind direction over Narragansett Bay in the summer is rarely southeastward, eastward, or southward.

The majority of sustained wind conditions leads to residence times that are detrimental for water quality. Flushing is most dramatically impaired when the wind blows toward the north. Residence times produced by a constant northward wind exceed 30 days in the inner basin. Inner basin dyes take between 21 days and 23 days to flush when a sustained northwestward wind is applied. The summer wind commonly has a dominant northward component. It is therefore expected that horizontal replenishment of water is impaired by the dominant summer wind.

Observations of summer wind over Narragansett Bay show a high incidence of diurnally varying wind [24]. The diurnal wind is associated with the growth and subsidence of sea breezes that exhibit predominantly northward directions in the afternoon due to differential ocean and land heating during the course of a day. The effect of diurnal wind with an average north-northeastward direction (SB2-WIND) is to increase flushing by a factor of two or more relative to the sustained northward wind case. A diurnal wind with an average east-northeastward direction (SB1-WIND) increases flushing even more with residence times resembling those produced under sustained northeastward wind.

Residence time characteristics present the following questions whose answers are summarized here, but are discussed in more detail in the remainder of the section:

- *What causes the differences in residence times of some surface and bottom dye patches?* The differences in surface and bottom dye patch dispersion are generally due to vertical two-layered advection forced by gravitational circulation in the NOTIDE and REF scenarios, and the wind in cases like

E-WIND and W-WIND.

- *What determines the horizontal flushing regimes where (i) the outer basin flushes prior to the inner basin, or (ii) the southern half of the bay flushes prior to the northern half of the bay?* The two horizontal flushing regimes occur as a result of differences in depth-averaged horizontal advective characteristics determined by the relative direction between the wind and the longitudinal axis of the bay. For example, a northward wind is aligned approximately perpendicular to the longitudinal axis, producing two distinct recirculation gyres that separate the inner and outer basin flow and cause the outer basin to flush first. Southeastward wind is aligned approximately parallel to the longitudinal axis, forcing a counterclockwise flow that extends from the mouth to the head of the bay, which allows the southern dyes to flush first.
- *Why is there a large discrepancy between the residence times of northwestern and southwestern quarter patches when the wind is directed toward the west?* Westward wind causes a separation in flow regimes of the northern and southern parts of the inner basin. A counterclockwise flow in the northwestern quarter enables exchange of inner basin dye with the outer basin, but a relatively weak clockwise flow in the southwestern quarter causes dye to become retained.
- *Why are high residence times obtained for the reference scenario, particularly in the inner basin?* High residence times produced by the reference run can partially be attributed to low depth-averaged velocity, particularly in the inner basin. Lower flushing in the inner basin relative to the outer basin is due to the interaction of the tide or the mean flow with topography, which

forces an outer basin recirculation structure that preferentially flushes the outer basin.

- *What are the mechanisms that lead to increased flushing when the wind is sustained southeastward and decreased flushing when the wind is sustained northward?* The relative alignment of the southeastward and northward wind to the longitudinal axis of the bay leads to differences in the orientation of the two-layered flow, explaining the higher residence times in the N-WIND case relative to the SE-WIND case. The two-layered flow is along the longitudinal axis in the SW-WIND case, and perpendicular to it in the N-WIND case.
- *By what means does diurnal wind improve flushing relative to sustained northward wind?* Flushing is improved during diurnal wind relative to sustained northward wind due to two-layered flow that is slightly less perpendicular to the longitudinal axis, allowing for dye to be more readily advected to the mouth.

3.3.2 Effect of tidal forcing on the mean flow

The REF run produces flushing characteristics that suggest that Greenwich Bay is predisposed to poor water quality, which can be interpreted by comparing the NOTIDE case. Inner basin residence times are higher when tides are absent relative to the reference scenario. The residence time of the bottom dye patch in the northwestern quarter is 13 days in the NOTIDE run and 9 days in the REF run. The lower dispersion rates of the inner basin dyes in NOTIDE can be attributed to a difference in the mean flow structure compared to REF. Additionally, it is expected that vertical turbulent mixing and shear dispersion is reduced in the absence of tidal current shear. Reduced vertical turbulent mixing in the NOTIDE run can be inferred from larger differences between residence times of surface and bottom dye

patches compared to the REF case (Figure 3.3). The difference in residence times between the surface and bottom dye patches released in the northwestern quarter is 12 days and 30 days in the REF run and NOTIDE run, respectively.

Differences in the dye patch flushing sequences of the NOTIDE and REF scenarios can be assessed to explain the discrepancy in flushing rates. Dye patch flushing in NOTIDE suggests counterclockwise advection throughout Greenwich Bay with the southeastern patch flushing first, the southwestern patch second, the northwestern patch third, and the northeastern patch fourth. The sequence is the same for the upper and lower water columns. The flushing sequence in the REF run indicates a separation between inner and outer basin flow regimes. Dye dispersion shows that the outer basin flushes before the inner basin as follows: the southeastern patch flushes first, the northeastern patch second, the southwestern patch third, and the northwestern patch fourth. This sequence occurs in both the upper and lower water columns.

Velocity vectors produced by the NOTIDE case indicate the processes governing the flow response (Figure 3.4). Depth-averaged velocity vectors of the NOTIDE run show a counterclockwise horizontal structure extending through much of the length of Greenwich Bay (Figure 3.4 (f)), suggesting that the depth-averaged flow determines the dye patch flushing sequence. Depth varying circulation is consistent with two-layered gravitational flow that is modified by topography. Near-surface flow is generally in the direction of the estuary mouth (Figure 3.4 (b)) and the near-bottom flow is in the direction of the estuary head (Figure 3.4 (d)). Near-surface flow experiences southward deflection and forms a counterclockwise recirculation near the mouth. The deflection and recirculation is likely due to the convergence of flow over a depth varying bottom in the northeastern quarter, as well as the down-estuary flow that encounters the coastline at Warwick Neck. Southward deflection

is consistent with the effect of Coriolis acceleration.

Velocity vectors forced by the REF and NOTIDE cases exhibit notable differences. Depth-averaged velocity of the REF scenario exhibits a counterclockwise flow structure in the outer basin with speeds in the northeastern and southeastern parts of Greenwich Bay notably higher than those computed in NOTIDE (Figure 3.4 (e)). The spatial- and depth-averaged velocity in the inner basin of Greenwich Bay increases from 0.017 m/s in NOTIDE to 0.046 m/s in REF. The depth varying velocity of the REF scenario resembles topographically modified, two-layered gravitational flow (Figure 3.4 (a) and (c)), but with some notable differences compared to the NOTIDE case. Flow structures of the surface and bottom layers produced by the REF case are similar to the NOTIDE case, but they exhibit increased speed. Near-surface flow of REF displays a more robust and spatially extensive counterclockwise flow structure in the outer basin than is produced by NOTIDE.

General flushing characteristics of the REF run can be explained by the depth-averaged circulation, but some aspects require further consideration. The discrepancy in residence times between the surface and bottom patches released in the southeast quarter is due to the fact that the bottom patch experiences a net advection toward the estuary head, but not the surface patch. The up-estuary advection causes a portion of the bottom dye to become trapped in the inner basin. The relatively large difference between the residence times of surface patches released in the northeast and the southeast in REF can be attributed to the effect of the tide on the mean dispersion. Northeastern surface dye is advected westward by the surface recirculation feature. This westward advection is enhanced by the flood tidal current, which causes dye in the northeastern quarter to be advected into the inner basin where it becomes trapped.

The differences between the REF and NOTIDE runs suggest that tidal forcing

Figure 3.4. The effect of tidal forcing on the mean flow structure is characterized by 17-day average velocity vectors of the reference scenario (REF), which excludes wind forcing and contains tidal and density forcing, compared to the scenario that differs only in that the tidal forcing is absent (NOTIDE). (a) and (b) are near-surface vectors of the REF and NOTIDE runs, respectively; (c) and (d) are near-bottom vectors of the REF and NOTIDE runs, respectively; and (e) and (f) are depth-averaged vectors of the REF and NOTIDE runs, respectively. Black arrows conceptualize the flow.

plays a role in the temporal mean circulation and dispersion of Greenwich Bay. Tidal forcing increases vertical turbulent mixing and the mean two-layered flow, consistent with conservation of mass theory [25]. Tidal forcing also enhances counterclockwise circulation in the outer basin by the interaction of the increased mean flow with topography and/or by tidal rectification. This outer basin recirculation preferentially flushes the outer basin and the weak inner basin flow limits the transport of inner basin dye to the outer basin in the reference scenario. Advective dispersion from the inner to the outer basin is achieved via a narrow pathway in the south.

3.3.3 Flow and dispersion forced by sustained wind

In all wind direction scenarios the wind forces two-layered flow consistent with wind-driven circulation theory in estuaries [26, 27]. The surface flow tends in the direction of the wind, which sets up a sea surface slope and an associated pressure gradient force in the opposite direction to the wind. The pressure gradient force causes bottom water to flow in the opposite direction to the wind. The two-layered flow is modified by topography and Coriolis acceleration. Wind-driven flow dominates circulation, such that the effects of gravitational forcing and tidal forcing become small. These characteristics can be explained in the context of extreme flushing scenarios by comparing the SE-WIND run and the N-WIND run.

The circulation produced when a constant southeastward wind is applied is shown in Figure 3.5. The near-surface flow is in a down-estuary direction in the southern half of the bay (Figure 3.5 (a)). Near-bottom water flows toward the estuary head (Figure 3.5 (c)). The structure of the near-bottom flow agrees with the observed structure under similar sustained wind conditions [28]. Both the near-surface and the near-bottom layers appear to be deflected to the right of the flow paths, suggesting that the effect of Coriolis acceleration is non-negligible. The

depth-averaged flow exhibits a robust counterclockwise structure that promotes exchange between the inner and the outer basin. The structure is attributed to the fact that the wind is in a direction nearly parallel to the longitudinal axis of the estuary and therefore nearly parallel to the two-layered gravitational flow. The depth-averaged flow is consistent with the flushing sequence of dye in the SE-WIND run, which also follows a counterclockwise pattern.

The flow response to constant northward wind is approximately perpendicular to the longitudinal axis of Greenwich Bay. Near-surface flow again tends in the direction of the wind (Figure 3.5 (b)), and near-bottom flow tends in the opposite direction to the wind (Figure 3.5 (d)). The structure of the near-bottom flow agrees with the observed structure under similar sustained wind conditions [28]. The surface and bottom circulation are deflected less than in the SE-WIND case, suggesting that Coriolis acceleration has less of an influence on the flow. The depth-averaged velocity shows very low currents in the inner basin that form a closed clockwise structure (Figure 3.5 (f)). The spatial- and depth-averaged velocity in the inner basin is 0.042 m/s. The clockwise circulation feature extends into a portion of the outer basin. A counterclockwise flow is produced toward the eastern side of the outer basin. Surface and bottom velocity vectors suggest that the counterclockwise structure is formed from the northward surface flow that encounters the shallow shoal in the northeastern corner of Greenwich Bay and a pressure gradient forcing bottom flow southward. Flow vectors indicate that exchange between the inner/mid-basin and the eastern side of the outer basin is very low. The depth-averaged circulation is consistent with the sequence by which dye patches flush. The southeastern, northeastern, northwestern and southwestern bottom patches take 9.33 days, 20.3 days, 31.1 days, and 31.5 days to flush.

Depth-averaged vertical velocity characteristics for the SE-WIND and N-

Figure 3.5. The effect of wind direction on the mean flow structure is shown with 17-day average velocity vectors of the SE-WIND scenario compared to the N-WIND scenario. (a) and (b) are near-surface vectors of the SE-WIND and N-WIND runs, respectively; (c) and (d) are near-bottom vectors of the SE-WIND and N-WIND runs, respectively; and (e) and (f) are depth-averaged vectors of the SE-WIND and N-WIND runs, respectively. Black arrows conceptualize the flow.

WIND scenarios exhibit large differences. A sustained southeastward wind causes two-layered flow and associated upwelling along the western and northwestern coastlines, as is expected for mass to be conserved (Figure 3.6 (a)). Upwelling of inner basin hypoxic bottom water under eastward and southeastward wind conditions has been hypothesized to cause mass mortality of near-surface organisms in Greenwich Bay by entrapment [4]. A sustained wind toward the north produces downwelling along the northern coastline and on the northeastern shoal of the outer basin, and upwelling along the southern coastline and on the southeastern shoal of the outer basin (Figure 3.6 (b)).

Vorticity maps show the recirculation features that enable exchange in the SE-WIND case and inhibit exchange in the N-WIND case. A high maximum depth-averaged positive vorticity value of 18.1×10^{-5} 1/s is found along the longitudinal axis of the bay when the flow is subject to a constant southeastward wind (Figure 3.6 (c)). Extensive depth-averaged negative vorticity is produced in the inner and mid-basin when the wind is northward (Figure 3.6 (d)). Minimum vorticity in the inner basin is -14.7×10^{-5} 1/s. Positive vorticity occurs toward the eastern side of the outer basin for the N-WIND scenario.

Disparate flushing of the SE-WIND and N-WIND scenarios is evident in the dye dispersion of each case, which can be explained in terms of their advective structures. Positive vorticity along the longitudinal axis in the SE-WIND case facilitates the flushing of the northwestern quarter bottom dye patch along the southern half of the bay, such that after four days only 26% of dye remains (Figure 3.6 (e)). The effect of the counter-rotating features in the N-WIND case is to trap the northwestern quarter bottom dye patch within the inner basin for extended periods of time, such that 98% of the dye remains after four days (Figure 3.6 (f)). The inner basin circulation feature also traps concentrations of dye patches released

Figure 3.6. 17-day averages of: depth-averaged vertical velocity from (a) the SE-WIND scenario and (b) the N-WIND scenario; depth-averaged vorticity produced by the (c) SE-WIND case and (d) N-WIND case; and the spatial dispersion of the northwestern quarter bottom dye patch four days after release in the (e) SE-WIND run and (f) the N-WIND run. Shown is the dye patch concentration averaged across the lower half of the water column. Note that color axis limits are assigned to emphasize variability; values outside the limits are clamped to the first or last color.

in the outer basin when they are advected into the inner basin via the tidal current. The outer basin counterclockwise rotation has a retentive effect on dyes released in the outer basin, since the direction of depth varying flow is perpendicular to the longitudinal bay axis.

The westward wind scenario brings about residence time characteristics that are generally different from the other cases. The W-WIND scenario produces flow that causes the bottom dye patches to flush faster than their surface counterparts. In the W-WIND case the discrepancy in residence times between the northwestern and southwestern quarter dyes is larger than in the other cases, namely eight days compared to less than four days.

Velocity vectors show that a sustained westward wind leads to two-layered flow and depth-averaged recirculation features. The wind forces a westward near-surface flow, which is modified by topography (Figure 3.7 (a)). The near-bottom flow is consistent with a response to a pressure gradient force directed in the opposite direction to the wind and deflection by Coriolis acceleration (Figure 3.7 (b)). Depth-averaged flow exhibits a high counterclockwise flow in the outer basin, and a lower counterclockwise structure that extends to the northwest corner of the inner basin (Figure 3.7 (c)). A relatively weak clockwise flow extends throughout the southwestern quarter of the bay.

Dye dispersion by the westward wind is dominated by the advective flow structures. The outer basin gyre enables the flushing of the northeastern and southeastern quarter patches at similar rates. The smaller gyre feature in the northwest enables exchange with the outer basin such that the northwestern quarter dye flushes only three days later than the outer basin dyes. The southwestern quarter dye remains trapped for eight days longer than the northwestern quarter dye by the weak clockwise flow structure in the southwestern inner basin (Figure

Figure 3.7. The effect of westward wind direction on the flow structure is shown with 17-day average velocity vectors produced by W-WIND. (a) and (b) are near-surface and near-bottom vectors, respectively; (c) is the depth-averaged velocity vectors; and (d) the spatial dispersion of the southwestern quarter bottom dye patch four days after release. Shown is the dye patch concentration averaged across the lower half of the water column. Black arrows conceptualize the flow.

3.7 (d)). The fact that bottom patches flush at slightly higher rates than surface patches of the same locations can be attributed to the down-estuary near-bottom flow orientation and the up-estuary near-surface flow orientation.

3.3.4 The influence of diurnal wind on flushing

Dye patch residence times indicate that diurnal wind improves flushing relative to sustained northward wind. Results show that the diurnal wind increases vertical mixing and causes increased advective transport. Despite increased flushing rates the diurnal wind scenarios produce inner basin residence times exceeding seven days, which are considered to be detrimental to water quality. The run that is designated SB2-WIND is compared to N-WIND to show the effect of diurnal wind on dispersion.

Increased vertical turbulent mixing in SB2-WIND can be inferred from the similarity between the residence times of surface and bottom dye patches. The southeastern quarter dye patch can be considered as an example (Figure 3.8). In the SB2-WIND scenario the surface and bottom patches flush at the same rates, but in the N-WIND case the bottom dye lags the surface dye.

Figure 3.8. The difference in flushing rates of the southeastern quarter surface (solid black line) and bottom (dashed grey line) dye patches produced by (a) the N-WIND run and (b) the SB2-WIND run.

Flushing rates are improved in the SB2-WIND scenario compared with the N-WIND case. The depth-averaged velocity vectors of the two cases are almost identical. However, the near-surface flow produced by the SB2-WIND run is oriented northeastward in the inner basin and predominantly eastward in the outer basin. The near-bottom flow is mostly southwestward in the inner basin, and the outer basin near-bottom flow is southward. The fact that the depth varying flow is aligned slightly less perpendicular to the longitudinal axis of the bay allows for increased flushing, despite counter-rotating depth-averaged flow structures.

3.4 Conclusion

Numerical forcing scenario studies show that the wind strongly determines temporal mean circulation and flushing of Greenwich Bay. The wind-driven response dominates the mean flow forced by gravitational and tidal forces. Wind direction causes high flushing variability mainly as a result of differences in advective characteristics.

Flushing characteristics are variable in the horizontal and vertical spatial directions. The bay exhibits two horizontal flushing regimes, one in which the outer basin flushes prior to the inner basin and the other in which the southern bay flushes prior to the northern bay. The regimes are determined by the depth-averaged advection as forced by the wind and tide. Relative flushing rates of the surface and the bottom depends on the orientation of the gravitational- or wind-induced vertical two-layered flow. The diurnal wind increases vertical turbulent mixing, which causes reduced disparities between flushing rates of the surface and bottom water compared to sustained wind. Vertical turbulent mixing is also increased in the presence of a tide relative to a case where the tide is absent.

Greenwich Bay is predisposed to water quality problems by the structure of its tidal- and density-driven flow and by its response to dominant summer wind

directions. In the absence of wind the bay exhibits low flushing as a result of weak flow and a separation of inner and outer bay circulation forced by the tide and horizontal density gradients. Retentive inner and outer basin flow structures are produced by the dominant summer wind due to its orientation relative to the longitudinal axis of the bay. Diurnal summer wind increases flushing relative to sustained northward summer wind, but increased flushing by the diurnal summer wind is not expected to improve water quality due to the high rates of oxygen consumption in the bay.

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APPENDIX

Appendix A: Turbulence closure theory

The theory that provides the foundation for statistical two-equation and empirical models starts with Reynolds averaging of the Navier-Stokes equations [1, 2, 3, 4, 5, 6]. A Reynolds averaging procedure is necessary to obtain turbulent stress terms in the Navier-Stokes equations for incompressible flows. The procedure involves the decomposition of a property into an ensemble mean and fluctuating part, such that, in the case of the x-component of velocity,

$$u = U + u', \quad (\text{A.1})$$

where U is the mean portion of velocity and u' is the fluctuating or turbulent part. Substituting the decomposed velocity into the Navier-Stokes equation lead to the following expression for x-component mean flow

$$\begin{aligned} \frac{DU}{Dt} + \frac{1}{\rho_0} \frac{\partial P}{\partial x} + 2\Omega W \cos\phi - 2\Omega V \sin\phi \\ = \frac{\partial}{\partial x} \left(\nu \frac{\partial U}{\partial x} - \langle u'u' \rangle \right) + \frac{\partial}{\partial y} \left(\nu \frac{\partial U}{\partial y} - \langle u'v' \rangle \right) + \frac{\partial}{\partial z} \left(\nu \frac{\partial U}{\partial z} - \langle u'w' \rangle \right), \end{aligned} \quad (\text{A.2})$$

where ρ_0 is the reference density, P is pressure, Ω is the angular velocity of the Earth, V and W are the ensemble means of the y-component and z-component of velocity, respectively, ϕ is latitude, ν is the kinematic molecular viscosity, v' and w' are the fluctuating parts of the y-component and z-component of velocity, respectively, and $\langle \rangle$ indicates an average.

A scaling exercise for equation A.2 reveals that the dominant force balance in ocean mixed layers and estuarine flows is between the horizontal component of the Coriolis term, $\mathcal{O}(10^{-4})$, and the pressure gradient term, $\mathcal{O}(10^{-4})$. The substantial derivative and the turbulent vertical transport scale as $\mathcal{O}(10^{-5})$. All other terms

are at least three orders of magnitude smaller than the latter two terms and are neglected to simplify the horizontal RANS equations. This includes the so called boundary layer approximation, which neglects all horizontal mixing terms. The same scaling result holds for the y-component of velocity. The vertical momentum equation is simplified to be hydrostatic with the vertical pressure gradient balancing the gravitational acceleration term.

Similar scaling can be done for mean scalar properties to simplify the scalar transport equation,

$$\begin{aligned} & \frac{\partial \Theta}{\partial t} + U \frac{\partial \Theta}{\partial x} + V \frac{\partial \Theta}{\partial y} + W \frac{\partial \Theta}{\partial z} \\ &= \frac{\partial}{\partial x} \left(\nu_\theta \frac{\partial \Theta}{\partial x} - \langle u' \theta' \rangle \right) + \frac{\partial}{\partial y} \left(\nu_\theta \frac{\partial \Theta}{\partial y} - \langle v' \theta' \rangle \right) + \frac{\partial}{\partial z} \left(\nu_\theta \frac{\partial \Theta}{\partial z} - \langle w' \theta' \rangle \right), \quad (\text{A.3}) \end{aligned}$$

where Θ is the ensemble mean of the tracer property, ν_θ is the molecular diffusivity of the tracer, and θ' is the turbulent fluctuating part of the tracer property.

The system of equations, which include the momentum and scalar transport equations, the equation of state, and the continuity equation, is simplified by formerly mentioned scaling assumptions, such that the only turbulent transport terms that remain are $\langle u'w' \rangle$, $\langle v'w' \rangle$, and $\langle w'\theta' \rangle$. Transport equations can be derived for the turbulent fluxes $\langle u'w' \rangle$, as in,

$$\begin{aligned} & \frac{\partial \langle u'w' \rangle}{\partial t} + \left[\frac{\partial}{\partial x} \left(U \langle u'w' \rangle + \langle u'u'w' \rangle - \nu \frac{\partial}{\partial x} \langle u'w' \rangle \right) + \frac{\partial}{\partial y} \left(V \langle u'w' \rangle + \langle v'u'w' \rangle - \right. \right. \\ & \quad \left. \left. \nu \frac{\partial}{\partial y} \langle u'w' \rangle \right) + \frac{\partial}{\partial z} \left(W \langle u'w' \rangle + \langle w'u'w' \rangle - \nu \frac{\partial}{\partial z} \langle u'w' \rangle \right) \right] \left[- \left(\frac{\partial}{\partial x} U \langle u'w' \rangle \right. \right. \\ & \quad \left. \left. + \frac{\partial}{\partial y} U \langle v'w' \rangle + \frac{\partial}{\partial z} U \langle w'w' \rangle \right) - \left(\frac{\partial}{\partial x} W \langle u'u' \rangle + \frac{\partial}{\partial y} W \langle v'u' \rangle + \frac{\partial}{\partial z} W \langle w'u' \rangle \right) \right] \\ & \left[- \left(2\Omega_x \langle u'v' \rangle + 2\Omega_y (\langle w'w' \rangle - \langle u'u' \rangle) + 2\Omega_z \langle -w'v' \rangle \right) \right] \left[- \frac{1}{\rho_0} (g_x \langle w'\rho' \rangle + g_z \langle u'\rho' \rangle) \right] \\ & \left[- \frac{1}{\rho_0} \left\langle u' \frac{\partial p'}{\partial x} + w' \frac{\partial p'}{\partial x} \right\rangle \right] \left[- 2\nu \left\langle \frac{\partial w'}{\partial x} \frac{\partial u'}{\partial x} + \frac{\partial w'}{\partial y} \frac{\partial u'}{\partial y} + \frac{\partial w'}{\partial z} \frac{\partial u'}{\partial z} \right\rangle \right]. \quad (\text{A.4}) \end{aligned}$$

Terms are grouped together by square brackets in equation A.4 to denote (i) shear production (second group), (ii) the redistribution due to rotation (third group), (iii) buoyancy production (fourth group), (iv) the pressure-strain correlator (fifth group), and (v) dissipation of the Reynolds stress (sixth group). Transport equations for $\langle v'w' \rangle$, and $\langle w'\theta' \rangle$ can be obtained in a similar manner.

To solve the transport equations for the second moments it is necessary to derive transport equations for the third moments and pressure-strain correlators that are unknown quantities in equation A.4. Transport equations for the third moments, in turn, contain unknown fourth moments. The procedure of continuously deriving transport equations for unknowns leads to an infinite series of equations, known as the Friedman-Keller series. This unsolvable series of equations is closed by assigning empirical information to the unknowns.

Second-moment closures pertain to the derivation of a closed system of equations from the second-moment equations, such as equation A.4. Simplifying assumptions allow for the computation of only $\langle u'w' \rangle$, $\langle v'w' \rangle$, and $\langle w'\theta' \rangle$, the latter comprising $\langle w'T' \rangle$ for turbulent heat fluxes and $\langle w'S' \rangle$ for turbulent salinity fluxes. The solution of the transport equations of these fluxes requires prognostic equations for the following second moments: $\langle u'u' \rangle$, $\langle u'v' \rangle$, $\langle u'w' \rangle$, $\langle v'v' \rangle$, $\langle v'w' \rangle$, $\langle w'w' \rangle$, $\langle u'T' \rangle$, $\langle v'T' \rangle$, $\langle w'T' \rangle$, $\langle u'S' \rangle$, $\langle v'S' \rangle$, $\langle w'S' \rangle$, $\langle T'T' \rangle$, and $\langle S'S' \rangle$. The dissipation rate of turbulent kinetic energy, ε , is unknown in the transport equation for $\langle T'T' \rangle$. By making a local equilibrium assumption these fourteen second moments can be expressed algebraically and solved as a system of equations, but this system of equations is regarded to be highly complex. Further simplifications are made to reduce the computational expense required to solve these equations. Assumptions include (1) parameterization of pressure-strain correlators, (2) parameterization of dissipation terms, (3) neglect of tracer cross-correlations, (4) boundary layer

approximation, (5) neglect or simplification of transport of second moments, and (6) neglect of rotational terms in the second-moment equations. The consequent system of equations result in simple turbulent flux formulations that resemble molecular principles, namely

$$\langle u'w' \rangle = -c_\mu \frac{k^2}{\varepsilon} \frac{\partial U}{\partial z}, \quad (\text{A.5})$$

$$\langle v'w' \rangle = -c_\mu \frac{k^2}{\varepsilon} \frac{\partial V}{\partial z}, \quad (\text{A.6})$$

$$\langle w'T' \rangle = -c'_\mu \frac{k^2}{\varepsilon} \frac{\partial T}{\partial z}, \quad (\text{A.7})$$

$$\langle w'S' \rangle = -c'_\mu \frac{k^2}{\varepsilon} \frac{\partial S}{\partial z}, \quad (\text{A.8})$$

where c_μ and c'_μ are stability functions. The eddy viscosity (K_M) and the eddy diffusivity (K_H) are defined as

$$K_M = c_\mu \frac{k^2}{\varepsilon}, \quad K_H = c'_\mu \frac{k^2}{\varepsilon}, \quad (\text{A.9})$$

where k is the turbulent kinetic energy, and ε is the turbulent dissipation rate. These variables are unknown in the second-moment closure formulations. However, the variables k and ε can be related through a turbulence macro length scale l according to Taylor scaling [7] such that

$$l = c_l \frac{k^{\frac{3}{2}}}{\varepsilon}, \quad (\text{A.10})$$

where $c_l = (c_\mu^0)^{\frac{3}{4}}$ is a macro length scale for energetic eddies. Statistical turbulence models and empirical turbulence models differ in the strategies employed to solve equations A.5 – A.8.

The nondimensional stability functions, c_μ and c'_μ , of equations A.5 – A.8 depend on two complicated nondimensional parameters. One parameter is a function

of the shear frequency and the other is a function of the Brunt-Väisälä frequency, which represent the external fields of shear and stratification, respectively. Stability functions are derived algebraically from the transport equations for Reynolds stresses after parameterizations of third-order moments and pressure strain correlations. Different formulations for the stability functions have been derived by making different closure assumptions, e.g. by [8] and [9] for non-equilibrium stability functions, and by [10] for quasi-equilibrium stability functions. In ROMS these stability functions are available for selection and are denoted KC [8], CA and CB [9], and G88 [10].

Statistical two-equation turbulence models

Two-equation models, as the name suggests, solve two key equations, namely a transport equation for turbulent kinetic energy (k) and a transport equation for a turbulence length scale [2, 3, 4, 5, 6]. Different variables are used to represent the length scale of turbulence in different two-equation turbulence closure treatments. In the case of the $k - \varepsilon$ two-equation model the length scale is represented by ε [11, 12, 13, 2]; the $k - kl$ and MY25 two-equation models use kl to represent the length scale [14, 4, 14]; and the $k - \omega$ two-equation model uses ω , the turbulence frequency, to represent length scale [15, 16, 17, 4].

Aspects of the Generic Length Scale (GLS) implementation of two-equation models in ROMS is now considered. This information is obtained from [4, 5]. The GLS method is based on the principle that the transport equation for turbulence length scale can be for any quantity $k^m \varepsilon^n$ with $n \neq 0$, since equation A.10 is a nonlinear relation between k , ε , and l . Kinetic energy per unit mass of the velocity fluctuations is defined as $k = 0.5\langle u'u' + v'v' + w'w' \rangle$. The transport equation for turbulent kinetic energy is then obtained from equations for $\langle u'u' \rangle$, $\langle v'v' \rangle$, and

$\langle w'w' \rangle$ in a similar way to equation A.4, as

$$\frac{\partial k}{\partial t} + U_i \frac{\partial k}{\partial x_i} = \frac{\partial}{\partial z} \left(\frac{K_M}{\sigma_k} \frac{\partial k}{\partial z} \right) + P + B - \varepsilon, \quad (\text{A.11})$$

where k is the turbulent kinetic energy, U is the mean component of velocity, K_M is the eddy viscosity coefficient, σ_k is the turbulent Schmidt number for the eddy diffusivity of turbulent kinetic energy, P and B represent production of turbulent kinetic energy by shear and buoyancy, respectively, and ε is the dissipation of turbulent kinetic energy. Turbulent kinetic energy production by shear is expressed as

$$P = -\langle u'w' \rangle \frac{\partial U}{\partial z} - \langle v'w' \rangle \frac{\partial V}{\partial z} = K_M \left(\left(\frac{\partial U}{\partial z} \right)^2 + \left(\frac{\partial V}{\partial z} \right)^2 \right), \quad (\text{A.12})$$

and by buoyancy as

$$B = -\frac{g}{\rho_0} \langle w'\rho' \rangle = -K_H \left(\frac{g}{\rho_0} \frac{\partial \rho}{\partial z} \right). \quad (\text{A.13})$$

The second transport equation describes a generic parameter ψ (modification for the k-kl scheme not shown here) as a generic function for the turbulent length scale as

$$\frac{\partial \psi}{\partial t} + U_i \frac{\partial \psi}{\partial x_i} = \frac{\partial}{\partial z} \left(\frac{K_M}{\sigma_\psi} \frac{\partial \psi}{\partial z} \right) + \frac{\psi}{k} (c_1 P + c_3 B - c_2 \varepsilon), \quad (\text{A.14})$$

where σ_ψ is the turbulent Schmidt number for the eddy diffusivity of ψ and c_1 , c_2 , and c_3 are empirical parameters. Dissipation is calculated using

$$\varepsilon = (c_\mu^0)^{3+\frac{p}{n}} k^{\frac{3}{2}+\frac{m}{n}} \psi^{\frac{-1}{n}}, \quad (\text{A.15})$$

where c_μ^0 is the stability coefficient, which is determined from experimental data for unstratified channel flow with a log-layer solution. Values for the stability coefficient range between 0.5270 and 0.5544 [5]. The generic parameter can be expressed as

$$\psi = (c_\mu^0)^p k^m l^n, \quad (\text{A.16})$$

where l is the turbulent length scale. The purpose of p , m , and n is to define the appropriate generic parameter, ψ , for equation A.14, depending on the specific turbulence closure scheme that is selected.

The empirical parameters that arise in equations A.11 and A.14 have to be determined by theoretical assumptions, observations, or by experiments done numerically or in the laboratory [3]. Early investigators estimated the empirical parameter c_2 from observations of freely decaying homogeneous turbulence by relating it to turbulence decay rates [3]. Parameter c_1 was determined by wind tunnel experiments with homogeneously sheared grid turbulence [3]. The parameter c_3 has not been directly determined in the laboratory, but it depends on the level of stability of the water column [3]. Under stable stratification c_3 is calculated with relation to a steady-state Richardson number, which is calibrated from idealized wind mixing experiments.

In the standard $k - \varepsilon$ approach [11, 12, 13, 2], the parameters p , m , and n are defined such that equation A.15 gives $\varepsilon = \psi$, such that the generic parameter is defined for ε and the second equation becomes

$$\frac{\partial \varepsilon}{\partial t} + U_i \frac{\partial \varepsilon}{\partial x_i} = \frac{\partial}{\partial z} \left(\frac{K_M}{\sigma_\varepsilon} \frac{\partial \varepsilon}{\partial z} \right) + \frac{\varepsilon}{k} (c_{1\varepsilon} P + c_{3\varepsilon} B - c_{2\varepsilon} \varepsilon), \quad (\text{A.17})$$

where σ_ε is now the turbulent Schmidt number for the eddy diffusivity of dissipation. The right hand side of the transport equation for ε thus varies as scaled linear combinations of the right hand side of the transport equation for k .

The $k - \omega$ model [15, 16, 17, 4] implementation defines ω as

$$\omega = \frac{\varepsilon}{(c_\mu^0)^4 k} \quad (\text{A.18})$$

and a second transport equation as

$$\frac{\partial \omega}{\partial t} + U_i \frac{\partial \omega}{\partial x_i} = \frac{\partial}{\partial z} \left(\frac{K_M}{\sigma_\omega} \frac{\partial \omega}{\partial z} \right) + \frac{\omega}{k} (c_{1\omega} P + c_{3\omega} B - c_{2\omega} \varepsilon), \quad (\text{A.19})$$

where σ_ω is the turbulent Schmidt number for the eddy diffusivity of ω .

The $k - kl$ model [14, 4] is a modification of an original MY25 model. In ROMS the $k - kl$ model is implemented through the GLS method, and the MY25 model is represented in its original form with a G88 stability function. The main differences between the two models are (1) the vertical diffusion coefficient for transport of k and kl , (2) the wall proximity function, and (3) the specification of the buoyancy parameter. The first transport equation characterizing k used in the $k - kl$ and MY25 models is

$$\frac{\partial k}{\partial t} + U_i \frac{\partial k}{\partial x_i} = \frac{\partial}{\partial z} \left(K_Q \frac{\partial k}{\partial z} \right) + P + B - \varepsilon. \quad (\text{A.20})$$

In the MY25 implementation, $K_Q = \sqrt{2kl}S_Q$ and S_Q is a constant stability function that causes higher entrainment compared to the standard model for turbulent kinetic energy transport (equation A.11). In the $k - kl$ implementation, $K_Q = K_M/\sigma_q$ where σ_q is the turbulent Schmidt number for k . The second transport equation characterizing length scale is

$$\frac{\partial(kl)}{\partial t} + U_i \frac{\partial(kl)}{\partial x_i} = \frac{\partial}{\partial z} \left(K_Q \frac{\partial(kl)}{\partial z} \right) + l(c_1P + c_3B + c_2\varepsilon F_{wall}), \quad (\text{A.21})$$

where F_{wall} is a wall proximity function needed to ensure a positive value for the diffusion coefficient and thus correctly reproduce near-wall flow. The parabolic wall proximity function specified for the MY25 model in ROMS is defined as

$$F_{wall} = \left(1 + E_2 \left(\frac{l}{\kappa} \frac{d_b + d_s}{d_b d_s} \right)^2 \right), \quad (\text{A.22})$$

where $E_2 = 1.33$, d_s and d_b are distances from a grid point to the surface and the bottom, respectively. A selection of wall proximity functions are available in ROMS for the $k - kl$ model. Available functions in ROMS produce parabolic (equation A.22), triangular, or linear near-wall shapes.

A qualitative analysis that indicated the sensitivity of the empirical parameters in two-equation models were conducted using a test application that is avail-

able in ROMS. The ROMS estuary test case [5] was run with $k - \varepsilon$ turbulence closure scheme and a KC stability function. The grid represents an estuary that is 100 km in length and 1 km in width. The estuary is characterized by linearly decreasing depth of 10 m at the mouth in the west to 5 m at the estuary head in the east. The grid has 200 x 3 horizontal points and twenty vertical sigma levels. Initially the estuary is well-mixed with a salinity of 35 in the west and 0 in the east. River input in the east is constant at 400 m³/s. In the west the flow is forced by a sinusoidal wave with a period of twelve hours for a total of ten days. The northern and southern boundaries of the grid are frictionless walls. Temperature is constant at 10 °C throughout the estuary.

The sensitivity of a number of key parameters was investigated by running a series of estuary test cases. One parameter at a time was varied, while holding all others constant. We tested values for σ_k , σ_ψ , $c1$, $c2$, and $c3^-$ that were factors of two and 0.5 relative to the standard magnitudes of the $k - \varepsilon$ turbulence closure scheme with the KC stability function.

The differences between these scenarios were qualitatively assessed by looking at the lateral and temporal averages of water properties for the final five days of the runs (Figure A.1). The variables considered were axial salinity, velocity, and vertical turbulent eddy diffusivity. The reference case showed a vertically well-mixed salt intrusion extending halfway up the estuary at the bottom of the water column separated from a shallow surface layer of fresh water by a strong vertical halocline centered around 3 m. The vertical eddy diffusivity ranged from approximately 10⁻⁵ m²/s in the surface freshwater layer to 10⁻² m²/s within the salt intrusion layer. Flow magnitude was directed toward the estuary head in the region of the salt intrusion and toward the mouth elsewhere.

Solutions were insensitive to most empirical parameters, which showed a lack

Figure A.1. Effects of doubling and halving values of turbulence closure parameters using the $k - \varepsilon$ scheme with KC stability as reference for the following parameter selections: (a) Reference, (b) $0.5\sigma_{kRef}$, (c) $2\sigma_{\psi Ref}$, (d) $2c_{3Ref}^-$, (e) $2c_{1Ref}$, (f) $0.5c_{1Ref}$. Variables evaluated are average axial salinity (top), velocity (middle), and vertical eddy diffusivity (bottom).

in variations in the salinity, velocity, and mixing structures (Figure A.1). Both high and low values of the turbulent Schmidt number for the eddy diffusivity of k (σ_k) produced results that resembled the reference case. Applying $0.5^*c3_{Ref}^-$ resulted in unchanged average water properties compared to the reference case. However, when $2^*c3_{Ref}^-$ was specified the result was a vertically well-mixed water column through the entire length of the estuary with saline water only intruding a relatively small distance up the estuary. The eastward flow is weak compared to that of the reference case. A doubling of the turbulent Schmidt number for the eddy diffusivity of ε , namely σ_ψ (or σ_ε), caused only slightly higher mixing than was seen in the reference case. Applying $0.5^*\sigma_{\psi Ref}$ produced lower vertical mixing compared to the reference case.

The $c1$ and $c2$ parameters affect the production of turbulent kinetic energy dissipation and the dissipation of turbulent kinetic energy dissipation, respectively. For high values of $c1$ the production of ε is reduced, and for low values of $c1$ the production of ε is increased. Similarly, for low values of $c2$ the dissipation of ε is reduced and for high values of $c2$ the dissipation of ε is increased. By applying 0.5^*c2_{Ref} the dissipation of ε was six times lower than in the reference case and by specifying 2^*c1_{Ref} the production of ε was six times higher than in the reference case. The effect of a lower (/higher) dissipation (/production) of ε is to cause higher (/lower) overall dissipation. It was found that the results of 0.5^*c2_{ref} was similar to those of 2^*c1_{Ref} and the result of 2^*c2_{Ref} was similar to 0.5^*c1_{Ref} . The latter scenario provided relatively high tidal mixing below the halocline and towards the estuary head (Figure A.1). This high mixing caused the salt intrusion to be limited to near the estuary mouth, which also resembled properties that were seen when $2^*c3_{Ref}^-$ was specified. In contrast, a doubling of $c1_{Ref}$ or halving of $c2_{Ref}$ produced relatively low mixing through the length and depth of the estuary (Figure A.1).

The latter specifications caused the salt intrusion to propagate to the head of the estuary and a relatively strong estuarine exchange flow was produced along the entire length of the estuary.

Empirical turbulence models

The K-profile parameterization (LMD) is an empirical turbulence model available in ROMS [1, 6]. The model employs parameterizations to separate the surface boundary layer and the ocean interior, and to specify vertical turbulent fluxes in each region. The boundary layer depth is calculated as the minimum of three depth levels, namely the Ekman depth, the Monin-Obukhov depth, and the shallowest depth at which the bulk Richardson number exceeds a critical value. The Ekman depth is dependent on the friction velocity and the Coriolis parameter. The Monin-Obukhov depth is related to the friction velocity, the von Kármán constant ($\kappa = 0.40$), and the surface buoyancy flux. The critical bulk Richardson number is typically in the range 0.25–0.5. The bulk Richardson number is calculated in terms of differences in density and horizontal velocity between the given depth and the surface layer. The vertical turbulent flux $\langle c'w' \rangle$ of a fluctuating variable c' can be expressed as

$$\langle c'w' \rangle = \langle c'w'_L \rangle + \langle c'w'_N \rangle, \quad (\text{A.23})$$

where $\langle c'w'_L \rangle$ indicates the local turbulent flux and $\langle c'w'_N \rangle$ represents the nonlocal turbulent flux. The local turbulent flux component is given by the familiar gradient parameterization

$$\langle c'w'_L \rangle = -K_c \frac{\partial C}{\partial z}, \quad (\text{A.24})$$

where $C(z, t)$ is the ensemble mean of a property and $K_c(z, t)$ is its turbulent diffusivity.

Within the surface boundary layer the local vertical diffusivity (K_c) is parameterized as the product of the boundary layer depth (h), a depth dependent

turbulent velocity scale (w'_c), and a cubic vertical shape function (G) defined as

$$G(\sigma) = a_0 + a_1\sigma + a_2\sigma^2 + a_3\sigma^3 \quad (\text{A.25})$$

such that

$$K_c = hw'_c(\sigma)G(\sigma), \quad (\text{A.26})$$

where $\sigma = z/h$ defines a nondimensional vertical location within the boundary layer. The polynomial coefficients of G are chosen such that (a) the interior viscosity at the bottom of the boundary layer is matched and (b) the Monin-Obukhov similarity theory [18] holds near the surface. Surface layer similarity theory is used to estimate the local turbulent velocity scale in the boundary layer. The turbulent velocity scale for a property c is defined as

$$w'_c = \frac{\kappa u^*}{\phi_c(\zeta)}, \quad (\text{A.27})$$

where κ is the von Kármán constant ($\kappa = 0.40$), u^* is the friction velocity, ϕ_c is a nondimensional flux profile that varies according to the stability of the boundary layer forcing, and $\zeta = z/L$ is the surface layer stability parameter with L the Monin-Obukhov length scale. Friction velocity is defined as $u^{*2} = |\tau_0|/\rho_0$ where τ_0 is surface stress and ρ_0 is surface density. The stability parameter (ζ) is assumed to vary over the entire depth of the boundary layer in stable and neutral conditions. In unstable conditions it is defined to vary only in the surface layer. The flux profile (ϕ_c) is defined via analytic fitting to atmospheric surface boundary layer data.

Below the surface boundary layer vertical mixing is assumed to be entirely local and result from the superposition of three processes, namely shear instability, internal wave breaking, and double diffusion. Each process is assigned a separate local vertical diffusivity parameter,

$$K_c = K_c^s + K_c^w + K_c^d. \quad (\text{A.28})$$

The shear mixing term, K_c^s is derived from a gradient Richardson formulation. It is assumed that the viscosities and diffusivities induced through shear instability are the same and decrease strongly with the gradient Richardson number. The internal wave breaking term is constant and defined as, $K_c^w = 1 \times 10^{-4} \text{ m}^2/\text{s}$ for momentum and $K_c^w = 1 \times 10^{-5} \text{ m}^2/\text{s}$ for scalar tracers based on findings from studies by [19, 20]. Mixing due to internal waves serves as the background mixing in the LMD turbulence closure scheme. Double diffusive mixing can occur when the vertical gradient of density is stable, but the vertical gradient of either salinity or temperature is unstable. In the presence of a stable vertical density gradient, salt fingering occurs with an unstable vertical salinity gradient, and diffusive convection occurs with an unstable vertical temperature gradient [21]. The model distinguishes between salt fingering and diffusive convection. The distinction is done based on the double diffusion density ratio, R_ρ ,

$$R_\rho = \left(\alpha \frac{\partial T}{\partial z} \right) / \left(\beta \frac{\partial S}{\partial z} \right), \quad (\text{A.29})$$

where T is potential temperature, S is salinity, and α and β are the thermodynamic expansion coefficients for temperature and salinity, respectively. Salt fingering ($R_\rho > 1.0$) is parameterized based on laboratory and field data. The mixing of momentum from diffusive convection ($0 < R_\rho < 1.0$) is modeled according to diffusive layer thickness.

The nonlocal component, $\langle c'w'_N \rangle$, is the surface boundary turbulent flux due to eddies that are comparable in size to the scale on which the background gradient $\partial C/\partial z$ varies. Nonlocal momentum flux is assumed to be zero and nonlocal scalar flux is nonzero only in unstable forcing conditions. In order to parameterize the nonlocal flux for any scalar, it is necessary to define an effective gradient of c , $-\gamma_c(z, t)$, such that

$$\langle c'w'_N \rangle = K_c \gamma_c. \quad (\text{A.30})$$

The effective gradient is defined such that equation A.30 depends only on surface fluxes and the surface layer thickness. The formulation for nonlocal scalar flux is based on a parameterization for pure free convection, i.e. convection that occurs in the absence of a mean shear, which is modified to include unstable surface forcing conditions such that

$$\gamma_S = C_s \frac{\langle w' S'_0 \rangle}{w'_s(\sigma)h}, \quad \gamma_T = C_s \frac{\langle w' T'_0 \rangle + \langle w' T'_R \rangle}{w'_s(\sigma)h}, \quad (\text{A.31})$$

where C_s is a proportionality constant that depends on the extent of the surface layer and the von Kármán constant ($\kappa = 0.40$), $\langle w' S'_0 \rangle$ and $\langle w' T'_0 \rangle$ are the surface turbulent fluxes of salinity and potential temperature, respectively, $\langle w' T'_R \rangle$ is the radiative turbulent flux contribution – it represents the amount of radiative heat absorbed in the boundary layer that effectively contributes to the nonlocal transport of heat, and w'_s is the turbulent velocity scale for scalars.

Characteristics of turbulence closure methods

There are a number of benefits and deficiencies of the different closure models. A drawback of the empirical approach is that its accuracy more strongly depends on the observations used to model fluxes than in the case of statistical models. This is due to the fact that the empirical approach relies almost entirely on observational evidence to justify flux parameterizations [6]. A benefit of the empirical approach is that it captures nonlocal fluxes [6]. These nonlocal fluxes are not included via the truncated Friedman-Keller series of the statistical approach. A disadvantage of the statistical two-equation approach, therefore, is that flux calculations are based on instantaneous local flow properties only [5, 6].

Both the statistical and empirical models neglect horizontal mixing processes as a result of the boundary layer assumption [6]. The assumption is considered realistic when the numerical grid has sufficiently high resolution to resolve horizontal eddy activity, although the subgrid scales remain unresolved. Separate subgrid

scale parameterizations for horizontal mixing are introduced in numerical models such as ROMS to account for these unresolved processes. In ROMS the mixing by Langmuir circulation is not represented by statistical and empirical models as a result of the hydrostatic assumption and the neglect of surface wave effects [6].

The distinguishing factor for the different statistical two-equation models is the treatment of the length scale variable [5, 3, 6]. Assumptions in the derivations of statistical models lead to several unresolved processes. Two-equation models exclude counter-gradient fluxes due to omission of the transport term in the heat flux equation [6]. The original MY25 and $k - \varepsilon$ models are known to incorrectly reproduce the vertical structure of the flow in the wave-enhanced layer below the free surface [6]. In the GLS treatment of ROMS the dissipation due to wave breaking is parameterized through a modification of the turbulent Schmidt number, which can be manually specified [5]. This specification was not made in the current application. The effects of internal waves on turbulence is not represented by statistical models due to the hydrostatic assumption and grid resolution [6].

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