THE HYDRODYNAMICS OF THE UPPER NEUSE RIVER ESTUARY, NC
AND THEIR INFLUENCE ON DISSOLVED OXYGEN DISTRIBUTION

by
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ABSTRACT

JANELLE V. REYNOLDS-FLEMING: The Hydrodynamics of the Upper Neuse River Estuary, NC and their Influence on Dissolved Oxygen Distribution
(Under the direction of Richard A. Lueettich, Jr.)

Four years of hydrographic data along the main channel of the Neuse River Estuary (NRE) were used to determine mean distributions of salinity, temperature and dissolved oxygen (DO). These distributions varied seasonally and direct relationships between salinity, salinity stratification and bottom DO concentrations were determined. A seasonal DO budget confirms that vertical mixing and biological demand primarily control DO concentrations.

On shorter timescales, lateral variability was noted through the deployment of two bottom-mounted Acoustic Doppler Current Profilers (ADCPs), two bottom-mounted Conductivity-Temperature and Depth (CTD) sensors, and two autonomous vertically profiling CTD/DO systems. The profiling systems were novel instrumentation developed specifically for this application to achieve high vertical and temporal resolution of salinity and DO data. Additionally, several data collection periods using a boom-mounted ADCP and a winch-driven CTD were used to increase the spatial coverage of the upper NRE. Data analysis techniques such as spectral and wavelet analysis were used to identify several physical phenomena, including barotropic and baroclinic seiches, lateral upwelling of high salinity/low DO water and wind-driven, low frequency lateral salinity variability.

The three-dimensional finite difference model, Environmental Fluid Dynamics Code (EFDC), was calibrated for the NRE and was used to simulate flow and transport conditions during the summers of 1999 and 2000. Model data were independently validated with observed data and were statistically shown to adequately reproduce conditions in the NRE. Experimentation with boundary conditions documented the sensitive nature of the upper NRE to both freshwater discharge and wind.

A linear, two layer theoretical model was applied to the upper NRE to quantify the vertical variability in the pycnocline caused by across channel wind. Wind speeds greater than 10 m s\(^{-1}\) can force the pycnocline to surface of the windward shore. This motion also causes the low DO water below the pycnocline into the upper regions of the water column.

Previous studies of the NRE had noted lateral variability, but had never tried to quantify this variability or to determine the cause. The development of novel profiling instruments provided data that linked physical phenomena to biological events and provided insight into the mechanisms behind a fish kill.
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Chapter 1

Introduction and Background

1.1 The Study Site

The Neuse River Estuary (NRE) in eastern North Carolina is a shallow, intermittently mixed estuary originating near New Bern, NC and extending eastward into the Albemarle-Pamlico Sound. The average depth is 3.6m, the average width is 6.5km and its length stretches 70km from the head to the mouth (figure 1.1). Km distance downstream from the head of the estuary to the mouth is presented in figure 1.2. It is classified as a mesohaline estuary and access to the Atlantic Ocean is restricted to three small inlets that punctuate the Outer Banks barrier islands. The small cross sectional area of the inlets in comparison to the volume of both the NRE and Pamlico Sound reduces tidal influence on NRE circulation. As tidal influence is minimal, the dynamics of the NRE are mainly controlled by freshwater discharge and atmospheric forcing.

1.2 The Problem

The NRE received public attention in the mid 1990's because of the notorious fish kills that occurred in its waters (Leavenworth, 1995b; Leavenworth, 1995a; Press, 1995). Longitudinal water quality data from the MODMON (Modeling and Monitoring of the Neuse River Estuary) documented the presence of hypoxic (DO<4mg l$^{-1}$ ) conditions during most of these fish kills (figure 1.3). While these large expanses of hypoxic bottom water had been cited as a cause for fish kills (Pael et al., 1998), it was unclear how bottom water DO could affect surface dwelling pelagic fishes like Atlantic Menhaden, Brevoortia tryannus.

Since the majority of the fish kill events occurred in the upper NRE (from New Bern towards the bend in the estuary, figure 1.1), the lateral variability of low DO water and salinity as well as the response of these variables to external stresses like atmospheric forcing and freshwater discharge was studied. The following research questions were posed: 1) what was the relationship between salinity, salinity stratification and hypoxic water (DO < 4 mg l$^{-1}$ ); 2) what was the effect of local atmospheric forcing and freshwater discharge on lateral variability in circulation, salinity, and hypoxia; 3) was there a relationship between wind events and fish kill occurrence due to hypoxia; and 4) could another conserved scalar be used as a proxy for oxygen to simplify model construction?
1.3 Physical Stratification and Low DO Studies

A direct relationship between the occurrence of stratification and the depletion of oxygen below the pycnocline was found throughout the literature for many estuaries. Stratification separated the water column into at least two vertical sections where the bottom layer was effectively cut off from re-aeration due to lack of mixing processes. This separation allowed benthic respiration to deplete DO in the lower layers. Increased benthic respiration and hence DO depletion were directly attributed to eutrophication.

One effect of eutrophication is the supply of excess phytoplankton biomass to the benthos. This supply of organic material eventually will be decomposed, absorbing DO in the process. The resultant complete lack of oxygen in the water column was defined as anoxia. Hypoxia was defined as the reduction of DO concentration to less than 4 mg l\(^{-1}\) for a period lasting 24 hrs or more (Thursby et al., 1999). The term severe hypoxia represented DO values less than 2 mg l\(^{-1}\).

Hypoxia in the main channel of the Chesapeake Bay was studied extensively and it was demonstrated that the main drivers for anoxia were benthic respiration and water column stratification during the summer (Taft et al., 1980; Officer et al., 1984). Oxygen depletion was also linked to nutrient loading (i.e., eutrophication) through the accumulation, deposition and decomposition of phytoplankton biomass (Smith et al., 1992). Hypoxia was also observed in the deeper waters of the Rappahannock and York rivers, but rarely in the James River. Though tributaries of the Chesapeake Bay, these estuaries were influenced by anoxic conditions in the Chesapeake via exchange at the mouth of the tributary with the Bay. Differences between the tributaries in terms of occurrence, extent, and frequency of hypoxia
Figure 1.3: Bottom dissolved oxygen (DO) contours and fish kill locations (blue diamonds) from 1994 through 2001.
were attributed not only to vertical stratification differences, but differences in the longitudinal salinity gradient as well (Kuo and Neilson, 1987; Kuo et al., 1991; Park et al., 1996).

Northward, vertical stratification played a major role in the development of hypoxia in the western Long Island Sound. There, the oxygencline always coincided with the pycnocline and periods of oxygen depletion corresponded with periods of thermally-controlled stratification (Welsh and Eller, 1991). The oxygencline also coincided with the pycnocline in Waquoit Bay, Massachusetts to exacerbate the formation of an anoxic region (D'Avanzo and Kremer, 1994). In that system, however, it was the combination of eutrophication, stratification and solar irradiance that created anoxic water volumes.

In a more temperate region, a 15 year study on salinity and bottom-water hypoxia was conducted in the shallow, partially mixed Pamlico River Estuary (Stanley and Nixon, 1992). Investigators concluded that hypoxia developed only when there was both vertical water-column stratification and warm water temperature. They assumed this relationship was a natural feature of the system and not attributable to eutrophication. Furthermore, in Mobile Bay, Alabama and the adjacent shelf bottom waters, oxygen depletion was also demonstrated to be directly related to the intensity of water column stratification (Turner et al., 1987). Finally, in an analysis of mid-river hydrographic data from the NRE, Buzzelli, et al (2002) showed that during warmer months, hypoxic conditions were present even in weakly stratified (Δ psu=1-2) conditions. It was therefore concluded that dissolved oxygen depletion was dependent on the development of stratification in the NRE during the summer and early fall (Buzzelli et al., 2002).

Low dissolved oxygen concentrations were detrimental to organisms living within these estuaries. Benthic, sessile organisms were most often affected as most lacked the necessary locomotion to avoid the hypoxia. In the Chesapeake Bay, macrobenthic communities affected by hypoxia were characterized by lower species diversity, lower biomass, a lower proportion of deep-dwelling biomass (deeper than 5 cm in the sediment), and changes in community composition (Dauer et al., 1992). The macrobenthos of the Pamlico Estuary had low species diversity and density in the summer as correlated with anoxic/hypoxic conditions (Tenore, 1972). In the NRE, changes of infaunal abundance and community structure were recorded as a result of hypoxic events (Luetich et al., 2000a).

Pelagic and mobile organisms were somewhat more fortunate than benthic organisms as they had the ability to navigate around hypoxia/anoxia. However, this avoidance measure sometimes proved to be futile as hypoxic/anoxic events decimated the benthic community (Diaz and Rosenberg, 1995). It was found that demersal and pelagic fishes dependent on the benthic community as a food source could eventually starve if the anoxic event was of a long enough duration and succession was not rapid enough (Peterson et al., 2000). Future generations could suffer as low DO conditions affected the growth rates and behavior of larval stages of organisms (Breitburg, 1992; Baker and Mann, 1992; Das and Stickle, 1993; Breitburg, 1994; Tankersley and Wieber, 2000) as well as affect the trophic interactions (Breitburg et al., 1994; Breitburg et al., 1997). Finally, if the anoxic volume was large enough, these organisms
might not be able to avoid the volume or they may become trapped in the hypoxic/anoxic volume due to the currents within the basin. In fact, this was often the case in Mobile Bay "Jubilees" as crab, flounder, and shrimp migrated to the shores to escape the leading edge of an anoxic zone (Schroeder and Wiseman, 1988).

1.4 Relationship between Stratification and Anoxia

The degree and occurrence of stratification was determined by the physical circulation of the system. Physical circulation in estuaries was a unique result of the complex interaction between forcing mechanisms like fresh water discharge, tidal influence and meteorological forcing (Pritchard, 1952; Pritchard, 1954; Pritchard, 1956; Schroeder et al., 1990; Simpson et al., 1990; Jay and Smith, 1990; Nepf and Geyer, 1996; Valle-Levinson et al., 1998; Geyer et al., 2000). As noted previously, hypoxic/anoxic conditions were intimately connected with stratification in eutrophic estuaries during the warmer months. Hence, information about hypoxia/anoxia development could be extracted from knowledge of the mechanisms controlling the development and maintenance of stratification. Indeed, a probabilistic model for predicting hypoxia based on various factors, including stratification, was developed for the NRE (Borsuk et al., 2001).

1.4.1 Physical Studies of the NRE

Initial studies of the physics of the NRE began in the late sixties. At that time, a study consisting of two dye releases was conducted (Woods, 1969). Through these dye releases, average downstream rates were calculated as 5.4 cm s⁻¹ during a six day release and 3.1 cm s⁻¹ during an 18 day release. No effort was made to describe the mechanisms behind the observed circulation, however these numbers were used to estimate flushing rates.

As technology improved, a more intensive study involving the simultaneous deployment of multiple current meters was conducted (Knowles, 1975). Seven near bottom and two near surface current meters were deployed in the NRE for 38 days beginning in August 1973. Analysis of the current records found significant spectral energy near the semi-diurnal period. At the time, the energy was attributed to the lunar semi-diurnal tide and the strongest evidence for wind forcing was due to the diurnal sea breeze.

Investigation of water level time series collected near Minnesott Beach during a turbidity maxima study suggested a tidal range of less than 3.1 cm (Kim, 1990). Power spectral analysis conducted on the same record noted a significant peak in the approximate semi-diurnal frequency band. This peak was attributed to the M₂ tidal period by default as comparison with power spectral density graphs of wind speed for two directions, NE and SE, did not contain a similar peak. Additionally, coherence tests between water level and wind speed in the NE direction showed high coherence for all periods longer than 1.5 days with a zero phase lag. These results validated earlier speculation of the effect of wind on
water level variations for longer periods (Roeloffs and Bumpus, 1953; Giese et al., 1979; Pietrafesa et al.,
1986), but not for the sea-breeze effect noted in a previous study (Knowles, 1975).

A roughly semi-diurnal response in water level from New Bern to Minneutott Beach was noted in a later study (Robbins and Bales, 1995). These investigators did not attribute this response to the M2 tidal period, but instead concluded that it was a combination of the sea-breeze effect and a Pamlico Sound seiche. Velocity data was also recorded as part of that study. Ten current meters were placed 1.5 m from the bottom along the estuary for a period of 17 days. Investigators noted lateral differences in both speed and direction. In general, slower, upstream flow predominated near the north shore of the estuary whereas faster, downstream currents predominated near the south shore. Further documentation of this pattern was provided by bottom-mounted S4s used during a study of the lateral variation of the halocline in the upper NRE (Sweet, 2000) and noted in the application of a two-dimensional, vertically-integrated model of the NRE (Robbins and Bales, 1994). These differences may have been due in part to the effect of the coriolis force (Giese et al., 1979).

A two-dimensional, vertically integrated, unsteady flow and transport model was used to identify the driving mechanisms behind circulation in the NRE (Robbins and Bales, 1995). The model was calibrated with field data and then used to simulate the circulation in the NRE for 4 periods: June 1-24, 1991; October 24-November 3, 1989; May 1-30, 1991; and September 1-30, 1991. Results showed that velocity was fastest through the narrow portions of the NRE due to the morphologic constriction and that recirculating eddies were generated near protrusions of land into the estuary. One of the main results of that model study was that directional differences in wind influenced the mean transport through the estuary. For example, strong winds from the N-NE increased transport in the upstream direction by over 50%.

Significant energy at the semi-diurnal frequency was noted in both water velocity and water level records recorded at Minneutott Beach (Luetttich et al., 2000a). Coherence tests between water level and wind speeds in the NE direction also verified earlier results (Kim, 1990). In particular, wind blowing toward the NE drove water out of the NRE and into Pamlico Sound whereas wind blowing toward the SW drove water from Pamlico Sound into the NRE (Luetttich et al., 2000b). A relationship between water velocity and water level was also noted by coherence tests. It was found that the pervasive peak in energy at the semi-diurnal frequency band was due to a wind-driven barotropic wave that traveled between the lower NRE and the Pamlico Sound. That finding was later supported using wavelet analysis and a two-dimensional, vertically integrated, finite element model (Luetttich et al., 2002). At longer periods, water level in the NRE responded to changes in freshwater discharge in the NRE (Luetttich et al., 2000a).

A synthesis of these observations characterizes the NRE as a wind and freshwater discharge-driven estuary having low flows that are spatially and temporally complex. The Outer Banks, which ring the Albemarle-Pamlico Sound system, allow limited water exchange with the Atlantic Ocean through very narrow inlets. The influence of the astronomical tide on the NRE is small. At relatively long time
scales (i.e. months) and during calm periods, the primary driver is fresh water discharge. However, at
shorter time scales (i.e. hours, days and weeks), circulation in the NRE is dominated by meteorological
forcing, either remotely due to the setup of Pamlico Sound or directly as wind stress over the NRE itself.
These driving forces determined the location of the salt-wedge, the degree of longitudinal stratification,
and ultimately the development and location of hypoxic/anoxic regions in the NRE.

Previous investigators noted lateral differences in velocity, but did not attribute it to a physical
mechanism (Robbins and Bales, 1995). Indeed, the role of freshwater discharge was completely over-
looked by previous investigators. Additionally, the instrumentation from previous investigators lacked
the temporal and spatial resolution to analyze higher frequency variability in flow and transport due to
higher frequency wind events. Finally, no attempt was made to quantify lateral variability in salinity or
DO or its possible role in fish kills.

1.4.2 Physical Studies of Wind Driven Circulation

Wind-induced circulation has been investigated in both laboratory and natural settings. Lab-

oratory studies allowed the detailed examination of the vertical current profile in a homogeneous fluid,

usually water. In general, a surface current was generated by the tangential shear stress applied by the
wind at the air-sea interface. Normally, the surface current flowed with the wind and water velocity in

the vertical decreased towards zero. Below that a return flow in the opposite direction of the surface
flow was established in the lower layers. That general pattern would be altered, however by the Coriolis
affect, water depth, and interaction with boundaries. Velocity in the lower layers eventually returned
to zero in the bottom boundary layer (Baines and Knapp, 1965; Wu, 1975; Tsuruya et al., 1985). This
type of current, at least in the surface, was documented in unstratified lakes (Bye, 1965) and generated
water level setup at the downwind side of a basin, which in turn generated a series of seiches (Dorn,
1953).

Mathematical quantification of the wind effects on a thermally stratified lake system of uni-
form depth, neglecting the effects of rotation, was conducted and applied to the Windermere Lake
System (Heaps and Ramsbottom, 1966). The investigators indicated that changes in the wind-stress
component parallel to the length of the lake set up longitudinal internal seiches centered on the ther-
omcline and longitudinal surface seiches. In a similar analysis, it was shown that wind stress applied at
the surface of a basin of variable depth sets up a circulation pattern characterized by relatively strong
barotropic coastal currents in the direction of the wind, with return flow occurring over the deeper
regions (Csanady, 1973b).

For larger lake systems, such as the Great Lakes of North America, it was necessary to incor-
porate the effect of earth’s rotation on any model study of forced circulation. Observational evidence
in Lakes Michigan, Huron and Ontario in the summertime showed the sudden displacement of warm
epilimnion water by cold hypolimnion water near the beaches (Csanady, 1968b). This type of upwelling
was always associated with an offshore wind that generated transverse internal seiches (Csanady, 1973a). However, it also occurred when wind direction was parallel to the shore. These shore parallel winds raised the thermocline near the shore and occasionally caused the thermocline to intersect the surface. It was also noted that a current regime flowing with the parallel wind was associated with a raised thermocline near the shore. By applying simplified equations of motion, the investigator showed that the current regime and subsequent thermocline tilt associated with parallel winds was due to an “Ekman” type drift under a steady wind stress (Csanady, 1963b). This model study also showed the existence of standing waves on the thermocline having odd modes that were induced by a moderate wind, standing surface seiches, and surface and internal seiches that rotate around the basin (Csanady, 1968a; Csanady, 1972).

Unlike the work on stratified lakes, meteorological forcing on estuarine circulation has always been overlooked or “averaged” out in observational studies prior to the 1970s (Pritchard, 1956). Since then, however, several studies sought to explain subtidal wind effects on circulation and more recently some supratidal wind effects (Wong and Moses-Hall, 1998b). Significant work has been done on the subtidal variability in estuaries resulting from atmospheric forcing. Subtidal circulation in the Providence River and the west passage of Narragansett Bay was found to be dominated by wind-induced fluctuations (Weisberg and Sturges, 1976; Weisberg, 1976). In the west passage, coherence between the longitudinal components of wind and net near surface current for periodicities of 2-3 days was high and the current lagged the wind by about 3 h. In the Providence River, coherence between the above variables was high for periodicities of 4-5 days and the current lagged the wind by about 4 h. Additionally, an 83 h record of salinity through the water column confirmed wind-driven upwelling and the generation of an internal seiche (Weisberg, 1976).

Subtidal circulatory response to wind forcing in the Chesapeake Bay was divided into remote and local effects (Elliott, 1978; Wang and Elliott, 1978; Wang, 1979b; Wang, 1979a). Investigators showed that almost half of the surface and mid-depth current fluctuations and most of the bottom current fluctuations were generated locally by winds and that subtidal sea level fluctuations at periods < 4 days were the result of seiche oscillations driven by the local, longitudinal winds. Additionally, barotropic volume exchange due to sea level fluctuations > 10 days in the lower bay was part of the response of the coupled bay-shelf system to non-local forcing over the continental shelf. A similar result was found for Corpus Christi Bay, Texas (Smith, 1977). Here, the meteorological exchanges appeared to be the result of wind stress forcing water directly onshore or offshore at periods of several days as well as the Ekman transport associated with the remote longshore component of the quasi-steady winds acting on the shelf.

Since the prior studies, the subtidal circulatory response to meteorological forcing has been broken into two types of effects: remote and local. Remote effects consist of winds acting on the continental shelf adjacent to a particular estuary to produce coastal sea level set up or set down at the mouth of the estuary and to possibly cause a coastal disturbance which propagates into the coastal areas
adjacent to the estuary in the form of a free wave (Noble and Butman, 1992). This effect dominated the subtidal variability in sea level data in the partially mixed San Francisco Bay (Walters, 1982; Walters and Gartner, 1985) as well as in the Delaware estuary (Wong and Garvine, 1984; Wong and Moses-Hall, 1998a), Great South Bay, New York (Wong and Wilson, 1984), the lower Chesapeake Bay (Paraso and Valle-Levinson, 1996), the mid-Chesapeake Bay (Vieira, 1986) and Sebastian Inlet, Florida (Liu, 1992). Local effects were more straightforward as they represented the effect of local wind stress acting on the surface of the estuary. They were most often seen in subtidal current variability (Wong and Garvine, 1984; Wong and Moses-Hall, 1998a; Liu, 1992; Vieira, 1986; deCastro et al., 2000). Additionally, local meteorological forcing changed flushing rates (Goodrich, 1988; Geýer, 1997) and was the cause of transverse variability in both current and salinity records (Wong, 1994; Wong and Moses-Hall, 1998b).

Wind-driven circulation altered the degree of stratification within the estuary. Often, wind speeds were sufficiently strong to vertically mix and destratify the water column (Goodrich et al., 1987). In other circumstances, however, wind-driven circulation was just strong enough to maintain stratification (de Kreeke and Robaczewska, 1989) or when combined with tidal effects, even intensified stratification (Valle-Levinson et al., 1998). Finally, one typical response to meteorologic forcing was the tilting of the pycnocline similar to the thermocline response in stratified lake systems. This tilting was also documented in the Chesapeake Bay (Sanford et al., 1990). In that investigation, it was found that wind forced lateral internal oscillations of the pycnocline in the mainstem of the Bay resulted in the advection of saline, hypoxic water from below the pycnocline onto the flanks of the Bay and into the lower reaches of adjoining tributaries.

1.4.3 Physical Studies of Freshwater Discharge

An estuary was defined as a semi-enclosed coastal body of water which has a free connection to the open sea, extending into the river as far as the limit of tidal influence, and within which sea water is measurably diluted with fresh water derived from land drainage (Dyer, 1997). Freshwater runoff affects circulation in an estuary by creating both a surface water slope from the head to the mouth of the estuary (barotropic effect) and a longitudinal density gradient through dilution of the saline water (baroclinic effect). The subtidal circulation within an estuary, termed estuarine or gravitational circulation, consists of a two-layer flow with fresh downstream flowing water at the surface and saltier, denser upstream flowing water near the bottom. Mixing of the two layers may occur along the zone separating the layers, changing the salinity structure both vertically and longitudinally (Stommel, 1953). Pioneering studies of the Chesapeake Bay established the framework for subtidal estuarine circulation and its effect on the salt balance in that coastal plain estuary (Pritchard, 1952; Pritchard, 1954; Pritchard, 1956). It was shown that the salt balance was maintained primarily by a longitudinal advective salt flux and a vertical non-advective salt flux.

Investigation into the axial variability of salinity stratification due to freshwater discharge was
conducted in the Delaware estuary. A seasonal relationship was found in that high discharge periods resulted in lower surface salinities and low discharge resulted in higher surface salinities at two locations in the estuary (Wong, 1995). With regard to isohaline displacement at shorter time scales, results indicated that within the ordinary range of river discharge variation, the Delaware estuary responded weakly to discharge. Instead, variations in isohaline displacement were associated with tidal advection (Garvine et al., 1992). In the Tamar estuary, England, the axial salinity distributions showed that the saltwater-freshwater interface was located closer to the head of the estuary during low runoff and was pushed further downstream during high runoff (Uncles and Stephens, 1990). A similar result held for the Gamtoos Estuary in South Africa (Schumann and Pearce, 1997). The classification of the Vellar Estuary in India varied depending on freshwater discharge (Dyer and Ramamoorthy, 1969). The estuary was classified as a salt wedge estuary during high discharge and well stratified at low discharge based on dimensionless stratification parameters. Stratification and destratification events in Mobile Bay were related to the relative strengths of the river discharge and winds (Schroeder et al., 1990). Destratification was observed during periods of strong river discharge where it essentially flushed the system, and during wind events. The river flow appeared to be the dominant control in destratification; winds became important in the absence of large freshwater discharge. In general, the halocline or thickness of the fresh, upper layer, remained substantially constant from the head to mouth of an estuary for a given river runoff (Pickard and Emery, 1990). However, that result changed as tidal and wind mixing were considered.

The Lateral variation of salinity and its relation to freshwater discharge has recently gained attention. In Mobile Bay, the near bottom salinity in the center channel was much higher than along both shores (Schroeder, 1977). This was often attributed to the effect of the Coriolis force on freshwater discharge (Pritchard, 1967; Wong, 1994), but recent evidence suggested that lateral variations may also be the result of channel geometry (Friedrich and Hamrick, 1996).

1.5 Numerical Studies of the NRE Circulation

The lower portion of the NRE was incorporated into a three-dimensional, finite difference model of Pamlico Sound (Pietrafesa et al., 1986). The model assumed time-dependence with a uniform wind and density field. Current velocities and sea surface elevations were output from that model. Wind-driven sea level setup and setdown in the lower NRE were simulated. The model verified that the wind field was the principal forcing function of the physical dynamics of Pamlico Sound and the lower NRE.

A two-dimensional, vertically integrated, unsteady flow and transport model was applied to the NRE from the head of the estuary to the mouth (Robbins and Bales, 1995). Like the model for Pamlico Sound, a uniform wind field was assumed. Current velocity, sea surface elevation and the salinity distribution were outputs from that modeling effort. The model was calibrated with data collected in the field and subjected to a sensitivity analysis. Results showed that flow in the NRE was highly nonuniform with opposing along channel flow along the shores and that magnitudes varied greatly throughout the
day and even in mean values for different periods. Simulated salinity exhibited lateral differences for all simulations.

A preliminary three-dimensional modeling application was developed for the NRE (Luettich et al., 1996). The hydrodynamic model was the state-of-the-art time-dependent, free-surface, nonlinear, three-dimensional and baroclinic model referred to as QUODDY (Lynch and Werner, 1991; Lynch et al., 1996). The density field evolved in time in response to prescribed buoyancy fluxes and with turbulence closure achieved with the Mellor-Yamada (1982) level 2.5 formulation and extensions described in Galperin et al (1988). The equations of motion were solved using the finite element method, in the horizontal and vertical, allowing for flexibility in the grid refinement required to capture the variability in the bathymetry and shoreline as well as within the water column in regions of the pycnocline. These were desirable features in a study of the Neuse as the exploration of the river’s circulation, its connections to creeks and neighboring sounds, and the simulation of the onset and sustenance of stratification required meshes capable of handling irregular shapes while embedding arbitrary resolution. For the NRE, the preliminary QUODDY application calculated velocity fields that looked reasonable based on the limited observational data that was available. However, no thorough model verification was attempted nor was any analysis work performed with the model to gain a deeper understanding of the behavior of the NRE.

As part of the modeling effort of the MODMON project, a two-dimensional, laterally averaged hydrodynamic/water quality model, CE-QUAL-W2, was applied to the NRE. This model considered the entire vertical water column and transport along the estuary. The model was used to examine the relationship between nutrient loading and the estuary’s water quality. Simulations of water motion and the longitudinal and vertical distributions of important water quality parameters such as dissolved oxygen and phytoplankton biomass were made for various nutrient loading and hydrologic scenarios. Here again, there was no attempt to discern causal factors driving system behavior.

As another part of the MODMON modeling effort, the 3-D hydrodynamic model EFDC and the biological process model, WASP, were calibrated for the NRE (US EPA Region 4, 2002). The EFDC model calculated flow fields that were then linked with the WASP model which was used to compute varying biological parameters like chlorophyll-a, dissolved oxygen and nutrient levels. The model was used to determine the effect of nutrient reduction schema on nutrient levels in the NRE in an effort to establish a TMDL for the NRE (NC Department of Environment and Natural Resources, 2002).

Finally, circulation and sea level were investigated for the lower NRE in its connection with Pamlico Sound (Luettich et al., 2002). In that study, the two-dimensional, finite element model ADCIRC was employed to determine the relationship between wind stress and circulation. It was found that winds directed along the axis of the lower NRE created a barotropic standing wave that traveled between Pamlico Sound and the NRE with a period 13.2 hrs.
Chapter 2

Methods and Materials

2.1 High Frequency, High Resolution Vertical Profiles

Accurate measurements of the onset of stratification and its vertical and temporal extent facilitated the understanding of the link between wind stress, stratification and the depletion of DO. Technological innovations such as water quality multiprobes with increased battery life, memory capacity and flexible programming capability allowed the design of a novel portable autonomous vertical profiler (P-AVP). The first P-AVPs for estuarine applications were produced specifically to study the lateral variation in the NRE and forward the knowledge of stratification and hypoxia development as well as salt advection (Reynolds-Fleming et al., 2002).

The core component of the P-AVP was a computer-controlled winch, which unwound and rewound cable to the appropriate depth. The P-AVP consisted of a computer controlled winch assembly and acoustical altimeter (WAAA), a water resistant housing and deployment platform, and a sensor package comprised of a multi-parameter water quality probe and an external data logger.

2.1.1 Computer Controlled WAAA

The WAAA consisted of a standard boat trailer winch (model T1200), a spring loaded tensioner, which provided constant tension on the winch spool (fabricated locally by Bircher Machinist), a 102 RPM DC motor (Dayton, model 1L475), a 12 V marine battery, an optical encoder, a digital acoustical altimeter (Datasonics, model PSA-916), and a micro-controller (ONSET Computer Corp., Tattletale model TFX-11).

The TFX-11 controlled power to the acoustical altimeter, optical encoder, and DC motor. It also queried the acoustical altimeter to determine the water depth and monitored the optical encoder to determine the angular displacement of the winch (figures 2.1 and 2.2).

The optical encoder was constructed at the Institute of Marine Sciences and consisted of a hand-fabricated encoder wheel and a slotted optical photoswitch (Opteck, model OPBS30W55). The encoder wheel was a thin plastic disk with two holes around the perimeter (figure 2.3). The wheel was attached to the motor shaft and positioned to allow the edge of the wheel to pass through the photoswitch. The photoswitch was a horseshoe-shaped electronic component with a light emitting diode on one side and a
phototransistor (detector) on the other side. If power was supplied to the photoswitch and the detector could see the emitter, the detector output was HIGH. If the detector could not see the emitter, the detector output was LOW. As the wheel turned, periodic interruptions in the beam, or state transitions, were counted by the TFX-11. With a winch spool (drum plus cable) diameter of 5 cm and an encoder wheel configured with two holes, vertical position was resolved to approximately 4 cm.

The TFX-11 had a real time clock, which allowed it to perform each profile according to pre-programmed times and duration. When the clock indicated that it was time to start a profile, the TFX-11 switched on power to the motor and optical encoder and raised the sensor package from its parked position to the surface. Upon reaching the surface, the TFX-11 switched off power to the motor and optical encoder and applied power to the acoustical altimeter. The TFX-11 collected an ensemble of depth readings (typically 10) from the altimeter and determined the average water depth. This water depth was logged by the TFX-11 if desired.

After the depth was determined, the TFX-11 calculated how many optical encoder state transitions were required to lower the sensor package to the bottom, based on the depth measurement and a calibration curve for the relationship between angular displacement of the motor shaft and the linear displacement of the winch cable. An estimate of 5-10 cm uncertainty existed in this calibration due to variability in the wrap of the cable on the winch drum. After calculating the number of state transitions, the motor was turned on and lowered the sensor package at a typical speed of 4 cm s⁻¹. When the bottom of the profile was reached, the sensor package was raised back to its parked position near mid-depth.

### 2.1.2 Water Resistant Housing and Deployment Platform

The design of the deployment platform (figure 4) was similar to that of a small floating dock. It was constructed of marine treated lumber and is 1.8 m x 1.8 m with a 0.31 m x 0.31 m hole in the center. Two marine grade floats provided buoyancy for both the platform and instrumentation. The platform was sturdy enough to moor a boat and large enough to hold two average sized adults. Two platforms were easily loaded onto a boat trailer for transport and were towed into place with a medium sized powerboat.

A water-resistant housing was supported by two pedestals on top of the platform. The WAAA package was raised above the platform to reduce corrosion from sea spray and to allow the sensor package to be brought to the surface for a complete sample of the water column.

The platform was moored using concrete anchors. Heavy (2.2 cm) chain was shackled to each anchor and connected to 2.54 cm nylon line. The nylon line was then secured to the corners of the platform so that approximately 1.5 m of chain was suspended in the water column for stability.
Figure 2.1: The top view of the winch assembly package showing the connections between motor, optical encoder, and winch. A 12 V marine battery supplied power to the 12 V DC motor, which turned both the optical encoder and trailer winch. The optical encoder measured the angular displacement of the winch and the tensioner provided constant tension for a smooth wrap on the drum.

2.1.3 Water Quality Probe and External Data Logger

A commercially available multi-parameter water quality probe (Hydrolab Datasonde, model 4a, HD4A) configured with conductivity, pressure, dissolved oxygen (with circulator) and temperature sensors was used to measure water quality conditions through the water column. The internal data logger that came with the HD4A was not flexible enough to accommodate our sampling schedule and therefore a separate, external data logger (ONSET Computer Corp., Tattletale Model 8) was used. The Model 8 and 6 "C" cell batteries were mounted in a waterproof housing that was strapped to the HD4A. The HD4A and external data logger housing were shackled to the winch cable and were easily removed for servicing.

The Model 8 had a real time clock that was synchronized with the clock on the TFX-11 and was used to turn the HD4A on and off according to the profiling schedule. When a profile began, the Model 8 activated its serial port, which was connected to the HD4A. The HD4A was configured to power itself up and send data at a rate of 1 Hz whenever its serial port was activated. The Model 8 wrote the incoming data to a flash memory card (Peripheral Issues, model Persistor CF8). When the profile was scheduled to end, the Model 8 turned off its serial port, causing the HD4A to stop sending data and to return to a low power mode. Recovering the stored data required swapping the flash memory card with a fresh card. Based on a typical lowering speed of 4 cm s$^{-1}$ and a 1 Hz sampling rate, profile data was collected at a vertical resolution of 4 cm.
Figure 2.2: A Tattletale mode TFX-11, the computer controller for the winch assembly and acoustical altimeter (WAAA), controlled the distribution of power to the DC motor as shown in this motor control design schematic.
Figure 2.3: The schematic of the optical encoder shows a plastic disc with two holes (gray area), which served as the encoder wheel. The photoswitch detected transitions in the light beam that were counted by the TFX-11 controller and used to determine how much cable had been paid out by the winch.
2.2 Measurement strategies

2.2.1 Weekly Hydrographic Sampling

Nineteen stations along the axis of the NRE were designated for hydrographic and nutrient data collection as part of the MODMON (MODeling and MONitoring) program initiated in 1997. UNC-IMS, Weyerhauser, and NC Division of Water Quality obtained salinity, temperature, dissolved oxygen, and pH data at 0.5 m increments from surface to bottom at these stations. All sampling was conducted using like instruments (Hydrolab Survey 3 and H2O CTD with dissolved oxygen and pH sensors) that were calibrated in the laboratory before each field deployment. Stations were sampled as synoptically as possible, typically within 3-4 hours.

A subsample of nine stations (Stations 0,20,30,60,70,100,120,140, and 160 on figure 2.4) was chosen from the original 19 and extends from the head of the estuary (at Streets Ferry Bridge) to near the mouth at station 160. The data were further reduced to near surface and near bottom hydrographic samples and sorted into seasonal time periods. The data spanned a time period from June 10, 1997 to November 27, 2000 and represented four summer and fall periods and three winter and spring periods. During the four summer periods, two of those samples represented major hurricane years with hurricane Bonnie in 1998 and Hurricanes Dennis, Floyd and Irene in 1999. There were no major hurricanes that affected the region during 1997 and 2000.

Mid-river contour plots were constructed from this data and provided a weekly snapshot of the hydrography and DO conditions and the river axis. Contours of the longitudinal profile can be viewed on the MODMON website. These data were also used to calibrate three water quality models including the 3-D finite difference model, EFDC. A portion of the data was also used as boundary condition data for driving the EFDC.

2.2.2 Monthly Pamlico Sound Hydrographic Sampling

Water quality data including salinity, DO, temperature, pH, and chl-a at 0.5m increments were collected at a selected site in Pamlico Sound (PS01 on figure 2.4). After Hurricane Floyd passed through the area in October of 1999, monthly sampling was initiated to determine the aftereffects of the Hurricane on the Sound. These data were used to determine a relationship between MODMON station 160 water quality and water quality at the mouth of the NRE. The selected Pamlico Sound station was located at the boundary of the 3-D EFDC model and served as a boundary condition for this model.

2.2.3 Lateral Hydrographic Samples

As part of the MODMON project, lateral hydrographic samples were taken in the summer of 1997 to determine if lateral variability was present in the NRE. Five transects (68,100,120,135,148; Figure 2.5) were sampled bi-weekly during calm days. Along each transect, vertical profiles were taken
Figure 2.4: A sub-sample of the 19 MODMON mid-river sampling stations were used to determine the seasonal climatology of the NRE. ◊ represent MODMON stations whereas ○ represents Pamlico stations at five sample stations approximately 1 km apart. Each vertical profile was taken with a Seabird SBE-19 with conductivity, temperature and dissolved oxygen sensors. The data were later quality checked, partitioned into near surface and near bottom values and interpolated using a kriging routine (Bohman-Carter, 1994) in Surfer.

2.2.4 High Frequency Spatial Sampling

A high frequency, high spatial resolution sampling trip was designed and then conducted July 7-8, 1999 to assess the spatial extent of wind-driven advection. Currents and hydrography were measured along three transects (T68, T95, and T120, fig.2.6) during July 7-8, 1999. The sampling design focused on repeated sampling of transect T95 while increasing spatial resolution by sampling transects T68 and T120. If sampling were limited by foul weather, the focus was shifted to transect T68 and T95. The general design was to begin at T95, collect velocity and hydrographic information along that transect and then move to T120. At T120, velocity and hydrographic information were collected before returning to T95 to collect the same information. Once T95 was completed, data at T68 were collected and then the cycle was completed with the return to T95.

Current speed and direction throughout the water column were measured using an Acoustic Doppler Current Profiler (ADCP) manufactured by RDInstruments. The 1200kHz ADCP was boomed-mounted and positioned 0.5m below the surface (Hench et al., 2000). Ship speed and flow magnitudes were calculated from Differential Global Positioning System (DGPS) and bottom tracking, respectively.
Depth bins of 25 cm were used and ensemble averaging of pings over a horizontal distance of 200m provided a flow speed accuracy of 2 cm s\(^{-1}\). Velocities from the near-surface and near-bottom depth bins were discarded due to velocity error associated with surface disturbances, ship wakes and side lobe reflection at the bottom. Velocity calculations from the ADCP thus provided an effective range from approximately 75cm below the surface to 25cm above the bottom.

Hydrography was measured by taking CTD casts at 5 stations along the transects (see fig. 2.6). CTD casts included collecting DO, salinity, pressure, and temperature data at each station using a Seabird CTD with attached DO sensor and pumped flow through system. The CTD was lowered continuously at a speed of approximately 1.5m min\(^{-1}\) and provided vertical profiles of conductivity, temperature, and DO at 2.5cm vertical resolution. DO and salinity measurements from the Seabird CTD were calibrated and verified in the field with either a YSI DO probe or a Hydrolab H20 CTD with DO sensor that was calibrated independently in the laboratory. Due to adverse field conditions, the sampling schedule was modified and is presented in table 2.1.
Figure 2.6: Locations for high frequency flow and hydrographic sampling for July 7-8, 1999.

<table>
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<tr>
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<th>Velocity</th>
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<th>Hydrograph</th>
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<td>H95</td>
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<td>15:38</td>
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Table 2.1: Actual velocity and hydrographic sample times from the high-frequency, high spatial sampling in July, 1999.
2.2.5 Long Term Data Acquisition

Velocity profiles

Two bottom-mounted, upward facing ADCPs were installed near the north and south shores of the upper NRE and collected data semi-continuously from June 13, 1999 to September 27, 2000 (figure 1.1). For reference purposes, these locations were termed 95N and 95S for the north and south shore locations, respectively. The ADCPs, 1200 kHz RDInstruments with high resolution/low flow enhancements, were deployed in approximately 4m of water in in-house designed stainless steel frames. The ADCPs measure flow velocities by transmitting sound at a fixed frequency and recording the echoes that have been Doppler shifted returning from sound scatterers in the water (Gordon, 1996).

The ADCPs were programmed to provide along and across channel velocities at a 20 cm vertical spacing throughout the water column (except for blanking regions of approximately 0.4 m near the surface and 0.8 m near the bottom). Data represented an ensemble average every three minutes. There were 120 pings (transducer emissions) per ensemble and this yielded a standard deviation of 0.4 cm s\(^{-1}\) in the velocity record at a sampling rate of 1.5 seconds. The instruments were serviced (data downloaded, cleaned, batteries replaced) every month during the summer (high bio-fouling period) and every 2 months otherwise. There were a total of 10 lateral deployments and one mid-channel deployment, see table 2.2.

CTD time series

Attached to each ADCP stand was a Microcat 37-SM manufactured by SeaBird Electronics, which recorded conductivity, temperature, and pressure at a fixed point near the bottom. The instruments were serviced (data downloaded, cleaned, batteries replaced) every month during the summer (high bio-fouling period) and every 2 months otherwise. When returned from the field, the salinity sensors were cross checked with a freshly calibrated Hydrolab.

CTD vertical profiles

Two P-AVPs were deployed approximately 25m shoreward of the ADCPs. As described earlier, the platforms raised and lowered a Hydrolab Datasonde 4a with conductivity, temperature, and DO sensor every 15min. The Hydrolab was lowered at a speed of 2.5m per minute and recorded samples at 1.0 Hz yielding a vertical resolution of 4cm (Reynolds-Fleming et al., 2002). The CTD/DO units were serviced approximately every 5 days to minimize the impact of bio-fouling on the DO measurements. Data from both P-AVPs were collected from June to August, 2000 (see table 2.3 for a complete deployment list).

Meteorological Data

Meteorological conditions such as wind direction, speed, air temperature and barometric pressure were collected at hourly intervals at the Cherry Point Meteorological Station. The station was located
10 m above the sea level. Large gaps in the Cherry Point data (> 5 days) were patched with New Bern meteorological data if available. Correlation between the two sites was high at $R^2=0.73$.

**Freshwater Discharge Data**

Quarter-hourly discharge data from a stream gauge located in Kinston, approximately 145 km upstream from New Bern was supplied by the United States Geological Survey (USGS). A drainage area ratio of 1.66 (11,600 km² drainage at New Bern vs. 7,000 km² drainage at Kinston) was applied to the data to estimate the average annual freshwater inflow at New Bern (Robbins and Bales, 1995).

**Moored salinity data**

Semi-continuous, quarter-hour water quality data from two stations (New Bern and Marker 11; figure 1.1) were collected by the USGS. These stations had both a near surface and near bottom sensor that measured temperature, salinity, pressure, and dissolved oxygen. These sensors were mounted on Coast Guard navigational pilings or on bridge struts and the data were available via the USGS web site.

**Water level gauge**

A Coastal Leasing MicroWave pressure sensor was deployed alongside navigational marker '1AC' (figure 1.1). Pressure and temperature were measured at 1Hz for 5 min every 30 min. Each 5min sample burst was averaged to remove high frequency surface gravity waves. This instrument was deployed specifically to provide a downstream boundary condition for the modeling efforts.

**2.3 Numerical simulation**

To better understand the driving forces as well as to increase the spatial and temporal coverage of the estuary, a mechanistic, three-dimensional, finite-difference model was calibrated for the NRE. Environmental Fluid Dynamics Code (EFDC), developed at Virginia Institute of Marine Science (J. M. Hamrick, 1992), was used to investigate the 3-D circulation and transport of the NRE. The physics and aspects of the computational scheme of the EFDC model were equivalent to the Princeton Ocean Model (Blumberg and Mellor, 1987) and the U.S. Army Corps of Engineers CH3D or Chesapeake Bay model (Johnson et al., 1993). The EFDC model solved the three-dimensional, vertically hydrostatic, free surface, turbulent averaged equations of motions for a variable density fluid. Dynamically coupled transport equations for turbulent kinetic energy, turbulent length scale, salinity and temperature were also solved. The turbulence parameter transport equations implemented the Mellor-Yamada level 2.5 turbulence closure scheme (Mellor and Yamada, 1982) that was modified (Galperin et al., 1988). The model used a stretched or sigma vertical coordinate and Cartesian or curvilinear, orthogonal horizontal coordinates and allowed for wetting and drying in shallow areas by a mass conservation scheme.
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Table 2.2: Deployment records for the ADCP and CTD.
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</tbody>
</table>

Table 2.3: Deployment record for the novel CTD profiling instrumentation.
The numerical scheme in the EFDC model was based on a second order spatial finite difference scheme on a staggered or C grid. Time integration employed a second order accurate three time level, finite difference scheme with an internal-external mode splitting procedure to separate the internal shear or baroclinic mode from the external free surface gravity wave or barotropic mode. The external mode solution was semi-implicit, and simultaneously computed the two-dimensional surface elevation field by a preconditioned conjugate gradient procedure. The external solution was completed by the calculation of the depth averaged barotropic velocities using the new surface elevation field (J. M. Hanrick, 1996).

The model’s semi-implicit external solution allowed large time steps that were constrained only by the stability criteria of the explicit central difference or high-order upwind advection scheme (Smolarkiewicz and Margolin, 1993) used for the nonlinear accelerations. The EFDC model’s internal momentum equation solution, at the same time step as the external, was implicit with respect to vertical diffusion. The internal solution of the momentum equations was in terms of the vertical profile of shear stress and velocity shear, which resulted in the simplest and most accurate form of the baroclinic pressure gradients and eliminated the over determined character of alternate internal mode formulations. Time splitting inherent in the three time level scheme was controlled by periodic insertion of a second order accurate two time level trapezoidal step. The EFDC model is also readily configured as a two-dimensional model in either the horizontal or vertical planes.

The EFDC model implemented a second order accurate in time and space, mass conservation fractional step solution scheme for the Eulerian transport equations for salinity, temperature, suspended sediment, water quality constituents, and toxic contaminants. The transport equations were temporally integrated at the same time step or twice the time step of the momentum equation solution (Smolarkiewicz and Margolin, 1993). The advective step of the transport solution used either the central difference scheme used in the POM model or a hierarchy of positive definite upwind difference schemes. The highest accuracy upwind scheme, second-order accurate in space and time, was based on a flux-corrected transport version of Smolarkiewicz’s multidimensional positive definite advection transport algorithm (Smolarkiewicz and Clark, 1986; Smolarkiewicz and Grabowski, 1990), which was monotonic and minimized numerical diffusion. The horizontal diffusion step, if required, was explicit in time, while the vertical diffusion step was implicit. Horizontal boundary conditions included time variable material inflow concentrations, upwinded outflow, and a damping relaxation specification of climatological boundary concentration.

Calibration of the model was completed using data collected as part of the MODMON project. Mid-channel data collected by IMS from January 1, 1998 through December 31, 2000 were used to calibrate the along channel portion of the model. The side channel of the estuary in the model was calibrated to data collected by NCSU in 1998 and 1999 (NC Department of Environment and Natural Resources, 2002). Independent validation of the model was done comparing model simulations of salinity, water level, and flow velocity for the 1999 year to moored CTD and ADCP data collected in the across
channel circulation study that was not used in the calibration procedure. Descriptive statistics such as the mean error and the root mean square error were calculated for the residuals of the model and field data to provide detail on model performance. Additionally, regression analysis was used to determine how well the model accounted for variability in the field data. Once it was determined that the model adequately reproduced conditions during the 1999 year, simulation of the year 2000 was completed using the same kinetic parameters with boundary and initial conditions based on the MODMON 2000 field data. Model performance was based on descriptive statistics comparing simulations of salinity and flow velocity with field data. Once the model performance was judged adequate, an in-depth investigation of the circulatory characteristics of the upper NRE was conducted during September and October, 2000 when massive fish kills were reported in the area.

Experimentation with the model was conducted to determine the influence that discharge and wind each had on lateral salinity variation in the upper NRE. In these cases, the initial condition for model salinity was set to a uniform 5 psu. The eastern boundary condition was steady in time, laterally uniform, but vertically non-uniform and represented a typical salinity curve with 9 psu at the surface and 20 psu at the bottom. Discharge and atmospheric input values were those collected during the year 2000. Four scenarios were addressed: 1) discharge as the only time dependent boundary condition; 2) atmospheric forcing as the only time dependent boundary condition; 3) atmospheric forcing and discharge as time dependent boundary conditions; 4) discharge as the only time dependent boundary condition but no coriolis force. The model was run for the first 25 days of 2000 and a snapshot at day 25 was compared for each scenario. This date was chosen so that freshwater discharge did not interfere with the east boundary condition.

### 2.3.1 Model Domain

The curvilinear-orthogonal horizontal model grid for the NRE was comprised of approximately 405 active quadrilateral water cells having faces ranging in length from approximately 2000m to less than 500m (figure 2.7). The model originally simulated four sigma or stretched layers which provided adequate vertical resolution for simulation year 1999 (US EPA Region 4, 2002). However, the P-AVPs were deployed during the year 2000 and provided better vertical resolution for model comparison. For this reason, the EFDC code was modified to increase vertical resolution to seven sigma layers during the year 2000 model runs, which gave a total of approximately 2800 computational cells.

Model bathymetry incorporated bathymetric data from previous modeling efforts (Bales and Robbins, 1999). As the model domain extended into the the southwest corner of Pamlico Sound, additional bathymetric data was acquired from NOAA and incorporated into the domain (figure 2.8).
Figure 2.7: The curvilinear-orthogonal, horizontal model domain for the NRE was comprised of 405 active quadrilateral cells.

Figure 2.8: The model bathymetry incorporated data from previous modeling efforts and bathymetric data from NOAA as it extended into the southwestern region of Pamlico Sound.
2.3.2 Model Boundary Conditions

The model grid was forced at its eastern open boundary by water surface elevation, temperature, and salinity data. Water level data was provided primarily by mooring 1AC as part of the MODMON program. Since this data was located inside the model domain, it was therefore projected onto the model boundary. Through an iterative process, a deterministic relationship was found whereby water level at the boundary was related to water level at marker 1AC with a lag of 1.25 hrs and a dampening of 0.8 (US EPA Region 4, 2002).

Salinity data for the boundary was gathered from MODMON stations 160, 170, 180, and 185 (see figure 2.4 as a reference). As the data was located inside the model domain, the data were interpolated to the boundaries. Originally, boundary salinity was determined by adding 3 psu to the salinity values recorded at station 160 of the MODMON program (US EPA Region 4, 2002). These values were visually compared with those at the downstream locations (for example, station 180). For simulation year 1999, this worked well as comparisons between model and field salinity were statistically similar.

For simulation year 2000, this relationship did not work well and a new one was determined. Based on the linear relationship between salinity ($S$) and distance downstream ($x$) from the climatology section, $dS/dx = 0.23$ psu km$^{-1}$ change along the bottom and 0.17 psu km$^{-1}$ change along the surface existed. Data from station 160 were divided into seven levels, which corresponded to the seven $\sigma$ levels in the model, and this relationship was applied to the data in each level. Overall, 3.45 psu to 4.67 psu in 0.2 psu increments was added to the MODMON 160 data to replicate salinity conditions in each of the $\sigma$ levels at the open boundary. To quality check this relationship, the salinity data were compared with salinity from the Pamlico Sound hydrographic sampling. There were 12 samples during the year 2000 and they were directly comparable to the model boundary conditions formulated from MODMON 160 data. The data were well correlated ($R^2 > 75\%$) and it was concluded that this was an adequate method for estimating the salinity boundary condition.

The western boundary of the grid was open and forced with freshwater discharge. Originally, discharge data in units of m$^3$s$^{-1}$ was provided by the following USGS markers: marker 02091814 in Fort Barnwell, marker 0209205053 at Swift Creek Station and marker 02092500 at Trent River. The majority of freshwater flows into the NRE from the Neuse and Trent Rivers (Giese et al., 1979). At the head of the estuary, the data from Swift Creek was added to the Ft. Barnwell data. At the Trent River Boundary, the data from the Trent River station were modified by a multiplier of 2.8 (US EPA Region 4, 2002). Monthly discharge data from Weyerhauser and the New Bern waste water treatment plant were also included as periodic inputs of freshwater and varying temperature (US EPA Region 4, 2002). Freshwater flows from tributaries were estimated daily using a watershed model that related freshwater flow to rainfall events (US EPA Region 4, 2002). Since prior analysis with field data was computed with discharge data from the Kinston gauge station, the Ft. Barnwell discharge data was replaced with the Kinston discharge data in order to better compare field and simulated data.
Surface boundary conditions were taken from available meteorologic data. Hourly air temperature, dew point temperature, precipitation, and wind direction and speed were acquired from the Cherry Point Marine Corps Air Station in Havelock, NC. Hourly solar radiation and cloud cover were provided by the Wilmington airport. The model assumed that atmospheric conditions were spatially uniform throughout the domain. For simulation year 2000, gaps in measured wind data were filled by linearly interpolating between values at the beginning and end of the gap. This interpolation was performed to maintain the temporal resolution of the surface boundary condition.

2.3.3 Statistical Comparison Techniques

Basic comparative statistics and their associated tests (Ambrose, 1987; Ambrose and Roesch, 1982; Reckhow and Chapra, 1983a; Reckhow and Chapra, 1983b; Reckhow et al., 1986; Reckhow et al., 1990; Thomann, 1982) were compiled into a recent book about hydrodynamic modeling (Martin and McCutcheon, 1999) and were used to validate the EFDC model.

The simplest statistical test was the mean error (ME) defined as

\[ ME = \frac{1}{N} \sum_{i=1}^{N} (O_i - S_i) \]

where \( O_i \) and \( S_i \) were observations and matching simulations for \( i = 1 \) through \( N \), the total number of observations. The units of the mean error were the same as those for the observations. The mean absolute error (MAE) defined as

\[ MAE = \frac{1}{N} \sum_{i=1}^{N} |(O_i - S_i)| \]

was also used. The closer to zero the MAE, the better the model accuracy.

Mean error or mean absolute error was a measure of model accuracy, bias, or systematic error. If \( ME > 0 \), the model systematically oversimulated and if \( ME < 0 \) the model systematically undersimulated the response of the natural system. Neither mean error nor mean absolute error measured imprecision or scatter in residuals (observed-simulated). Overall, the ME was useful for guiding calibration and judging validation testing, but other statistics were necessary to determine precision and correlation.

Precision of a simulation was determined from the standard error of estimate or root-mean-square error,

\[ RMSE = \sqrt{\frac{\sum_{i=1}^{N} (O_i - S_i)^2}{N}} \]

where the root mean square error, \( RMSE \), had the same units and dimensions as the observations and simulations. The dimensionless coefficient of variation or % root mean square error was \( \%RMSE = RMSE/O_a \) where \( O_a \) was the average of observations \( O_i \).
Correlation between model and observed data were completed using regression analysis and were more useful in evaluating the calibration or validation of a model. The regression equation was

\[ O_i = \alpha_1 + \beta_1 S_i + \epsilon \]

where \( \alpha_1 \) and \( \beta_1 \) were dimensionless regression coefficients representing the true y-intercept and slope, respectively, of a plot of observed versus simulated values, and \( \epsilon \) was the error in simulation \( S_i \). The coefficient of determination \( R \) (or for correlation coefficient \( R^2 \)) was

\[
R = \frac{N \sum_{i=1}^{N} O_i S_i - \sum_{i=1}^{N} O_i \sum_{i=1}^{N} S_i}{\sqrt{[N \sum_{i=1}^{N} O_i^2 - (\sum_{i=1}^{N} O_i)^2][N \sum_{i=1}^{N} S_i^2 - (\sum_{i=1}^{N} S_i)^2]}}
\]

and provided an indication of the degree of correlation between the observed \( (O_i) \) and predicted \( (S_i) \) data for a given number of observations, \( N \). The dimensionless correlation coefficient, \( R^2 \), approached one when a high degree of correlation occurred, and approached zero when data were not correlated. Whether the simulation was accurate was determined by testing the significance of the intercept \( \alpha_1 \) and the slope \( \beta_1 \). If the slope was not significantly different from one and the intercept was less than or equal to some tolerance centered about zero, the statistics indicated accuracy.

Salinity and velocity were chosen as the observed and simulated values to test the accuracy, precision, and correlation of the model to field data. The ME was used to determine model accuracy where the more accurate the model, the closer to zero the ME approached. The RMSE was used as an indicator of model precision with values close to 0 being more precise. Finally, the \( R^2 \) statistic was used as a measure of the variability in the observed data that was explained by the model.

2.4 Analysis Techniques for Measured Data

2.4.1 Descriptive Statistics

As an initial step, the measured data were quality checked and visualized in MATLAB. Means, variances, standard deviations, minimums, maximums and trends were estimated from the series. Climatology data were pooled into different seasonal categories such as winter (defined as December, January and February), Spring (March, April and May), Summer (June, July and August), and Fall (September, October and November) and statistics were calculated for each season individually.

2.4.2 Coordinate Transformations for Measured Data

Current velocity data from the ADCPs and wind velocity data from Cherry Point Marine Corps Air Station were transformed into an across and along channel coordinate system from the traditional N/E coordinate system. The new coordinate system for ADCP current velocity data was oriented 35° clockwise from the standard coordinate system. Flows downstream were represented by positive along channel flow whereas upstream flow was represented by negative along channel flow. Positive across channel flow was directed toward the north shore of the upper NRE.
Coherence was calculated between transformed wind data and water level data to determine the most appropriate coordinate transformation for the wind data. To accomplish this, coherence squared between wind data and water level were calculated (Wang and Elliott, 1978). Significant coherence (> 0.3) was then contoured over all frequencies, however, the focus was on frequencies lower than 0.05 cph (fig. 2.9). Highest coherences at diurnal to lower frequencies were found in a band of wind between 40° and 100° (fig. 2.9). 70° wind fell in the middle of this band. A similar wind was identified in previous studies (Pietrafesa et al., 1986; Luetich et al., 2000a) and was nearly aligned to the across channel orientation of the upper NRE and to the along channel orientation of the lower NRE (west of Minnesott Beach, NC). As a result, the new coordinate system for wind velocity data was oriented 70° clockwise from the standard coordinate system.

2.4.3 Correlational Analysis

The collected field data represented a discrete, non-stationary, and stochastic time series. Relationships between variables were inferred through correlational analysis, however cause and effect relationships were determined through experimental analysis.

As an initial step, the data were pooled for a particular deployment and placed into a large data matrix. The columns of the matrix represented the different variables collected and the rows represented the data at hourly intervals. From this matrix, pair-wise correlation was conducted and significant linear relationships identified.
Regression

As significant linear relationships were identified, a linear model was formulated using the method of least squares. The correlation coefficient, $R$, is used to determine the sign of correlation and then used to calculate the coefficient of determination, $R^2$. $R^2$ is used to construct a significance level for hypothesis testing whereby the null hypothesis is that the true correlation squared is zero (Emery and Thomson, 2001).

Spectral Analysis

Spectral analysis is a modification of Fourier analysis used for the analysis of stationary, stochastic functions of time. It was used to partition the variance of a time series as a function of frequency. Conventional spectral methods for oceanographic applications favor the direct periodogram approach whereby the data are transformed directly to obtain the Fourier components using the following (Emery and Thomson, 2001):

$$Y_k = \Delta t \sum_{n=1}^{N} y_n e^{-i2\pi f_k n \Delta t}$$

where

$Y_k$ = the fourier transform of the series $y_n$

$f_k$ = the frequencies that are confined to the Nyquist interval

$\Delta t$ = sampling interval

Depending on how the data is represented, the total area under the curve of the spectrum may be equal to the variance or the pseudo-variance of the process (Emery and Thomson, 2001). A peak in the spectrum shows an important contribution to the variance at frequencies within the corresponding interval (Chatfield, 1996).

The highest frequency that one can fit to the data is termed the Nyquist frequency, defined by $f_N = \frac{1}{2\Delta t}$ and expressed in cycles per unit time. For the CTD data, the Nyquist frequency was 10 cph and for the ADCP data, $f_N$ varied between 10 cph and 5 cph depending on the sampling schedule. The highest frequency that was resolved in the wind data is 0.5 cph since it was sampled hourly.

Prior to performing spectral analysis, the time series was demeaned and detrended in order to make the data stationary. In practice, the data were broken into a series of short overlapping segments and smoothed with a hanning window in order to increase the degrees of freedom per spectral estimate and to narrow the confidence limits to lend greater reliability to any observed spectral peaks (Emery and Thomson, 2001). A periodogram (or power spectra energy diagram) was then generated which related the frequencies to percentage of variance attributed to that frequency.
Cross-spectrum and coherence

Bivariate analysis involved the investigation of a relationship between two time series. In many ways, cross-spectral analysis is the time series analog of correlation and regression (Chatfield, 1996). The cross-spectrum of two series, which tells how oscillations within specific frequency bands are related in the frequency domain, is combined with the spectra of each series to create the squared coherency between the two series. The squared coherency is a dimensionless number between 0 and 1 that represents the fraction of the variance in one time series ascribable to the other time series through a linear relationship between the series. Additionally, the phase spectrum was calculated and indicates the angle (or time) by which one series leads or lags the other series as a function of frequency (Emery and Thomson, 2001).

Wavelet Analysis

Demeaning and detrending the data were essential preliminary steps for spectral analysis in order to make the data stationary. However, most oceanographic data are non-stationary and therefore the change in spectral energy over time is essential information. Wavelet analysis generates a localized, “instantaneous” estimate for the amplitude and phase of each spectral component in the data set. Where a Fourier transform of the non-stationary time series would smear out any detailed information on the changing processes, wavelet analysis tracks the evolution of the signal characteristics through the data set (Emery and Thomson, 2001).

Wavelet analysis involved the convolution of a real time-series, \( x(t) \), with a set of functions \( g_a(\tau, a) \) that were derived from a “mother wavelet” or analyzing wavelet, \( g(t) \), which was generally complex. In particular

\[
g_a(\tau, a) = \frac{1}{\sqrt{a}} g[a^{-1}(t - \tau)]
\]

where \( \tau \) (real) was the translation parameter corresponding to the central point of the wavelet in the time series and \( a \) (real and positive) was the scale dilation parameter corresponding to the width of the wavelet. For the Gaussian-shaped Morlet wavelet, the dilation parameter was related to a corresponding Fourier frequency (or wavenumber) (Emery and Thomson, 2001).

The continuous wavelet transform, \( X(t) \), of the time series with respect to the analyzing wavelet, \( g(t) \), was defined through the convolution integral

\[
X_g[\tau, a] = \frac{1}{\sqrt{a}} \int_{-\infty}^{\infty} g^*[a^{-1}(t - \tau)]x(t)
\]

in which \( g^* \) denoted the complex conjugate of \( g \) and variables \( \tau, a \) were allowed to vary continuously through the domain \((-\infty, \infty)\). Wavelet analysis provided a two-dimensional breakdown of a one-dimensional time series into position, \( \tau \), and amplitude scale, \( a \), as new independent variables.
A morlet wavelet for the transform was used for analysis as well as software provided by investigators at the University of Colorado (Torrence and Compo, 1998).

2.4.4 Upwelling Index

Data from the two P-AVPs were used to create an upwelling index. Data from the depths of 1.0m, 2.0m, and 3.0m were designated to represent the near surface, middle, and near bottom locations. A 30 day near continuous time series of data at 15-min intervals from these depths was used to compute means and standard deviations.

The lateral salinity index (LSI) and the tilt of constant salinity surfaces were used to pinpoint the occurrence of upwelling in time and space. The LSI was defined as the difference between demeaned salinity at the north shore profiling system location and demeaned salinity at the south shore profiling system location at each vertical depth (1.0m, 2.0m, and 3.0m). Demeaning the salinities removed any bias of inherent variability along one shore due to long term lateral variability and focused solely on the variation along each shore. A positive LSI indicated that salinity variability was higher along the north shore whereas a negative LSI indicated that salinity variability was higher at the south shore.

The information from the LSI was further supported by establishing the tilt in a particular salinity surface. Using both the LSI and this tilt information, it was determined that $LSI > 1$ indicated both a lateral salinity variation and a tilt in the isohaline downward towards the south shore. $LSI < -1$ indicated a lateral salinity variation and a tilt in the isohaline downward towards the north shore.

The severity of upwelling was defined and quantified based on the salinity surface depth difference in the middle of the water column. If the salinity surface appeared deeper in the water column along the opposite shore by less than 1.5m, then it was concluded that moderate upwelling or tilting occurred. If the salinity surface appeared deeper by more than 1.5m, then it was concluded that extreme upwelling occurred. These depths corresponded to an $LSI > 2$ for isohaline tilting downward toward the south shore and a $LSI < -2$ for isohaline tilting downward toward the north shore. Finally, this novel technique was used to compute the percentage of time that upwelling was present near each shore at each depth.
Chapter 3

Results: General Climatology

3.1 Discharge

Daily averaged discharge from Kinston is presented for years 1997 through 2000 (figure 3.1). The 1997 year was a comparatively dry one whereas year 1998 was comparatively wet. Hurricanes Dennis, Floyd and Irene passed through the region during September of 1999 and their passage was reflected in the large volume of freshwater influx in the later portion of 1999. The flooding associated with Hurricane Floyd has been termed a “Hundred Year” flood. Discharge ranged from $19 \text{ m}^3 \text{ s}^{-1}$ to $1050 \text{ m}^3 \text{ s}^{-1}$ over this four year time period.

All years were averaged together to determine a “mean” picture of annual discharge (figure 3.2). The influence of the flooding from Floyd biased this average and therefore another average was computed neglecting year 1999 (figure 3.3). Discharge was generally higher in the late winter and early spring. The summer season was normally dry with low freshwater flow. However, the random occurrence of hurricanes sometimes produced isolated cases of comparatively large freshwater flows in the late summer and early autumn.

3.2 Atmospheric Forcing

An average yearly time series of wind direction and speed was computed from wind data for the years 1997 through 2000. The mean wind direction rose and wind speed histogram are presented (figure 3.4). Winds from NE/N/NW dominated the yearly record and over 80% of the wind speeds were less than 4 m s$^{-1}$.

The breakdown into seasonal time periods showed somewhat different trends. Winds from the N/NE dominated winter and fall seasons (figure 3.5). Over 80% of the wind speeds remained less than 4 m s$^{-1}$ (figure 3.6). Winds during the Spring and Summer seasons were from all directions, however there were more occurrences of S/SW winds during the spring and summer. A higher percentage of winds ($> 20\%$) were “strong” ($> 5 \text{ m} \text{ s}^{-1}$) during the spring (figure 3.6).
Figure 3.1: Yearly time series of Kinston freshwater discharge for years 1997-2000.

Figure 3.2: Average daily discharge from Kinston for the years 1997-2000.
Figure 3.3: Average daily discharge from Kinston for the years 1997, 1998, and 2000.

Figure 3.4: Time series average wind direction rose and wind speed histogram for the years 1997 through 2000. Wind direction fractions indicate the direction wind is coming from.
Figure 3.5: Seasonal average wind roses from 1997 through 2000. Wind direction fractions indicate the direction wind is coming from.

Figure 3.6: Seasonal average wind speed histogram from 1997 through 2000.
3.3 Mean Longitudinal Temperature, Salinity and DO Distribution

The main definition of an estuary may be taken as “a semi-enclosed coastal body of water which has free connection to the open sea, extending into the river as far as the limit of tidal influence, and within which sea water is measurably diluted with fresh water derived from land drainage” (Dyer, 1997). As such, the salinity of the estuary and its range from the head to the mouth aid in characterizing the system. Additionally, since most estuarine systems are shallower than the open sea with which they are connected, various temperature gradients may exist in part due to this bathymetric discrepancy. Hence, the characterization of an estuarine system depends on understanding the salinity and temperature dynamics in that system which in turn, may aid in characterizing the DO distribution.

Water quality data from the MODMON project were used to investigate the relationship between the longitudinal distribution of salt, temperature, and dissolved oxygen. The four year mean longitudinal distributions of salinity and dissolved oxygen (DO) in the surface and bottom waters are shown in figure 3.7. The mean surface to bottom difference in salinity ranges from 0 psu at the head to 3 psu at the mouth. The mean near bottom salinity increased linearly from 0 psu at the head to 12 psu at the mouth. The mean near surface salinity increased linearly from 0 psu at the head to 9 psu at the mouth. A significant linear relationship between longitudinal distance from the head and salinity existed for both the near surface and near bottom salinity. Mean axial salinity gradients of $dS/dx = 0.23$ psu km$^{-1}$ existed near bottom and $dS/dx = 0.17$ psu km$^{-1}$ existed near the surface. The linear model for the bottom salinity was $S = 0.23x + 0.20$ and $S = 0.17x - 0.55$ for the surface salinity with correlation coefficients of $R^2 = 0.98$. Hence, longitudinal distance accounted for 98% of the variance in both the bottom and surface salinity.

The mean vertical salinity stratification was calculated as the absolute value of the difference between surface and bottom salinity. A significant linear relationship existed between longitudinal distance and vertical salinity stratification. The linear model of $\Delta S = 0.654x + 0.75$ fit the data with an $R^2$ value of 0.81. Hence, salinity in the bottom and surface water and vertical salinity stratification increased in the downstream direction.

The mean longitudinal bottom DO was fairly constant at approximately 6 mg l$^{-1}$ with decreased surface DO near the head of the estuary (figure 3.7). The mean surface to bottom difference in DO from approximately 13 km downstream of the head to the mouth was approximately 3 mg l$^{-1}$ . From the head of the estuary to $\sim 13$ km downstream, the difference reduced to less than 1 mg l$^{-1}$.

The four year mean longitudinal temperature distribution in the surface and bottom water is shown in figure 3.8. A fairly consistent surface to bottom temperature difference was present with a slight increase in this difference downstream, presumably the result of increasing depth.

For each of the nine stations, regressions were performed between salinity stratification and bottom DO and temperature stratification and bottom DO. An increase in both salinity and temperature stratification was weakly correlated with a decrease in bottom DO. A linear relationship between salinity
stratification and bottom DO explained between 20-35% of the variability in bottom DO. A linear relationship between temperature stratification and bottom DO explained only between 0.2-19% of the variability in bottom DO. An exponential regression was also conducted and it was found that significant relationships existed between variables, however, salinity stratification explained less than 30% of the variability in bottom DO and thermal stratification explained less that 10% of the variability in bottom DO. Pooling the data for all stations showed that changes in the vertical salinity stratification represented 41% of the variability in bottom water DO, however no relationship existed for vertical temperature stratification and bottom water DO.

The 4-yr mean data indicate a significant relationship between salinity stratification and DO stratification and between salinity stratification and bottom water DO. Since, benthic biological activity may be strongest during the warmer months (Lenihan and Peterson, 1998) it seemed appropriate to consider these relationships on a seasonal basis.

### 3.4 Mean Seasonal Longitudinal Temperature, Salinity and DO Variation

Mean seasonal trends were identified in the salinity, temperature and DO data (figures 3.9 and 3.10). Variations in salinity are primarily horizontal due to the conflicting buoyancy forces at the head and mouth of the system. Seasonally varying significant relationships between longitudinal distance and salinity are presented in table 3.1. Variations in temperature are mainly a vertical process due to differential heating or cooling. In some cases, temperature may be used as tracer as was done in the
Chesapeake Bay (Seitz, 1971). Seasonally varying significant relationships between longitudinal distance and temperature are presented in table 3.2.

### 3.4.1 Summer (June-August)

Maximum salinity and temperature values in both the bottom and surface waters occurred in the summer (figs. 3.9a and 3.10a). Salinity stratification increased rapidly from the head to approximately 24 km downstream where the estuary widens. A linear relationship suggests that distance downstream explains over 98% of the variability in salinity stratification (fig. 3.11a; table 3.3). Beyond 24 km, salinity stratification slightly decreased downstream but remained near 3 psu. Distance downstream explains over 97% of the variability in salinity stratification beyond 24 km. There was also a reasonably constant thermal stratification of about 1°C (fig. 3.10). Bottom and surface DO were lowest during this time period (fig. 3.9a). Mean bottom DO fell below 5 mg l⁻¹ for the entire length of the estuary. Salinity stratification explains over 75% of the variability in bottom DO when all the data were considered. When the near zero salinity stratification data are omitted, however, salinity stratification explains roughly 52% of the variability in bottom DO (fig. 3.12a; table 3.4). These zero salinity stratification points are omitted since the focus is on the effects of stratification on bottom DO. There was no relationship between thermal stratification and bottom water DO. The NRE appears divided by the bend as bottom DO showed a significant relationship with distance downstream. Distance downstream explained over 93% of the variability in bottom DO from the head of the estuary to about 42 km, the bend (fig. 3.13a; table 3.5).
<table>
<thead>
<tr>
<th>Season</th>
<th>location</th>
<th>Linear model</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summer</td>
<td>Surface</td>
<td>$S = 0.24x - 0.44$</td>
<td>0.98</td>
</tr>
<tr>
<td>Summer</td>
<td>Bottom</td>
<td>$S = 0.28x + 1.02$</td>
<td>0.95</td>
</tr>
<tr>
<td>Fall</td>
<td>Surface</td>
<td>$S = 0.16x + 0.13$</td>
<td>0.98</td>
</tr>
<tr>
<td>Fall</td>
<td>Bottom</td>
<td>$S = 0.21x + 0.64$</td>
<td>0.99</td>
</tr>
<tr>
<td>Winter</td>
<td>Surface</td>
<td>$S = 0.15x - 0.81$</td>
<td>0.98</td>
</tr>
<tr>
<td>Winter</td>
<td>Bottom</td>
<td>$S = 0.19x + 0.42$</td>
<td>0.98</td>
</tr>
<tr>
<td>Spring</td>
<td>Surface</td>
<td>$S = 0.14x - 1.16$</td>
<td>0.93</td>
</tr>
<tr>
<td>Spring</td>
<td>Bottom</td>
<td>$S = 0.21x - 1.42$</td>
<td>0.96</td>
</tr>
</tbody>
</table>

Table 3.1: Linear regression models of longitudinal distance and salinity in the near surface and near bottom waters for all four seasons.

<table>
<thead>
<tr>
<th>Season</th>
<th>location</th>
<th>Linear model</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summer</td>
<td>Surface</td>
<td>$T = -0.02x + 28.18$</td>
<td>0.88</td>
</tr>
<tr>
<td>Summer</td>
<td>Bottom</td>
<td>$T = -0.02x + 27.19$</td>
<td>0.70</td>
</tr>
<tr>
<td>Fall</td>
<td>Surface</td>
<td>$T = 0.02x + 19.27$</td>
<td>0.78</td>
</tr>
<tr>
<td>Fall</td>
<td>Bottom</td>
<td>$T = 0.02x + 18.92$</td>
<td>0.75</td>
</tr>
<tr>
<td>Winter</td>
<td>Surface</td>
<td>$T = 0.001x + 9.78$</td>
<td>0.98</td>
</tr>
<tr>
<td>Winter</td>
<td>Bottom</td>
<td>NS</td>
<td></td>
</tr>
<tr>
<td>Spring</td>
<td>Surface</td>
<td>$T = -0.03x + 18.09$</td>
<td>0.90</td>
</tr>
<tr>
<td>Spring</td>
<td>Bottom</td>
<td>$T = -0.04x + 17.51$</td>
<td>0.88</td>
</tr>
</tbody>
</table>

Table 3.2: Linear regression models of longitudinal distance and temperature in the near surface and near bottom waters for all four seasons. NS stands for Not Significant.
Figure 3.9: Mean seasonal trends in the longitudinal distribution of salinity and DO. Error bars represent the 95% confidence interval about the mean.
Figure 3.10: Mean seasonal trends in the longitudinal distribution of temperature and DO. Error bars represent the 95% confidence interval about the mean.
Beyond the bend, mean bottom DO increased considerably to just above hypoxic values.

3.4.2 Fall (September-November)

Salinity and temperature values during the fall were the second highest behind mean summer values (figs. 3.9b and 3.10b). Salinity stratification appeared to increase linearly downstream and in fact, downstream distance explains about 86% of the variability in salinity stratification (fig. 3.11b; table 3.3). Thermal stratification was minimal and less than 0.5°C (fig. 3.10b). DO values in both the surface and bottom waters were higher than summer values but lower than the other seasons. No statistically significant relationship existed between salinity stratification and bottom water DO (fig. 3.12b; table 3.4) or distance downstream and bottom water DO (fig. 3.13b; table 3.5).

3.4.3 Winter (December-February)

Temperatures were lowest during the winter and salinities were the third highest behind the summer and fall values (fig. 3.9c and 3.10c). Similar to the dynamics during the mean summer months, the NRE was partitioned into two separate regions based on salinity stratification (fig. 3.11c; table 3.3). Salinity stratification increased rapidly from the head of the estuary to about 25 km downstream where the estuary widens. Distance downstream explains over 98% of the variability in salinity stratification. Beyond 25 km, salinity stratification reduces linearly to almost 2.5 psu. Temperature stratification increased linearly from the head of the estuary, yet values were typically well below 0.5°C. DO reached maximum mean values in both the surface and bottom layers in the winter (fig. 3.9c). The mean DO stratification was also the smallest during this time. No statistically significant relationship was found between salinity stratification and bottom water DO (fig. 3.12c; table 3.4) or thermal stratification and bottom water DO. From the head of the estuary to approximately 13 km downstream, bottom DO decreases fairly rapidly, but still remained well above hypoxic values (fig. 3.13c). Beyond this distance, however, bottom DO increases linearly and distance explains about 92% of the variability in bottom DO (fig. 3.13c; table 3.5).

3.4.4 Spring (March-May)

The effect of freshwater discharge was evident during the spring as increased discharge pushed the salt wedge downstream and generally freshened the system (fig. 3.9d). Temperature decreased downstream in both the surface and bottom data during spring and thermal stratification was evident further downstream (fig. 3.10d). Salinity stratification was minimal until 20 km downstream and gradually increased towards the mouth. A significant linear relationship between longitudinal distance and ΔS existed with $R^2 = 0.91$ (fig. 3.11d; table 3.3). A significant linear relationship existed between longitudinal distance and ΔT with $R^2 = 0.44$. DO values were above 7 mg l$^{-1}$ in both the surface and bottom layers and represent the second highest values behind winter values (fig. 3.9d). Salinity stratification
Figure 3.11: Seasonal comparisons between distance downstream and salinity stratification.

explains almost 79% of the variability in bottom DO when zero stratification values were neglected (fig. 3.12|table 3.4). The near zero stratification values coincide with the upper NRE before it widens and reflects the effect of freshwater discharge on the system pushing the salt wedge towards the mouth of the estuary. In fact, distance downstream explains about 66% of the variability in bottom DO from 13km to the bend in the estuary (fig. 3.13|table 3.5).
<table>
<thead>
<tr>
<th>Season</th>
<th>Distance km</th>
<th>Linear model</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summer</td>
<td>0-25</td>
<td>$\Delta S = 0.16x + 0.11$</td>
<td>0.98</td>
</tr>
<tr>
<td>Summer</td>
<td>25-60</td>
<td>$\Delta S = -0.03x + 4.38$</td>
<td>0.97</td>
</tr>
<tr>
<td>Fall</td>
<td>0-60</td>
<td>$\Delta S = 0.07x - 0.25$</td>
<td>0.86</td>
</tr>
<tr>
<td>Winter</td>
<td>0-25</td>
<td>$\Delta S = 0.13x + 0.14$</td>
<td>0.99</td>
</tr>
<tr>
<td>Winter</td>
<td>25-60</td>
<td>$\Delta S = -0.02x + 4.07$</td>
<td>0.84</td>
</tr>
<tr>
<td>Spring</td>
<td>13-50</td>
<td>$\Delta S = 0.06x + 0.51$</td>
<td>0.91</td>
</tr>
</tbody>
</table>

Table 3.3: Linear regression models of longitudinal distance and salinity stratification for all four seasons.

<table>
<thead>
<tr>
<th>Season</th>
<th>$\Delta S$</th>
<th>Linear model</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summer</td>
<td>1-4</td>
<td>$DO = -0.17x + 5.21$</td>
<td>0.52</td>
</tr>
<tr>
<td>Fall</td>
<td>1-4</td>
<td>$DO = 0.18x + 4.92$</td>
<td>NS</td>
</tr>
<tr>
<td>Winter</td>
<td>1-4</td>
<td>$DO = 0.14x + 8.69$</td>
<td>NS</td>
</tr>
<tr>
<td>Spring</td>
<td>1-4</td>
<td>$DO = -0.33x + 8.31$</td>
<td>0.79</td>
</tr>
</tbody>
</table>

Table 3.4: Linear regression models of salinity stratification and bottom water DO for all four seasons. NS means not significant.

Comparing seasons, salinity stratification followed nearly linear relationships with longitudinal distance for both spring and fall. Spring is generally a period of high freshwater discharge whereas the fall tends to be drier with periodic influxes of freshwater due to hurricanes. The linear models are summarized in table 3.3. The estuary appeared divided into two systems for both the summer and winter means. Salinity stratification rapidly increased from the head to approximately 25km downstream. This distance represents the location where the estuary widens. Beyond this distance, salinity stratification slightly reduced. These relationships are also summarized in table 3.3. Summer is generally a low discharge period and although higher discharge values are recorded at the end of the winter season, there is generally a time delay between when discharge occurs and when the higher volumes of water influence salinity in the estuary. Hence, the reduction in freshwater discharge may play a role in the

<table>
<thead>
<tr>
<th>Season</th>
<th>Distance km</th>
<th>Linear model</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summer</td>
<td>0-40</td>
<td>$DO = -0.05x + 5.03$</td>
<td>0.93</td>
</tr>
<tr>
<td>Fall</td>
<td>0-60</td>
<td>$DO = 0.008x + 5.2$</td>
<td>NS</td>
</tr>
<tr>
<td>Winter</td>
<td>0.13</td>
<td>$DO = -0.07x + 9.35$</td>
<td>0.99</td>
</tr>
<tr>
<td>Winter</td>
<td>13.60</td>
<td>$DO = 0.03x + 7.92$</td>
<td>0.92</td>
</tr>
<tr>
<td>Spring</td>
<td>13.42</td>
<td>$DO = -0.03x + 8.54$</td>
<td>0.95</td>
</tr>
</tbody>
</table>

Table 3.5: Linear regression models of longitudinal distance and bottom water DO for all four seasons. NS means not significant.

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Figure 3.12: Seasonal comparisons between salinity stratification and bottom water DO.
Figure 3.13: Seasonal comparisons between distance downstream and bottom water DO.
unique division of the estuary based on salinity stratification.

Bottom water DO generally decreased downstream during the summer and spring seasons. Interestingly, these two seasons were the only ones in which a significant relationship between bottom water DO and salinity stratification existed. Salinity stratification explained approximately 79% of the variability in bottom water DO during the spring and 52% of the bottom water DO variability in the summer when near zero salinity stratification was not considered. In fact, mean bottom DO dropped by 1.7 mg l\(^{-1}\) from the winter to the spring season and by 3.5 mg l\(^{-1}\) from the spring to the summer season whereas it rose by 1.5 mg l\(^{-1}\) from the summer to the fall season and by 3.7 mg l\(^{-1}\) from the fall to the winter season. As salinity stratification values did not change very much between seasons an additional causative agent is required to explain this phenomenon. The increase in temperature which would allow for benthic respiration to begin may explain this relationship. The importance of biological activity is especially evident when comparing the stratification of the winter and summer seasons. These two seasons showed similar partitions of the estuary based on salinity stratification, yet, hypoxia is evident only in the summer mean. It is well known that biological activity is diminished during the winter as metabolisms slow due to cooler temperatures. This likely explains the temporal difference in oxygen distribution during the summer and winter despite similar salinity stratification.

3.5 Generation of Low Dissolved Oxygen

Mid-river salinity stratification was fairly consistent seasonally. However, substantial differences in DO existed between seasons: Low DO bottom water was generated during warmer weather and was coincident with strong stratification. In the spring, summer, and winter, strong relationships between distance downstream and bottom water DO existed. During the spring and summer, there were also strong relationships between salinity stratification and bottom water DO suggesting that increases in stratification lead to decreases in bottom water DO. Lowest DO values generally occurred near the bend in the estuary which also coincided with regions of strong stratification. Hypoxia is particularly evident during the summer seasons.

Comparisons of the important processes necessary for the generation of low DO water can be addressed with first order knowledge of the advective and dispersive properties behind conservative tracers. For example, variability in bottom DO can be explained in the simplest form as a dynamic balance between horizontal advection, vertical mixing and oxygen sources and sinks. Over the seasonal mean, the lateral variation in bottom DO in the upper NRE can be assumed negligible (Buzzelli et al., 2002) and an oxygen balance equation focused in the along channel orientation may be written for the NRE.

Following a technique used in the Chesapeake Bay (Boicourt, 1992), a simple balance for the bottom water DO, following a parcel of water is written as:
\[
\frac{DO_2}{Dt} = \frac{\partial O_2}{\partial t} + u \frac{\partial O_2}{\partial x} = Q_o + S_i
\]

where \( u \) is the mean velocity component in the along channel direction, with positive flow directed out of the estuary; \( Q_o \) is the vertical exchange between layers term; and \( S_i \) is the net source/sink term representing the difference between consumption and production in both the water column and bottom sediments. Knowledge of the physical terms in the above balance can be used to determine rates of the biological term. This was done for a section of the upper NRE at approximately 33 km downstream (see figure 1.2).

Bottom water DO decreases during the spring and summer seasons. From the spring to the summer, there is a mean loss of 4.5 mg l\(^{-1}\). Over the season, which consists of 90 days, this rate is -0.05 mg l\(^{-1}\) d\(^{-1}\).

The mean summer distribution of DO and mean along channel velocity were used to determine the advective contribution of low DO. From the view of an observer at one location in the estuary, the upestuary flow in the upper NRE represents an influx of lower DO water. The longitudinal oxygen gradient was estimated as linear from 0-40 km (fig. 3.13a) and a rate of -0.05 mg l\(^{-1}\) km\(^{-1}\) of oxygen decrease was formulated from this relationship. As is shown in Chapters 4 and 5, an along channel deep water velocity \( u \) is approximately -3 cm s\(^{-1}\). The advective contribution to the DO balance in the lower layer is then 0.13 mg l\(^{-1}\) d\(^{-1}\).

\( Q_o \), the source term due to vertical exchange from the upper layer, can be estimated from a conservative tracer, like salinity. For example, a mean seasonal salt balance would constitute a balance primarily between upstream advection of salt and vertical mixing and may be represented as

\[
\frac{\partial S}{\partial t} + u \frac{\partial S}{\partial x} = Q_s.
\]

Bottom salinity increases from the spring to the summer seasons by approximately 6 psu over a 90 day period. This corresponds to a gain in salinity of 0.067 psu d\(^{-1}\). From the perspective of a volume of water, saltier water enters the region from the downstream direction and less salty water leaves the region towards the head of the estuary. Hence, there is a gain of salinity in the upper layer, \( Q_s \), \( \partial S/\partial x \) during the summer was given as 0.28 psu km\(^{-1}\) in the bottom water (see table 3.1). With a mean velocity of -3 cm s\(^{-1}\), the salt advective term becomes -0.726 psu d\(^{-1}\) and hence \( Q_s = -0.66 \) psu d\(^{-1}\).

Vertical exchange itself, \( Q_s \), may be represented as \( K_s(S_b - S_u) \) where \( S_b \) is the mean seasonal salinity in the bottom layer and \( S_u \) is the mean seasonal salinity in the upper layer. On a seasonal time scale, this vertical exchange may be due in part to the two-way turbulence process characteristic of partially mixed estuaries (Dyer, 1997), but it also may be due to wind-driven upwelling events and near complete mixing events. Near 34 km downstream, mean summer salinity in the lower layer is around 12 psu whereas salinity in the upper layer is around 8 psu. Hence, \( K_s = -0.19 \) d\(^{-1}\).
Assuming the same mechanisms that cause the vertical exchange of salinity also cause the vertical exchange of dissolved oxygen, this parameter may be used to estimate $Q_o$ as $Q_o = K_s(O_L - O_U)$. Here, mean summertime oxygen in the bottom layer is 3.2 mg l$^{-1}$ and mean summertime oxygen in the upper layer is 7.6 mg l$^{-1}$. Hence, $Q_o = 0.85$ mg l$^{-1}$ d$^{-1}$, a gain in DO due to exchange with the upper layer.

The rate of consumption, $S_i$, during the summer season was estimated from the oxygen balance to be -0.77 mg l$^{-1}$ d$^{-1}$ (see table 3.6). Rates of consumption were also determined for the spring and winter seasons and are presented in table 3.6. $S_i$ was not calculated for the fall season since no relationship existed between bottom DO and distance downstream (see figure 3.13 and table 3.4). The highest oxygen demand occurred in the summer, followed by the spring and was lowest during the winter season (see table 3.6). This is consistent with known temperature controls on consumption (Buzzelli et al., 2002).

Benthic oxygen demand has been measured in the NRE over the years. Alperin, et al (2000) computed an average benthic demand from previous studies combined with their study based on the use of sediment chambers (Luetich et al., 2000a). The benthic flux of oxygen was estimated as $25 \pm 10$ mmol m$^{-2}$ d$^{-1}$, which is equivalent to a rate of $0.4 \pm 0.16$ mg l$^{-1}$ d$^{-1}$, assuming a characteristic lower layer thickness of 2m. Pelagic oxygen demand was estimated at $32 \pm 23$ mmol m$^{-3}$ d$^{-1}$ (Sauber, 1998). This is a pelagic demand rate of $1.024 \pm 0.736$ mg l$^{-1}$ d$^{-1}$. The total mean oxygen demand, $S_i$ would therefore be $1.424 \pm 0.896$ mg l$^{-1}$ d$^{-1}$. The mean measured rate of oxygen consumption is somewhat higher than the $S_i$ calculated from the oxygen budget although the calculated $S_i$ falls within the uncertainty range of the measured $S_i$. Since the calculated $S_i$ includes both production and consumption, it is reasonable to expect it to be lower than measurements of DO consumption alone.

It is clear that vertical exchange, $Q_o$, and biological consumption, $S_i$, are the lead players controlling the oxygen balance in the bottom waters. However, the advective contribution to the balance does play an important role in the upper NRE from about 20km to 40km downstream, an area upstream of the bend (see figure 1.2). Ignoring the change in oxygen over time exemplifies the fact that upstream advection of low DO water and biological consumption combined are larger than the vertical mixing term during the summer and therefore a decrease in bottom DO results. This is also true during the spring. Interestingly, the advective term becomes a positive contributor to the balance during the winter and the advective term and biological demand together are less than vertical mixing. Hence, bottom DO increases in the upstream direction during the winter.

Spatial location in the NRE is important and the above relationship between advective processes, vertical mixing, and biological demand may not be consistent at other locations in the estuary. For example, east of the bend, bottom DO increases towards the mouth in the spring, summer, and winter (figure 3.13). Hence, the advective process in a region defined downstream of the bend would be an additive contribution to the bottom DO. In fact, $\partial O / \partial x \approx 0.05$ east of the bend and hence, the advective term would be -0.13 mg l$^{-1}$ d$^{-1}$. If biological demand is considered constant throughout the NRE, then
<table>
<thead>
<tr>
<th>Season</th>
<th>$\partial O_2/\partial t$</th>
<th>$u(\partial O_2/\partial x)$</th>
<th>$Q_o$</th>
<th>Calculated $S_i$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summer</td>
<td>-0.05</td>
<td>0.13</td>
<td>0.85</td>
<td>-0.77</td>
</tr>
<tr>
<td>Spring</td>
<td>-0.02</td>
<td>0.08</td>
<td>0.53</td>
<td>-0.47</td>
</tr>
<tr>
<td>Winter</td>
<td>0.04</td>
<td>-0.08</td>
<td>0.33</td>
<td>-0.38</td>
</tr>
</tbody>
</table>

Table 3.6: Seasonal components of the oxygen balance equation for the region of the NRE from 13km downstream towards the bend.

during the summer, vertical mixing would become smaller than biological consumption and $Q_o < S_i$. There is no innate reason to consider a constant biological demand, however, a reduced biological demand combined with the advective term would only mean a reduced vertical mixing term. The relationship between vertical mixing and biological demand would still hold.

The importance of vertical mixing as an explanation for the variable nature in bottom DO during the summer months in the NRE has been alluded to prior to this study. As this budget confirms, vertical mixing primarily balances the biological oxygen demand. Spatial variation in vertical mixing exists in the NRE. Specifically, vertical mixing is of a larger magnitude in the upper NRE than in the lower NRE. As is presented in the following chapters, the orientation of the upper NRE to the prevailing winds causes considerable lateral variability. This may be an explanation for the disparity evident in vertical mixing values between the upper and lower NRE.

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Chapter 4

Results

4.1 Measured Lateral Variability

The basic circulation theory of estuaries was developed from studying the Chesapeake Bay and assuming lateral uniformity (Pritchard, 1952; Pritchard, 1954; Pritchard, 1956). Since that time, it has become widely recognized that lateral variation in salinity and circulation can be an important mass transport mechanism (Fischer, 1972) which influences the water quality in the Chesapeake and other estuaries. Lateral variability within an estuary can be attributed to many things: variation in the topography, tidal asymmetry, effects of the Coriolis force, and wind-driven variations (Wong and Moses-Hall, 1998b; Kjerve and Proehl, 1979; Chant and Wilson, 1997; Mertz and Gratton, 1995; Valle-Levinson and Atkinson, 1999; Blanton and Andrade, 2001). Since the tidal influence on the NRE is small (Luettich et al., 2000a), reasonable mechanisms that may lead to lateral variability in circulation, salinity and therefore dissolved oxygen, are freshwater discharge and meteorological forcing.

4.1.1 Influence of Freshwater Discharge

General Circulation

As part of this study, two bottom-mounted, upward facing ADCPs were deployed along the shores of the upper NRE. The data were divided into appropriate seasonal time periods and averaged providing an average vertical profile of velocity at the two sites. The focus of this research was on the summer 1999 and the winter 2000 time periods since the data collected then were the most continuous.

Downstream flow predominated along the south shore for the summer season with stronger flow towards the surface (fig. 4.1). However, in the winter of 2000, the mean velocity profile followed that defined as “typical estuarine flow” with small upstream flow in the near bottom waters (fig. 4.2). Upstream flow was consistent along the north shore for both seasons. A three layer flow was evident along the north shore with upstream flow near the surface, downstream flow in the middle of the water column and upstream flow near the bottom. Comparatively weaker flows were recorded in the across channel velocity. Along the south shore, the mean velocity was oriented northward throughout the water column. A two layer flow persisted at the north shore with southward flow in the surface and northward or very low flow near the bottom.
Figure 4.1: Average velocity profiles along the north and south shores for the summer season of 1999. The dotted lines surrounding the mean profile represent the 99% confidence interval about the mean. Positive values represented downstream and northward flow.

This information implied that the Coriolis force played a role in the deflection of more buoyant waters toward the south shore of the estuary. The Rossby number,

\[ R_o = \frac{U}{L f} \]

where \( f = 2\Omega \sin \phi \), is a dimensionless number that assesses the importance of rotation for a particular phenomenon. Using characteristic values for the NRE (\( U=10 \text{ cm/s}, L=5 \text{ km}, \Omega = 7.29 \times 10^{-5} \text{ s}^{-1} \)), and \( \phi = 35^\circ \), the Rossby number was calculated as \( R_o = 0.2743 < 1 \) and therefore the earth’s rotation should have played a role in the along channel circulation of the NRE and may have in fact deflected lighter, fresher water towards the southern shore (Pedlosky, 1987).

**Water Level Variation**

Water level that spanned a 308 day period from June, 1999 through the September, 2000 were averaged together and were compared with freshwater discharge. A lagged correlation was computed to determine how discharge affects water level (figure 4.3). Here, a small, but statistically significant positive correlation between water level and discharge existed. As discharge increased, water level also increased. The change in water level lagged changes in discharge by approximately 30 days.
Figure 4.2: Average velocity profiles along the north and south shores for the winter season of 2000. The dotted lines surrounding the mean profile represent the 99% confidence interval about the mean. Positive values represented downstream and northward flow.

Figure 4.3: Lagged correlation between water level and discharge from June 1999 through September 2000. Values outside -0.1 and 0.1 are significant with a 95% confidence interval.
Salinity Variation

Lateral variation was present in the bottom salinity. Table 4.1 lists the mean salinity and standard deviation for all the bottom-mounted CTD deployments (see table 2.2 for schedule). T-test comparisons with the null hypothesis, \( H_0 \), defined as “the means of a deployment are equal” showed that mean salinity values along the north and south shore were statistically similar for deployment one with a p-value of 0.165. If rejection of the null hypothesis was statistically possible, alternative hypotheses were formed. Alternative hypothesis were presented as \( H_1 \) = “the mean of the north shore is greater than the mean of the south shore” and \( H_2 \) = “the mean of the south shore is greater than the mean of the north shore.” Accordingly, mean salinity on the south shore was statistically higher for deployment 8 (June, 2000; table 2.2) only and the mean salinity on the north shore was statistically higher for the other eight deployments. Combining the entire data set into more than a year long record and computing an ANOVA between the north and south shores yielded the fact that statistically, the mean bottom salinity along the north shore was higher than the mean bottom salinity along the south shore.

Peaks in the spectral energy of bottom salinity and discharge were investigated using the same 308 day data set that spanned the summer of 1999 through the fall of 2000 used for water level. Power spectral density plots showed peaks in energy at low frequencies (fig. 4.4). Bottom salinities were directly influenced by freshwater discharge as seen in squared coherence calculated from the same data set (fig. 4.5). Salinity along both shores were strongly coherent with discharge in a range of periods from 200-1000 hrs. They were also 180° out of phase, indicating that high volumes of discharge decreased bottom salinities. Similarly, when discharge was low, bottom salinities increased.

Quantification of the relationship between freshwater discharge from Kinston and bottom salinity was sought and lagged correlation between the two conducted. A sample data set of January through September 2000 was used to compare freshwater discharge and salinity. These dates were chosen as more representative of average conditions in the NRE since the fall of 1999 witnessed the worst flooding in 100 years associated with hurricane Floyd. Lagged correlations were computed between discharge and bottom salinity at New Bern, Marker 11, 95 S, and 95 N (see fig. 1.1 for locations). The lag times associated with the largest correlation value grew progressively downstream (fig. 4.6). At New Bern, bottom salinity and discharge were negatively correlated at a value of -0.48 and a lag of 18.4 days whereby discharge led bottom salinity. This linear relationship explained 23% of the variation in bottom salinity and indicated that with higher discharge, bottom salinity at this location decreased.

Moving downstream, bottom salinity at US Coast Guard Marker 11 was negatively correlated with discharge at a correlation coefficient of -0.62 and a lag of 29.5 days with discharge leading bottom salinity. 38% of the variability in bottom salinity was explained through this relationship and as flows increased, bottom salinities decreased. At 95 S, discharge and bottom salinity were again negatively correlated with a correlation coefficient of -0.64 and a lag of 31.4 days with discharge leading bottom salinity. Almost 41% of the variability in bottom salinity at this site was explained by this relationship.
with discharge. Across the estuary at 95 N, discharge and bottom salinity were also negatively correlated with a correlation coefficient of -0.62 and a lag of 33 days.

Lateral variability in bottom salinity was evident from the lagged correlations between discharge and salinity at 95 N and S. A lag of 1.6 days existed between the lagged relationship of discharge and salinity. Between bottom salinities, however, a positive correlation existed at 0.93 and the south shore led the north shore by 1.5 days. These relationships suggested that it took approximately 31 days for discharge from Kinston to decrease near bottom salinity at 95S. There was also the suggestion of discharge controlling the lateral variability in bottom salinity as fresher, lighter water was deflected along the southern shore due to the coriolis effect through the constriction in the upper NRE. As the bathymetry opened, south of New Bern, the fresher water traveled along the southern shore in a “plug-type” flow and spread downstream and northward as it reached the north shore.

4.1.2 Measured Wind Influence

Low Frequency Variability

Typical low frequency wind-driven variability was evident in salinity and water level data presented in a 20-day record of both unfiltered and 30 hr low-pass filtered data from the summer of 2000 (figure 4.7). Synoptic band winds blowing toward the southwest pushed water from Pamlico Sound into the NRE as noted by higher filtered water levels (JD 181-185, 187-190, and 193-195, figure 4.7d).
Figure 4.5: Squared coherence and phase lag between discharge and bottom salinity data from June, 1999 through September, 2000.

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Table 4.1: Bottom Salinity means from CTD deployments from June 1999-December 2000
Figure 4.6: Lagged correlation between discharge from Kinston and bottom salinities from four locations in the NRE (see figure 1.1 for locations).
Likewise, synoptic band winds blowing towards the northeast pushed water out of the NRE into Pamlico Sound and lowered water level (JD 175-181, 185-187, 190-193, figure 4.7d). Lateral variability of water level was below the precision of the bottom-mounted CTDs and data from both instruments were therefore averaged for presentation.

Unfiltered and filtered salinity data (fig. 4.7b and c) showed wind-driven lateral advection of saline waters as winds blowing southwestward moved fresher water towards the south shore and reduced south shore bottom salinity below that of the north shore (fig. 4.7c). Winds blowing northeastward moved fresher water towards the north shore and reduced bottom salinity along the north shore below that of the south shore (fig. 4.7c).

Power spectra as a function of wind direction for an 84-day record spanning summer 2000 was contoured in figure 4.8a. Energy was concentrated around three frequencies and in a band of wind between 40° and 100° CCW from E or in the across channel direction. Energy peaks were at low frequencies (<0.01 cph or more than 4 days), at approximately 0.025 cph or 40 hours and at 0.042 cph or approximately 24 hours. Power spectra of water level and salinity (figure 4.9e-d) showed responses at similar frequencies. Energy peaked at lower frequencies (<0.01 cph) and again near 0.025 cph and 0.042 cph.

Two variable coherence tests were conducted between wind and salinity for the same 84-day summer 2000 period to determine the angular direction of wind that most influenced salinity variability. Contours of coherence squared greater than 95% significance level (> 0.3) are presented similar to that for water level done in the Material and Methods section (figure 4.8). Strong coherence between wind and salinity existed with across channel winds, defined in the Materials and Methods section as NNE/SSW winds, at a range of frequencies. The squared coherence between bottom salinity along the north and south shores was also computed to aid in understanding the relationship between wind and salinity (figure 4.9e,f).

At very low frequencies, on the order of a week or more (<0.005 cph), salinity was strongly coherent with NE/SW wind (figure 4.8). Bottom salinities were also strongly coherent and in phase (figure 4.9e,f). The combination of this information showed that salinities along the shores vary uniformly as the salt wedge was pushed along the upper NRE by these NE/SW winds.

Within the synoptic time band (2-5 days; 0.008<cph<0.021), bottom salinity was coherent with wind at the upper end of this range (0.02 cph) and with wind rotated into the across channel orientation. Bottom salinities, themselves, were also coherent at this frequency and approximately 140° out of phase (figure 4.9e,f). As the phase was near 180°, this suggested that salinities were responding to both an along and across channel process. Across channel winds blowing towards the S/ SW pushed the salt wedge up the estuary increasing bottom salinities while concurrently advecting fresher surface water towards the southern shore where it was downwelled. Likewise, positive across channel winds pushed the salt wedge out of the estuary while downwelling fresher water along the north shore.
Higher frequency variability was evident in the power spectra of both water level and salinity (figure 4.9a-d), especially at 0.035 cph or approximately 28 hrs. Bottom salinities were coherent at this frequency and range from 90° to 180° out of phase, however, there was no significant coherence relationship with the wind (figure 4.8b).

The predominance of low-frequency wind-driven, across-channel advection and mixing was clearly illustrated in time series salinity values along the north and south shores from November, 1999 (figure 4.10). This represented a recovery period from the effects of hurricane Floyd, which passed through eastern North Carolina on September 15 bringing large quantities of rain and flooding that pushed the salt wedge completely out of the estuary and into Pamlico Sound. Approximately two months later, saline water entered the upper estuary along the channel as noted by weekly mid-channel sampling trips (November 4, 1999 or JD 307 and November 9, 1999 or JD 312; figure 4.11). Moderate south-southwestward winds began near JD 310 and caused upwelling of saltier water along the north shore. The wind speed dropped for half a day and then picked up again northeastward. The switch in the wind caused saltier water to be upwelled along the south shore. As the wind regime continued to shift from south-southwestward to north-northeastward, a corresponding pattern of up and downwelling developed. Evidence of wind-driven mixing was also apparent from this record at approximately JD 318. Here, winds were south-southeastward and initially created an upwelling of saltier water along the north shore. As the wind forcing continued over a period of 2-3 days, however, the clear evidence of upwelling diminished and instead, salinity values at both the north and south shore equilibrated at approximately 2 psu.

**Diurnal Frequency Response**

Diurnal variability in salinity and wind was clearly evident in the 20-day record from the summer of 2000 (figure 4.7). Coherence between across channel wind and bottom salinities showed strong coherence at a diurnal frequency (0.04 cph, figure 4.8b). Bottom salinities were also coherent and near 180° out of phase (figure 4.9e,f). Nightly decreases in across channel wind speed (figure 4.12a) influenced the bottom salinity along the shores (figure 4.12b). Specifically, across channel winds towards the NNE moved freshwater towards the north shore where it was downwelled as indicated by decreases in bottom salinity. As wind speeds decreased, the advection and downwelling ceased and salinities rose to pre-advection levels (JD 180-184, 187-191, 193-195).

The episodic nature of this variability was quantified in the wavelet spectra of across channel wind (figure 4.12c) and north (figure 4.12d) and south (figure 4.12e) shore bottom salinity. Significant energy existed at roughly a diurnal period in all the records. Peaks in the wavelet energy of the north shore bottom salinity occurred at the diurnal period when across channel wind was directed towards the north. Similarly, along the south shore, diurnal peaks in the wavelet energy of south shore bottom salinity occurred with winds blowing towards the south. Since bottom salinities were coherent and
Figure 4.7: Wind vectors from CPMCAS (a), unfiltered bottom salinity (b), 30-hr low pass filtered bottom salinity (c), and filtered and unfiltered water level (d) from the summer of 2000.

Figure 4.8: Power spectra contours of a wind component rotated by 5° increments (a) and coherence contours between wind and bottom salinity (b) for an 84-day record during the summer of 2000.
Figure 4.9: Power spectra of north shore water level (a), south shore water level (b), north shore bottom salinity (c), south shore bottom salinity (d), bottom salinity coherence (e) and bottom salinity phase (f) for an 84-day record during the summer of 2000.

Figure 4.10: Wind vectors from CPMCAS (a), bottom salinity from the north and south shores (b), a zoomed in portion of the wind record (c) and bottom salinity record (d) from November, 1999. Diamonds along the top of the vector plot represented missing wind data.
approximately 180° out of phase (figure 4.9e,f), this diurnal salinity variability was an across channel phenomena directly controlled by local wind forcing.

Evidence for a wind-generated baroclinic seiche at roughly a diurnal period was presented in the November, 1999 time series at approximately JD 311.5 (figure 4.11c,d). Prior to this period, winds were moderate towards the SW advecting saltier water towards the north shore. After a day, the wind dropped off completely and an oscillation in the north shore salinity appeared until moderate winds began to blow towards the NE. The oscillation along the north shore was attributed to a wind-driven baroclinic seiche of approximately 16.5 hours. This observed period agreed with the theoretical baroclinic seiche period calculated by $T = 2L/MC$ with $C = \sqrt{g\frac{h_1 h_2}{h_1 + h_2}}$, the baroclinic wave speed, $g' = g\Delta \rho / \rho_0$, the reduced gravity, and $M$ as the mode number. Typical values for the NRE include the following: $L = 5500 m$, $\rho_0 = 1000$, $\Delta \rho = 4$, $h_1$, the surface layer depth, is $3H/4$ where $H$ is the water column depth of 4m, and $h_2$, the bottom layer depth, is $H/4$, yielding a period that agrees with the theoretical, but depends heavily on the layer depth. This was illustrated by figure 4.13 as the fundamental baroclinic wave speed varied between 10 and 50 hours depending on the degree of stratification ($\Delta \rho$) and the layer depth ($h_1$). During warmer months, a baroclinic seiche may have been masked by the fairly consistent diurnal variability in wind speed mentioned above.

Higher Frequency Response

A higher frequency response in both water level and vertically averaged across channel velocity was investigated using Fourier spectral analysis. Example spectra presented in figure 4.14 showed a
Figure 4.12: Across channel wind speed (a), bottom salinity from the north and south shores (b), wavelet spectra of across channel wind speed (c), wavelet spectra of bottom salinity along the north shore (d) and wavelet spectra of bottom salinity along the south shore (e) from the summer of 2000. The dashed lines along the edges of the wavelet graphs represent the cone of influence for which edge effects may become important in the analysis (Torrence and Compo, 1998)
Baroclinic wave speed as a function of $\Delta \rho$ and depth

Figure 4.13: The wave period of baroclinic seiches dependent on surface layer depth, $h_1$, and the degrees of stratification $\Delta \rho$.

clear peak at approximately 2 cph. Coherence tests between these variables showed significant coherence between 1.8 and 2.0 cph. Along both shores the water level and velocity were approximately 90° out of phase (figure 4.15). This relationship indicated the possibility of a barotropic surface seiche in the lateral direction. Usually referenced in context to a tidal wave trapped inside a bay, seiches or standing waves can be generated by the tides or wind surface stress. The period of the wave is calculated using Merian’s formula and assuming the simplest oscillation (the fundamental, $M = 1$). According to wave theory, the period of a surface seiche is equal to $T = \frac{2L}{C}$, where $L$ is the length of the body investigated, $M$ is the mode number and $C$ is the shallow water wave speed, $C = \sqrt{g\bar{h}}$ which is dependent on the height ($\bar{h}$) of the water column. Using typical values for the NRE defined above, the period of a first mode barotropic surface seiche would be approximately 30 min. This suggests that a wind-driven, lateral, barotropic seiche with a period of approximately 30 min occurs in this section of the estuary.

Contours of salinity along the north and south shore are presented from a 30 day period of the summer of 2000 (figure 4.16c,e). Generally, winds blowing towards the NNE moved fresher surface water towards the north shore where it was downwelled. These winds moved more saline mid-channel bottom waters towards the south shore where it was upwelled. Significant movement of saline waters sometimes occurred within an hour of wind onset. In extreme cases, high salinity water was upwelled throughout the water column (figure 4.16c JD 181-184, 187-190, and 201-205 and figure 4.16c JD 176-180, 184-187, and 199-201).

Contours of dissolved oxygen along the north and south shores are presented for the same time
period (figure 4.16b,d). Low DO water was present at the same time and to the same vertical extent as high salinity waters. Therefore, the wind driven movement and upwelling of high salinity water coincided with that of low dissolved oxygen water. In this case, across channel winds were directed towards the N/NE and low DO/high salinity water was upwelled along the southern shore (ie JD 175-180, 184-186, 192-193, 199-201). Conversely, higher DO/lower salinity water was downwelled along the northern shore. The situation reversed with across channel winds towards the SSW. This phenomenon was highly sensitive to changes in wind speed and direction as seen in JD 175-180. At that time, diurnal variability in the wind field was reflected in the variability in up/downwelling. As wind speed decreased, the downwelling along the north shore stopped and switched to upwelling as the change in surface stress initiated a baroclinic seiche.

Salinity/Dissolved oxygen relationship

There was a strong relationship between high salinity and low DO. This relationship was different than the Δ psu and bottom DO relationship mentioned in the Climatology section because the vertical extent of low DO water was considered in this case. The data recovered from the P-AVPs aided in determining this relationship. Correlation between salinity and DO were computed and the coefficients of determination (R) and correlation coefficient (R²) were calculated and are presented in table 4.2. Along the north shore, there was negative correlation between salinity and DO at the 2.0 m and 3.0 m levels. Variations in salinity explained approximately 20% and 40%, respectively, of the variability in DO at these levels. Along the south shore, negative relationships existed between salinity and DO at the 1.0 m, 2.0 m, and 3.0 m levels. Here, variations in salinity explained approximately 38%, 45%, and 42%, respectively, of the variability in DO at these levels.

Low DO water occurred at the same time that higher salinity water appeared, hence an upwelling of higher salinity/low DO water (figure 4.16). The LSI, as defined in the Methods section, was used to quantify the occurrence of upwelling and the occurrence of hypoxia (DO < 4 mg l⁻¹). This hypoxia definition was chosen to correspond to the thresholds that induce fish emigration (Tyson and Pearson, 1991) and invertebrate stress (Diaz and Rosenberg, 1995) and to prior studies done on the NRE (Buzzelli et al., 2002). The time series of the LSI and the occurrence of hypoxia were graphed with the across channel wind speed (fig. 4.17). It was evident that across channel winds blowing towards the north shore (>0) generated an upwelling of saltier water along the south shore (LSI < -2 at the 2.0 m level; figure 4.17c) and hypoxia was also evident. Likewise, across channel winds blowing towards the south shore (>0) generated upwelling of saltier water along the north shore with concurrent hypoxia (LSI > 2 at the 2.0 m level; figure 4.17c).

Along the shores of the estuary, at these locations, a strong relationship existed between low dissolved oxygen and higher salinity water. Upwelling of hypoxic water was clearly connected to upwelling of saltier water and likewise, downwelling of higher oxygen water was connected with downwelling of
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Table 4.2: Coefficient of determination and correlation coefficients for correlation between salinity and dissolved oxygen along the north and south shores at three different vertical levels.

Figure 4.14: Example spectra of vertically averaged across-channel velocity (a) and water level (b) from a 30-day period beginning July 1999.

fresher water. This upwelling appeared to be controlled by across channel winds as described above. Because of this observation, a linear model relating across channel wind speed to LSI, the upwelling index, was constructed. Across channel wind speed explained 39% of the variability in the near surface salinity upwelling, 54% in the middle salinity upwelling and 59% in the near bottom salinity upwelling. The model equations for the three levels are $LSI_1 = -0.55x + 0.62$, $LSI_2 = -0.68x + 0.8$, and $LSI_3 = -0.7x + 0.42$. Figure 4.18 shows the results of applying this linear model to the across channel wind data from the summer of 2000. The model performed extremely well in reproducing the upwelling phenomena. It also captured the physical effect of sudden transitions in wind speed and direction. This elegant, simple model could be applied as a quick first check to determine the likelihood of upwelling phenomena in the temporal vicinity of fish kills.
Wind-driven Advection of Salinity and DO: an example

A clear example of the above-mentioned characteristics for the upper NRE is presented during a 3-day period that began on August 25, 2000 (figure 4.19). Light, southward wind from JD 237.5-238.5 created a variable flow pattern of southward surface flow and northward bottom flow near both the north and south shores (figure 4.19b,f). Saltier water was advected along the bottom and up along the north shore as the location of the halocline moved upward in the water column (figure 4.19c). Near the south shore, fresher water was advected along the surface and down the shore as the location of the halocline moved down in the water column (figure 4.19g). Profiles of DO showed a similar trend in the up and downwelling of low DO and subsequent location of the oxycline (figures 4.19d,h). The sensitivity of the upwelling event to wind speed was illustrated as the wind speed decreased during the night (ie JD 238), allowing the equilibration of both the halocline and oxycline as they move vertically downward along the north shore and upward along the south shore.

As the wind shifted to northwestward, the water velocity reversed causing the downwelling pattern to reverse in both salinity and DO. A strong northward wind at approximately JD 239.5 pushed well-oxygenated, fresher water down vertically along the north shore and anoxic, saltier water up vertically along the south shore so that the whole water column was anoxic. This condition persisted for almost 3 hours (figure 4.19e). When the winds ceased for almost a quarter of the day, the beginning a baroclinic seiche was evident as both the halocline and oxycline moved up along the north shore while simultaneously moving down along the south shore. The movement then reversed as westward winds
Figure 4.16: Wind vectors from CPMCAS (a), contours of dissolved oxygen along the north (b) and south (d) shores, and contours of salinity along the north (c) and south (e) shores for the summer of 2000.
Figure 4.17: The across channel wind component (a), the lateral salinity index (LSI) at 1.0m (b), 2.0m (c), and 3.0m (d). Hypoxia along the shores is noted by dots above and below the time series. LSI values outside the gray boxes indicate tilting or upwelling.
Figure 4.18: Across channel wind speed (a) was used as a linear predictor for the LSI at the 1.0m, 2.0m and 3.0m depths (b,c,d). The recorded data is in black while the model data is in grey.
Spatial Distribution of Salinity and DO

A snapshot of the estuary is seen in figures 4.20 and 4.21 over a 6 hr time span. In this case, across channel winds were blowing towards the north shore. Fresher water was being advected towards the north shore as salinity contours along the north shore were approximately 8 psu whereas salinity contours along the south shore were approximately 13 psu. Along the north shore, fresh water was also being downwelled as bottom salinities became 8 or 9 psu. Along the south shore, salinities were higher as the salt wedge was being pushed towards the south shore. Lateral variability in bottom and surface salinity was due to the combination of the upstream extent of the salt wedge and how surface stress due to wind moved it laterally. Spatial quantification of the areas affected by up/downwelling was necessary to determine if physical circulation was a cause of fish kills.

The bottom-mounted CTDs were also deployed during this collection period. The salinity record confirmed observations during the high spatial collection period (figure 4.22). At approximately JD 187.6, across channel wind speed increased to near 10 m s⁻¹. Bottom salinity along the north shore responded by decreasing almost 8 psu as this wind created an extreme downwelling event. At the same time, there was a slight increase in bottom salinity along the south shore.

4.2 Numerical Simulation of Lateral Variability

4.2.1 EFDC Model Calibration

The 3-D EFDC model was calibrated in the along channel direction to water quality data collected by IMS as part of the MODMON project. Side channels were calibrated with data from NCSU collected during 1998 and 1999 (US EPA Region 4, 2002).

Salinity was calibrated with the following two objectives: 1) establish the model's ability to predict stratification and destratification by comparing model data with data from the USGS Marker 11; and 2) establish the model's ability to accurately predict salinity distribution in the longitudinal and lateral direction by comparing model output with MODMON and NCSU data. Statistical comparison between the model salinity and observed salinity for the MODMON and NCSU data are presented in table 4.3.

4.2.2 EFDC Model Validation

The 3-D EFDC model did an good job of reproducing bottom salinity along both shores during the summer of 1999 (figure 4.23). Basic statistics comparing the model and field data were calculated and are presented in table 4.4. The ME for both locations were close to 0 as is the RMSE. The $R^2$ statistic represented the coefficient of determination of a linear regression between field and model data.
Figure 4.19: Wind vectors from CPMCAS (a), across-channel water velocity contours (b), salinity profile contours (c), DO profile contours (d), and bottom salinity (e) along the north shore and across-channel water velocity contours (f), salinity profile contours (g), DO profile contours (h), and bottom salinity (i) along the south shore during August, 2000. Diamonds along the top of the vector plot represent missing wind data.
Figure 4.20: Surface salinity contours from July 7, 1999.

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Table 4.3: Calibration statistics between predicted and observed data for the EFDC model year 1998.
Figure 4.21: Bottom salinity contours from July 7, 1999.
Figure 4.22: Across channel wind speed (a) and bottom salinity data from July 6-8, 1999. The shaded region represents the time period for the spatially intensive field experiment on July 7.
<table>
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<th>Model Mean</th>
<th>Field Std</th>
<th>Model Std</th>
<th>ME</th>
<th>RMSE</th>
<th>R²</th>
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</table>

Table 4.4: Comparative statistics between bottom salinities of the model output and recorded field data during the summer 1999.

(figure 4.24). The model did a good job of simulating field conditions without over or underestimating. The model data represented over 80% of the variability in the field data.

Statistical comparisons between model and ADCP recorded flow velocities were also computed. Both the ADCP and the simulated data represented vertical bins in the water column. Simulated flow represented 4 bins in the water column and the ADCP was able to capture at least 15 bins in the water column. The bins of both sets of data were uniformly spaced with simulated bins being larger than observed bins. Representative time series of ADCP and model data comparisons were presented (figures 4.25 and 4.26) and the statistics are summarized in table 4.5. Visual inspection of the data showed good correlation between the model and recorded data. The bottom bin comparison was not very good and this may be due to the fact that the ADCP was raised approximately 65 cm from the bottom and did not capture the flow near the bottom.

A separate validation of the spatial performance of the model was done by comparing the surface and bottom salinity contours from the high frequency, high spatial sampling conducted in the summer of 1999. Figure 4.27 shows how across channel winds blowing towards the north shore advect fresher surface salinity towards the north shore. Model salinity values were similar to those recorded during the experiment (figure 4.20). Likewise, figure 4.28 indicates that this fresher water was downwelled to the bottom as bottom salinities were approximate to surface salinities along the north shore. Similarly, salinity values along the south shore increased at both the surface and the bottom indicating the advection and upwelling of high salinity water. Again, this was consistent with what was recorded during the spatial experiment (figure 4.21).

4.2.3 Model Experimentation

All three test cases with the coriolis force (a) discharge only; b) wind only; and c) discharge and wind) produced varying degrees of lateral salinity variations. Surface and bottom salinity contours for the test cases are presented in figures 4.29 and 4.30, respectively. The most pronounced lateral variability in surface salinity derived from the discharge only boundary condition case (figure 4.29a). In that case, fresher water flowed down the upper NRE along the southern shore and into the lower NRE. Saltier water pushed up the estuary main channel along the bottom (figure 4.30a) in response to the
Figure 4.23: Model salinity data (black) and recorded salinity data (grey) for the summer 1999.

Figure 4.24: Linear regression between model salinity data and field salinity data for the summer 1999.
Figure 4.25: Time series comparison of model (black) and field (grey) velocity in the east direction at 4 different vertical levels for the summer 1999.

Figure 4.26: Time series comparison of model (black) and field (grey) velocity in the north direction at 4 different vertical levels for the summer 1999.
Table 4.5: Comparative statistics between model velocities and observed data for the summer 1999 period.

<table>
<thead>
<tr>
<th>Location</th>
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<th>Dir</th>
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<th>Model Mean</th>
<th>Field Std</th>
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<th>ME</th>
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<td>-1.36</td>
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</table>

Figure 4.27: Model surface salinity calculated for JD 187 of 1999.
Figure 4.28: Model bottom salinity calculated for JD 187 of 1999.

Baroclinic pressure gradient. There were indications of lateral variability along the bottom salinity with saltier water along the northern shore (figure 4.30a).

A less pronounced effect occurred with the discharge and wind boundary condition case (figures 4.29c and 4.30c). In that case, fresher water was prevalent along the southern shore in both the upper NRE and lower NRE. Wind induced more vertical mixing than was present in the discharge only case and the upper NRE was less salty on the bottom, yet more salty on the surface (figure 4.29c). Lateral salinity variability was present as saltier water entered the upper NRE along the north shore (figure 4.30c).

The wind only boundary condition resulted in fresher water along the southern shore in both the surface (figure 4.29b) and bottom (figure 4.30b). Winds during this time were directed predominantly towards the SW/S/SE and accounted for the saltier water along the north shore and fresher water along the south shore (figure 4.31). Saltier water entered the upper NRE along the north shore (figure 4.30b). This tendency reversed when winds were blowing toward the NE/N/NW.

It appeared that the Coriolis force played a role in controlling lateral variability as seen in the discharge only and no coriolis force example (figure 4.29d). In that case, discharge flowed down the center of the estuary and did not favor a particular side. Likewise, higher salinity water appeared to move upstream along the deeper central portion of the NRE and did not favor a shore as in previous cases (figure 4.30d).
Figure 4.29: Model surface salinity contours after 25 days from experimental runs: a) Discharge only boundary condition; b) Wind only boundary condition; c) Discharge and wind boundary condition; and d) Discharge only and no coriolis.
Figure 4.30: Model bottom salinity contours after 25 days from experimental runs: a) Discharge only boundary condition; b) Wind only boundary condition; c) Discharge and wind boundary condition; and d) discharge only and no coriolis.
Figure 4.31: Discharge conditions (a), wind speed (b) and wind direction (c) for the first 50 days of the year 2000 were used to drive the experimental model.

4.2.4 Model Application

After the model was validated, it was used to simulate hydrodynamic and transport conditions in the NRE for the year 2000. This simulation was computed in order to both compare with observational data and to fill in the gaps where data was unavailable and fish kills occurred. Simulated salinity was statistically compared to data collected by the USGS at Marker 11. Salinity from the near surface and near bottom monitors were compared with model output from levels $\sigma_1$ and $\sigma_7$. This served to check the fidelity of the model as well as the stratification reproduction performance of the model. The ME between surface comparisons was 0.159 and -1.024 in the bottom. The RMSE in the surface was 1.371 and 2.983 in the bottom. Coefficients of determination between the model and field data were 0.89 at the surface and 0.87 near the bottom. So, the model explained roughly 86% of the variability seen in the field data. Time series graphs of the near surface and near bottom salinity are presented in figure 4.32.

Bottom salinity data from the lateral study were also compared with model output. Time series from the north and south shore are presented during the summer of 2000 (figure 4.33). Along the North shore, statistics between the model and observed data yielded a ME of 0.094, RMSE of 1.81 and $R^2$ of 0.80 whereas along the south shore there was a ME of 0.068, RMSE of 2.17 and $R^2$ of 0.78. So model simulations explained more than 78% of the variability in field data. These results were comparable to simulation year 1999.
The model reproduced peaks in the power spectra of the bottom salinity at key frequencies noted in the observed data (figure 4.34). Both the observed and model salinity data had peaks in energy at low frequencies (<0.02 cph). A clear diurnal signal was present in both the observed and model power spectra density plots. Higher frequency peaks were also synchronous. At those same frequencies, coherence between simulated and observed data was high and nearly in phase 4.35.

The ability of the model to capture the low frequency and higher variability recorded in the observed bottom salinity data was clear in a twenty-day data set (figure 4.36). Here, filtered model salinity showed the same trends as the filtered observed salinity. Specifically, winds blowing toward the NE reduced bottom salinities along the north shore and increased salinity along the south shore. The situation reversed with winds blowing toward the SW (figure 4.36 d and e). The model also responded to decreases in the wind speed as seen in the unfiltered data (figure 4.36 b and c).

Vertical resolution of salinity variability compared well with the data collected from the autonomous vertical profiling system (figure 4.37). The model did an excellent job of replicating conditions during the data collection period and also filled in information where there were gaps in the field data. For example, the autonomous profiling system did not collect data from approximately JD 193-195.5. During this time, winds were blowing strongly towards the S/SW and then tapered off. The model showed an upwelling of higher salinity water along the north shore that seemed to agree with what was
Figure 4.33: Time series of model and field near bottom salinity along the north (a) and south (b) shores for the summer of 2000.

Figure 4.34: Low frequency power spectral density plots of the field salinity along the north (a) and south (c) shores and of the simulated salinity along the north (b) and south (d) shores.
Figure 4.35: Coherence between simulated and field bottom salinity along the north (a) and south (c) shores and their phase (b and d).

hinted at by the observed data (figure 4.37 b and c). The LSI was also computed and compared well with the field LSI (figure 4.38). Because the model used the full 3-D momentum equations, it was more able to represent complex physical phenomena and provide considerable physical insight into causality. It was still evident, however, that wind played a critical role in upwelling and advection of salinity.
Figure 4.36: Wind vectors from CPMCAS (a), unfiltered observed bottom salinity (b), unfiltered simulated bottom salinity (c), 30-hr low pass filtered observed bottom salinity (d), and 30-hr low pass filtered simulated salinity (e) from the summer of 2000.
Figure 4.37: Wind vectors from CPMCAS (a), observed salinity along the north shore (b), simulated salinity along the north shore (c), observed salinity along the south shore (d), and simulated salinity along the south shore (e) during the summer of 2000.
Figure 4.38: Across channel wind speed (a), observed and simulated LSI at 1.0m (b), 2.0m (c), and 3.0m (d).
Chapter 5

Results: Dynamics

Results from the previous chapter documented many interesting phenomena within the upper NRE. Of particular importance is the lateral variability in the pycnocline and wind-driven upwelling of high salinity/low DO waters along the shores. Additionally, the longitudinal mean flow showed considerable lateral variability. Specific questions about the dynamics involved with these phenomena evolved, for example, what type of wind stress is required to ventilate the pycnocline and how does this vary with stratification? Furthermore, is the longitudinal mean flow interpretable solely based on gravitational circulation, or is it necessary to invoke a long term wind stress? This chapter seeks to determine the dynamics behind the observed phenomena and answer questions posed as a result of the previous chapter.

5.1 General characteristics of the upper NRE

In many respects, the upper NRE can be considered a tideless estuary with friction. Dyer (1997) explains that in a tideless estuary, river water would flow outward over the surface of the saline water. The velocity and thickness of the surface layer decrease towards the mouth as the estuary widens. The Coriolis force would cause river water to be concentrated on the right-hand side (looking downstream) in the northern hemisphere. Shear between the two layers occurs as friction is incorporated. The salt wedge would be pushed downstream until its upper surface has a slope sufficient to resist this force. The tip of the salt wedge will be blunted due to bottom friction effects. The Coriolis force will affect the lateral water slopes, with the interface sloping downwards towards the right and the sea surface sloping downwards towards the left in the northern hemisphere.

This simplified view of an estuary is complicated by the orientation and bathymetry of the system. The NRE contains a near 90° bend in the middle of the estuary. This bend may separate the NRE into two dynamically coupled systems: the upper NRE which is directly influenced by the main freshwater source at New Bern and its connection with the lower NRE and the lower NRE whose dynamics are influenced more by its connection with Pamlico Sound and, to a lesser degree, with the upper NRE (Pietrafesa et al., 1986; Luetich et al., 2002).

Normally, tides represent a deterministic mechanism for oscillating volumes of water within the estuary at well-known, regular intervals and are a source for internal mixing. The NRE, however is
micro-tidal. The mouth of the estuary empties into Pamlico Sound which itself is separated from the Atlantic by a chain of barrier islands. Since tidal influence is small, the main control for the dynamics of the NRE in addition to freshwater discharge mentioned above is meteorological variation. Specifically, a NE/SW wind pattern is prevalent in the NRE system. This wind pattern is aligned in an across channel manner in the upper NRE and an along channel manner in the lower NRE. The influence of freshwater discharge and wind affect the upper NRE at different time scales. At long time scales, perhaps most noticeably at a seasonal time scale, freshwater discharge plays a major role in controlling circulation and salt wedge location. At shorter time scales, the prevailing winds may generate lateral variability in the upper NRE while at seasonal time scales winds may also control the salt wedge position.

5.2 Local Mixing vs. Advection

The competition between stratification and mixing plays a crucial part in estuarine dynamics because when the fluid is stratified, the density gradient resists the exchange of momentum by the turbulence and an extra velocity shear is necessary to cause mixing (Dyer, 1997). The Gradient Richardson number, Ri is a comparison of the stabilizing forces of the density stratification to the destabilizing influences of velocity shear and is defined by

\[ Ri = \frac{g \rho}{\rho} \left[ \left( \frac{\partial u}{\partial z} \right)^2 + \left( \frac{\partial v}{\partial z} \right)^2 \right] \]

For \( Ri > 0 \) the stratification is stable, for \( Ri = 0 \) it is neutral and the fluid is unstratified, and for \( Ri < 0 \) it is unstable. When the stratification exceeds a certain value turbulence will be damped, mixing will be limited and the flow will be essentially laminar. There have been many laboratory and theoretical investigations of the mechanisms of formation and growth of instabilities within the stratified interface, and the transition from laminar to turbulent flow under conditions of uniform flow is generally taken to occur at \( Ri = 0.25 \) (Dyer, 1997).

In essentially two layered systems, the gradient Richardson number may be simplified by a type of layer or bulk Richardson number (Dyer, 1997). Goodrich, et al (1987) used data from a near surface and near bottom mooring along the main stem of the Chesapeake Bay to compute a bulk Richardson number. A time series of this bulk Richardson number was then used to determine when wind-driven destratification events occurred along the main stem of the Bay.

The NRE may be considered a two-layer system as seen from representative density profiles in figure 5.1. Here, salinity, temperature and depth data from the profilers were used to compute instantaneous density profiles along the north and south shores. The upper layer of the NRE extends to between -1.5m and -2.25m with \( \sigma_t = 4 \) or 5. A corresponding lower layer has a \( \sigma_t = 8 - 10 \). The bulk Richardson number defined as

\[ Ri_B = \frac{g \sigma_t}{\rho} \left[ \left( \frac{\Delta u}{\Delta z} \right)^2 + \left( \frac{\Delta v}{\Delta z} \right)^2 \right] \]
could be computed for the north shore as velocity data was only available from the north shore for 23 days during the summer of 2000.

Time series data from the 23 day record from the summer of 2000 of $\Delta S$, $\partial \rho / \partial z$, the arctangent of the bulk Richardson number, across channel wind speed, across channel wind stress, and shear are presented in figure 5.2 for the north shore site. Data gaps less than half a day were linearly interpolated over time whereas they were omitted if they were longer than half a day. The along and across channel wind speed as defined in the Methods chapter were determined and then used to compute wind stress. Wind stress calculations were based on a drag coefficient reduced to 10 m height and neutral conditions (Large and Pond, 1981). Along the shores, salinity stratification is a clear indicator of density stratification (figure 5.2a,b). This is consistent along the south shore as well although the data are not presented. There was no obvious correspondence between peaks in the shear stress and either the across
or along channel wind stress (figure 5.2d,e,f and 5.3d,e,f) indicating that the velocity shear stress was not directly attributable to wind stress.

The arctangent of the Richardson number was graphed for visualization purposes and the 0.25 threshold was marked with a horizontal line. The behavior of the bulk Ri indicated a near binary state where the water column was either strongly stratified and stable or well-mixed and unstable (figure 5.2c). The bulk Ri dropped below 0.25 as stratification dropped to near 0. The transition was almost instantaneous from stratified to well-mixed. Rarely did low bulk Ri numbers correspond to periods of large velocity shear (figure 5.2c,f). Low bulk Ri did, however, correspond to peaks in the wind stress, corresponding wind speeds greater than or equal to 5 m s⁻¹, oriented towards the north shore (figure 5.2d,e). The “sudden” jumps from stratified to non-stratified conditions appear to be the result of advective properties due to upwelling rather than mixing. There was no apparent correspondence with the bulk Ri and along channel winds (figure 5.3c,e).

5.3 Lateral Dynamics

The results in Chapter 4 and the above observations suggest that the nearshore stratification is strongly influenced by across channel winds. A lateral section of the upper NRE can be said to behave like a two layer lake system with movement directed only in the lateral direction (see figure 5.4 for a representation of a cross section of the upper NRE).

Following work done on the Windermere system (Heaps and Ramsbottom, 1966), a momentum balance can be constructed considering a lateral section of the upper NRE. While in equilibrium, this section would consist of two horizontal layers of water, each homogeneous, with different densities. The length of this cross section in the upper NRE is approximately 5.5 km and the depth is 4.0 m and therefore, the depth is small in comparison with the length. Assuming that (i) lateral water movement is due to surface wind stress, (ii) non-linear values and vertical accelerations are negligible, (iii) density and eddy viscosity are constant within each layer, but may differ between layers, and (iv) internal friction at the interface between the two layers may be considered negligible as advective processes dominate over local mixing processes, the continuous hydrodynamic equations governing the motion of water of uniform constant density $\rho$ reduce to

$$\frac{\partial w}{\partial y} + \frac{\partial w}{\partial z} = 0 \quad (5.1)$$

$$\frac{\partial w}{\partial t} = -\frac{1}{\rho} \left( \frac{\partial p}{\partial y} + \frac{\partial F_x}{\partial z} \right) \quad (5.2)$$

$$0 = -\frac{1}{\rho} \frac{\partial p}{\partial z} + g \quad (5.3)$$
Figure 5.2: Salinity stratification (a), density stratification (b), the arctangent of the bulk Richardson number (c), across channel wind speed (d), across channel wind stress (e), and the squared velocity shear (f) at the north shore site during the summer of 2000.
Figure 5.3: Salinity stratification (a), density stratification (b), the arctangent of the bulk Richardson number (c), along channel wind speed (d), along channel wind stress (e), and the squared velocity shear (f) at the north shore site during the summer of 2000.
where \( v \) is the component of current in the \( y \) direction and \( w \) is the component of current in the \( z \) direction, positive downward; \( p \) is the pressure at a point in the water; \( F_{x y} \) is the frictional stress in the \( y \) direction; \( g \) is the acceleration due to the earth’s gravity; and \( t \) is time.

When the upper NRE is at rest, a homogeneous layer of density \( \rho_1 \) lies to a depth \( h_1 \) over another homogeneous layer of density \( \rho_2 \) and depth \( h_2 \). The total depth is \( h = h_1 + h_2 \). When there is motion, the elevation of the upper surface is \( \zeta_1 \), and that of the surface between layers is \( \zeta_2 \) as shown in figure 5.4. Dividing the continuous hydrodynamic equations 5.1, 5.2, and 5.3 into two separate layers where

\[
v = v_1, \ w = w_1, \ p = p_1, \ F_{xy} = F_1, \quad \text{in the upper layer } (-\zeta_1 \leq z \leq h_1 - \zeta_1),
\]

\[
v = v_2, \ w = w_2, \ p = p_2, \ F_{xy} = F_2, \quad \text{in the lower layer } (h_1 - \zeta_2 \leq z \leq h),
\]

then

\[
\frac{\partial v_1}{\partial y} + \frac{\partial w_1}{\partial z} = 0 \tag{5.4}
\]

\[
\frac{\partial v_1}{\partial t} = -\frac{1}{\rho_1} \left( \frac{\partial p_1}{\partial y} + \frac{\partial F_1}{\partial z} \right) \tag{5.5}
\]

\[
\frac{\partial p_1}{\partial z} = \rho_1 g \tag{5.6}
\]

for the upper layer and
\[
\frac{\partial v_2}{\partial y} + \frac{\partial w_2}{\partial z} = 0 \quad (5.7)
\]

\[
\frac{\partial v_2}{\partial t} = -\frac{1}{\rho_2} \left( \frac{\partial p_2}{\partial y} + \frac{\partial F_2}{\partial z} \right) \quad (5.8)
\]

\[
\frac{\partial p_2}{\partial z} = \rho_2 g \quad (5.9)
\]

for the lower layer.

The \(z\)-momentum equations in each layer (equations 5.6 and 5.9) may be integrated to solve for pressure in both layers subject to the boundary conditions, \(p_1 = p_a\) when \(z = -\zeta_1\) and \(p_1 = p_2\) when \(z = h_1 - \zeta_2\) where \(p_a\) is atmospheric pressure that is assumed constant and uniform,

\[
p_1 = p_a + g \rho_1 (\zeta_1 + z),
\]

\[
p_2 = p_a + g \rho_1 (\zeta_1 + h_1 - \zeta_2) + g \rho_2 (\zeta_2 + z - h_1).
\]

The pressure terms may be substituted into equations 5.5 and 5.8 to get

\[
\frac{\partial v_1}{\partial t} = -g \frac{\partial \zeta_1}{\partial y} - \frac{1}{\rho_1} \frac{\partial F_1}{\partial z} \quad (5.10)
\]

\[
\frac{\partial v_2}{\partial t} = -g \frac{\rho_1}{\rho_2} \frac{\partial \zeta_1}{\partial y} - g \left( 1 - \frac{\rho_1}{\rho_2} \right) \frac{\partial \zeta_2}{\partial y} - \frac{1}{\rho_2} \frac{\partial F_2}{\partial z} \quad (5.11)
\]

Vertically integrating equations 5.4 and 5.10 between \(z = 0\) and \(z = h_1\) and equations 5.7 and 5.11 between \(z = h_1\) and \(z = h\) with the boundary conditions of \(w_1(z = 0) = -\partial \zeta_1 / \partial t\), \(w_1(z = h_1) = w_2(z = h_1) = -\partial \zeta_2 / \partial t\), and \(w_2(z = h) = 0\) yields

\[
\frac{\partial}{\partial y} \int_0^{h_1} v_1 \, dz + \frac{\partial \zeta_1}{\partial t} - \frac{\partial \zeta_2}{\partial t} = 0, \quad (5.12)
\]

\[
\frac{\partial}{\partial t} \int_0^{h_1} v_1 \, dz = -gh_1 \frac{\partial \zeta_1}{\partial y} + \frac{1}{\rho_1} (F_S - F_D) \quad (5.13)
\]

\[
\frac{\partial}{\partial y} \int_{h_1}^{h} v_2 \, dz + \frac{\partial \zeta_2}{\partial t} = 0 \quad (5.14)
\]

\[
\frac{\partial}{\partial t} \int_{h_1}^{h} v_2 \, dz = -gh_2 \frac{\rho_1}{\rho_2} \frac{\partial \zeta_1}{\partial y} - gh_2 \left( 1 - \frac{\rho_1}{\rho_2} \right) \frac{\partial \zeta_2}{\partial y} + \frac{1}{\rho_2} (F_D - F_B) \quad (5.15)
\]

where \(F_S = F_{zy}(z = 0)\) = the component of wind stress over the surface of the cross section in the direction of \(y\), \(F_D = F_{zy}(z = h_1)\) = internal friction at the surface of discontinuity and \(F_B = F_{zy}(z = h)\) = bottom friction.

The depth mean values of \(v_1\) and \(v_2\) are

\[
v_{1m} = \frac{1}{h_1} \int_0^{h_1} v_1 \, dz, \quad v_{2m} = \frac{1}{h_2} \int_{h_1}^{h_2} v_2 \, dz,
\]

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and equations 5.12, 5.13, 5.14 and 5.15 may be written as

\[
\begin{align*}
\frac{h_1}{\partial y} + \frac{\partial \xi _1}{\partial y} - \frac{\partial \xi _2}{\partial y} &= 0, \\
\frac{\partial \xi _1}{\partial t} &= -g \frac{\partial \xi _1}{\partial y} + \frac{1}{\rho_1h_1}(F_S - F_D), \\
\frac{\partial v_{1m}}{\partial t} &= \frac{h_2}{\partial y} + \frac{\partial \xi _2}{\partial y} = 0, \\
\frac{\partial v_{2m}}{\partial t} &= -g \frac{\rho_1}{\rho_2} \frac{\partial \xi _1}{\partial y} - g \left(1 - \frac{\rho_1}{\rho_2}\right) \frac{\partial \xi _2}{\partial y} + \frac{1}{\rho_2h_2}(F_D - F_B).
\end{align*}
\] (5.16)

When there is no wind stress and no internal friction, then \( F_S = F_D = F_B = 0 \) and equations 5.16, 5.17, 5.18, and 5.19 reduce to the equations governing surface and internal seiches (Proudman, 1953). In this case, \( v_1 \) and \( v_2 \) are independent of depth, so \( v_1 = v_{1m} \) and \( v_2 = v_{2m} \). Periods of the internal seiches were investigated earlier in the Chapter 4, Results section and showed how baroclinic wave periods varied as a function of upper layer depth, \( h_1 \) and stratification. Elevations within the layers were not investigated, however.

Equations 5.16, 5.17, 5.18, and 5.19 can be solved for \( v_1, v_2, \xi _1, \) and \( \xi _2 \) with the following boundary conditions:

\[
\begin{align*}
v_1 &= v_2 = 0 \text{ at } y = 0, y = l, \\
F_1 = F_s, \text{ i.e. } -\rho_1N_1 \partial v_1/\partial z &= F_s, \text{ at } z = 0, \\
F_1 = F_2 = F_D = 0, \text{ i.e. } \partial v_1/\partial z = \partial v_2/\partial z = 0, \text{ at } z = h_1,
\end{align*}
\]

and an approximate form for the bottom friction (Proudman, 1953) is given by

\[
F_B/\rho_1 h_2 = 2k v_{2m} \text{ for } k \text{ a constant.}
\]

Solutions for \( \xi _1 \) and \( \xi _2 \) were sought when a sudden increase in the wind stress \( F_S \) at \( t = 0 \) is applied to the system and remains constant. Solutions of the form \( \xi _1 = Z_1(t) \cos(n\pi y/l) \) and \( \xi _2 = Z_2(t) \cos(n\pi y/l) \) were determined in order to satisfy boundary conditions where \( n \) is the mode number, \( y \) is the distance across channel and \( l \) is the length of the channel. Details of the solution are provided in (Heaps and Ramsbottom, 1966) and this allows for the quantification of wind stress required to create a pycnocline excursion as stratification varies.

The mean profile of density in the upper NRE (figure 5.1) separates the upper NRE into 2 layers, an upper layer that extends to 1.5m with a density of 1005 kg m\(^{-3}\) and a lower layer with a density of 1008 kg m\(^{-3}\). For the analysis, \( k = 1.1 \times 10^{-5} \) and the highest energy mode is viewed and hence, \( n = 1 \). Figure 5.5 provides elevation time series at two locations, a south shore site \( (y = 0.5\text{km}) \) and a north shore site \( (y = 5.0\text{km}) \), when wind stress is directed towards the north shore. Magnitudes of the elevations of the lower layer \( h_2 \) are much higher than those for the upper layer \( h_1 \). With a northward wind, upper layer elevations at the south shore site decrease while those at the north shore site increase.
within the mm range. In the bottom layer, elevations increase at the south shore site and decrease at the north shore site within the m range. The effects of \( \zeta_1 \) and \( \zeta_2 \) based on the wind stress from three wind speeds (2 ms\(^{-1}\), 5 ms\(^{-1}\) and 10 ms\(^{-1}\)) are presented (figure 5.5). The largest elevations in both layers occur naturally with the wind stress associated with the highest wind speed.

The effects of stratification on the elevation of the bottom layer were investigated. Here, it was assumed that the layer depths remained as defined in the above case, yet density in the bottom layer varied. Figure 5.6 represents the magnitude of the maximum elevation of the lower layer vs. stratification based on different wind speeds. Stratification is simply \( \rho_2 - \rho_1 \). First, higher winds speeds allow for larger pycnocline excursions. Not surprisingly, as stratification increases, the pycnocline excursions due to each wind speed decrease.

Ventilation of the pycnocline is important for biological reasons as it brings higher density, lower oxygen water from the lower layer depths into the upper layer regions. The wind speed required to elevate this pycnocline varies according to the stratification between the layers. The wind speed required to create a pycnocline excursion, dependent on stratification, was investigated in figure 5.7. A lower layer ventilation of 0.5m can occur with wind speeds as low as 2 ms\(^{-1}\) however, the minimum wind speed required increases as stratification increases. Even with strong stratification (\( \Delta \sigma_1 = 3 \)) a 5 ms\(^{-1}\) wind will produce a lower layer ventilation of 0.5m. For a pycnocline ventilation of 1.5m, typical for the upper NRE, and minor stratification (\( \Delta \rho = 2 \)) wind speeds of 12 ms\(^{-1}\) or greater are required. The required wind speed varies linearly as stratification increases. With minimal stratification, wind speeds of 8 ms\(^{-1}\) are required.

Following Heaps and Ramsbottom (1966), \( \zeta_2 \) was calculated for a time-varying wind stress based on time-varying wind speed. Since the wind speeds and direction vary over time, the above model was applied to the NRE during the profiler collection period. Figures 5.8 and 5.9 present a time series of \( \zeta_2 \) based on model calculations and field data for both the north and south shore locations. This linear model does a rough job of predicting pycnocline excursions in both the north and south shore. Similar trends are identified, if not fully realized at the same magnitudes. The pycnocline excursions noted by the model are, however, larger than those presented for wind speed vs. stratification (figure 5.7). This may be because the variable wind forcing sets up a series of baroclinic seiches on \( \zeta_2 \) that may be additive in nature. There is more variability in the field data and a more accurate model may be necessary to fully capture the pycnocline response. In fact, a similar analysis was completed with model results from EFDC and showed a better correlation with the field data.

5.3.1 Along channel wind vs. upwelling

The influence of along channel wind stress in causing across channel upwelling was shown to be relatively minor in comparison to the influence of across channel wind stress (figure 5.2 and 5.3). However, an Ekman layer transport relationship leading to upwelling along the shore to the left of the
Figure 5.5: Elevation time series for (a) the upper layer south shore site, (b) the upper layer north shore site, (c) the lower layer south shore site, and (d) the lower layer north shore site as computed from the dynamic model.

Figure 5.6: Maximum lower layer elevation magnitude vs. stratification for (a) a 2 ms$^{-1}$ wind, (b) a 5 ms$^{-1}$ wind, (c) a 7 ms$^{-1}$, and (d) a 10 ms$^{-1}$ wind as computed from the dynamic model.
Figure 5.7: Wind speed vs. stratification required to raise the lower layer 0.25m, 0.5m, 1.0m, and 1.5m.

Figure 5.8: Across channel wind speed (a), model and field $\zeta$ (b). The model assumes a lower layer thickness of $h_2 = 2.5\text{m}$ and $\Delta \rho = 2$ and the field data are from the summer of 2000.
Figure 5.9: Across channel wind speed (a), model and field $\zeta_2$ (b). The model assumes a lower layer thickness of $h_2 = 2.5$ m and $\Delta \rho = 2$ and the field data are from the summer of 2000.

The Ekman depth may be determined by $D_E = \sqrt{2A_v/f}$ where $f$ is the Coriolis parameter and $A_v$ is the Austausch coefficient or vertical eddy coefficient. $A_v$ is associated with the turbulent shear stresses and is challenging to measure accurately, but may be estimated using the $Ri$ number. In fact, a low-order turbulence closure model may be used so that $A_v = 0.01(1 + 5Ri)^{-2} + 10^{-4}$ (Pacanowski and Philander, 1981). This closure model was applied to the James River Estuary (Valle-Levinson et al., 2000) because the model was shown to perform best among the lower order schemes (Nunes-Vaz and Simpson, 1994). Using a mean bulk $Ri$ over the 30 day period during the summer of 2000, the eddy viscosity reverts to background levels of $10^{-4}$ and the Ekman layer depth was determined to be $D_E = 3.8$ m. In the James River estuary, values of $A_v$ varied between a minimum of $5 \times 10^{-4}$ m$^2$s$^{-1}$ and a maximum of $20 \times 10^{-4}$ m$^2$s$^{-1}$ (Valle-Levinson et al., 2000). Applying these values, $D_E$ ranged between 8.67 m and 17.26 m.

In the upper NRE, depth, $H$, varies from 2 to 4 m. It is approximately 3.5-4 m deep from 25 km downstream to the bend at 40 km downstream. If the minimum background eddy viscosity is assumed, $H < D_E$ in the region of interest. In general, if $H > 1.25D_E$, Ekman transport is to the right of the wind. As $H < D_E$, that transport is more in line with the direction that wind is blowing. Based on the comparison between the upper NRE water depth and the Ekman layer depth, it seems infeasible for Ekman type upwelling associated with along channel wind to occur.
The conclusion that Ekman type upwelling associated with along channel wind does not occur is also supported by the fact that Ekman layer dynamics depend on fairly regular along channel winds. The wind system in the NRE is dominated by NE/SW (across channel) winds and not NW/SE (along channel) winds.

5.4 Mean longitudinal flow in the upper NRE

Freshwater discharge in the Neuse River varies from low discharge conditions that normally occur during the summer and fall seasons to high freshwater conditions that normally occur during the late winter and into spring (see Chapter 3 for average discharge). Bottom mounted ADCPs along the shores were deployed during both periods and the data from these instruments, combined with a single ADCP deployment in the center of the estuary, were used to construct contour plots of the mean along channel flow for each of these periods (figure 5.10).

Similar pictures are evident for both high and low discharge periods. Downstream flow dominates along the south shore and upstream flow dominates along the north shore. There is a transition zone in the middle of the estuary with typical “estuarine flow” (downstream near the surface and upstream near the bottom). The Rossby number which assesses the importance of rotation is \( R_o = 0.2392 < 1 \), as was shown in Chapter 4, and hence the earth’s rotation should play a role in deflecting lighter freshwater along the southern shore of the estuary. Velocities are slightly reduced during the low discharge periods.

Bi-weekly salinity and temperature data were collected at 3 stations in a cross section of the upper NRE during both high and low discharge periods as part of the MODMON project and were used to compute mean conditions during those periods. Cross channel isohalines for the high discharge period (figure 5.11) indicate fresher water at the surface along the southern shore. The isohalines are tilted down toward the southern shore and tilted up along the northern shore indicating a component of vertical flow. Since salinity is the most important component for density calculations, this pattern persists with the isopycnals (figure 5.12). Salinity values are overall higher during low discharge periods.

5.4.1 Mean lateral dynamic balance in the upper NRE

Over a seasonal mean, internal motions may be dominant and a simplified mean lateral momentum balance may be represented as

\[
f u = - \frac{1}{\rho_0} \frac{\partial p}{\partial y},
\]

a balance between the Coriolis force and the seasonally averaged lateral horizontal pressure gradient. Here \( f \) is the Coriolis parameter, \( \rho_0 \) is a reference density, \( p \) is pressure, \( u \) is the mean seasonal longitudinal component of velocity, \( y \) is positive towards the north shore and \( z \) is positive upwards.

The vertical momentum equation can be integrated and solved for \( p \). The derivative of the equation for \( p \) in the \( y \) direction allows the division of the pressure term into a baroclinic and barotropic component. The atmospheric pressure was assumed constant over \( y \) and the pressure term in the
Figure 5.10: Mean along channel velocity contours (cm/s) in a transect of the upper NRE (+ downstream and - upstream).

Figure 5.11: Mean salinity contours (psu) in a transect of the upper NRE
longitudinal momentum equation reduces to
\[ \frac{\partial p}{\partial y} = g\rho_0 \frac{\partial \eta}{\partial y} + g \int_{z}^{\eta} \frac{\partial \rho}{\partial y} dz \]
allowing the longitudinal momentum equation to be rewritten as
\[ f u = -\frac{g}{\rho_0} \int_{z}^{\eta} \frac{\partial \rho}{\partial y} dz - g \frac{\partial \eta}{\partial y}. \]  

(5.21)

To test the appropriateness of equation 5.21, the simplified lateral momentum equation was evaluated for a two-layer system representing a lateral section of the upper NRE. Similar to what has been done previously, the layer integrated terms of the force balance were examined. In the upper layer, with depth \( h_1 \), the integrated force balance would be represented by
\[ f u_U h_1 = -g \frac{\partial \eta}{\partial y} h_1 - \frac{g}{\rho_0} \left( \int_{z}^{\eta} \frac{\partial \rho}{\partial y} dz \right)_U h_1 \]  

(5.22)

and
\[ f u_L h_2 = -g \frac{\partial \eta}{\partial y} h_2 - \frac{g}{\rho_0} \left( \int_{z}^{\eta} \frac{\partial \rho}{\partial y} dz \right)_L h_2 \]  

(5.23)
in the lower layer, with thickness \( h_2 \). Here, the subscripts \( U \) and \( L \) represent the mean upper and lower layer values. Hence, \( u_U \) is the mean upper layer along channel velocity. Likewise, \( \frac{g}{\rho_0} \left( \int_{z}^{\eta} \frac{\partial \rho}{\partial y} dz \right)_U \) represents the mean baroclinic pressure gradient in the upper layer.

Assuming equation 5.21 represents the main balances in the lateral flow for the upper NRE, across channel velocity in the upper and lower layers, \( u_U \) and \( u_L \), should approximate those values recorded in the upper NRE. The determination of the along channel velocities was completed using data
Table 5.1: Components of equations 5.22 and 5.23 evaluated with model data provided that $h_1 = h_2 = 2$. Surface stress data provided by Climatology section. $U_U$ and $U_L$ are derived.

<table>
<thead>
<tr>
<th>Term</th>
<th>Units</th>
<th>Value</th>
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</thead>
<tbody>
<tr>
<td>$g\left(\frac{\partial h}{\partial y}\right)$</td>
<td>m$^2$s$^{-2}$</td>
<td>$-0.236 \times 10^{-6}$</td>
</tr>
<tr>
<td>$g\left(\frac{\partial h}{\partial y}\right)$</td>
<td>m$^2$s$^{-2}$</td>
<td>$-0.236 \times 10^{-6}$</td>
</tr>
<tr>
<td>$\frac{\rho_0}{\rho_0} \left( \int_z \frac{\partial h}{\partial y} \frac{\partial h}{\partial x} dz \right)_{h_1}$</td>
<td>m$^2$s$^{-2}$</td>
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<tr>
<td>$\frac{\rho_0}{\rho_0} \left( \int_z \frac{\partial h}{\partial y} \frac{\partial h}{\partial x} dz \right)_{h_2}$</td>
<td>m$^2$s$^{-2}$</td>
<td>$1.92 \times 10^{-6}$</td>
</tr>
<tr>
<td>$U_U$</td>
<td>m s$^{-1}$</td>
<td>0.016</td>
</tr>
<tr>
<td>$U_L$</td>
<td>m s$^{-1}$</td>
<td>-0.054</td>
</tr>
</tbody>
</table>

from the EFDC model simulation for the year 2000 that are presented in table 5.1. Here, model data were time averaged over the summer season and values from the upper NRE (table 5.3) were used to evaluate the terms in equations 5.22 and 5.23.

The bottom layer along channel velocity, $u_L$, was calculated from equation 5.23 to be 5.4 cm s$^{-1}$ in an upstream direction, which is comparable to both the model bottom layer velocity (table 5.3) and field data (figure 5.10). The upper layer along channel velocity, $u_U$, was calculated from equation 5.22 to be 1.6 cm s$^{-1}$ in a downstream direction. This value is lower than the summer model mean, however, it is within less than a factor of two of the mean model value. It would appear that equation 5.21 is an adequate representation for the main force balance in the lateral momentum equation and hence a geostrophic balance exists in the lateral direction.

5.4.2 Mean longitudinal dynamic balance in the upper NRE

Over a seasonal mean, the simplified longitudinal momentum balance may be represented as

$$-fv = -\frac{1}{\rho_0} \frac{\partial p}{\partial x} + \frac{\partial}{\partial z} \left( \frac{F_{xx}}{\rho_0} \right),$$

a balance between the Coriolis force, the seasonally averaged horizontal pressure gradient, and vertical shear stress. Here $f$ is the Coriolis parameter, $\rho_0$ is a reference density, $p$ is pressure, $u$ is the mean seasonal longitudinal component of velocity, $x$ is positive down the estuary, $z$ is positive upwards, and $F_{xx}$ is the vertical frictional stress in the $x$ direction.

The vertical momentum equation can be integrated and solved for $p$. The derivative of the equation for $p$ in the $x$ direction allows the division of the pressure term into a baroclinic and barotropic component. The atmospheric pressure was assumed constant over $x$ and the pressure term in the longitudinal momentum equation reduces to

$$\frac{\partial p}{\partial x} = g \rho_0 \frac{\partial \eta}{\partial x} + g \int_z \frac{\partial \rho}{\partial x} dz$$

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allowing the longitudinal momentum equation to be rewritten as

\[-f v = -\frac{g}{\rho_0} \int_z^n \frac{\partial \rho}{\partial x} dz - \frac{\partial \eta}{\partial x} + \frac{\partial}{\partial z} \left( \frac{F_{xz}}{\rho_0} \right). \quad (5.25)\]

To test the appropriateness of equation 5.25, the simplified longitudinal momentum equation was evaluated for a two-layer system representing a longitudinal section of the upper NRE. As has been done previously, the layer integrated terms of the force balance were examined. In the upper layer, with depth \( h_1 \), the integrated force balance would be represented by

\[-f v_U h_1 = -g \frac{\partial \eta}{\partial x} h_1 - \frac{g}{\rho_0} \left( \int_z^n \frac{\partial \rho}{\partial x} dz \right)_U + \frac{1}{\rho_0} (F_S - F_D) \quad (5.26)\]

and

\[-f v_L h_2 = -g \frac{\partial \eta}{\partial x} h_2 - \frac{g}{\rho_0} \left( \int_z^n \frac{\partial \rho}{\partial x} dz \right)_L + \frac{1}{\rho_0} (F_D - F_B) \quad (5.27)\]

in the lower layer, with thickness \( h_2 \). Here, the subscripts \( U \) and \( L \) represent the upper and lower layer values. Hence, \( v_U \) is the mean upper layer across channel velocity. Likewise, \( \frac{g}{\rho_0} \left( \int_z^n \frac{\partial \rho}{\partial x} dz \right)_U \) represents the mean baroclinic pressure gradient in the upper layer. \( F_S = F_{xz}(z = 0) \) is the local along channel wind stress in the upper NRE, \( F_D = F_{xz}(z = -h_1) \) is the internal friction between the two layers and \( F_B = F_{xz}(z = -h) \) is bottom friction equal to the quadratic drag law, \( F_B = C_D \rho_0 u_L^2 \) with \( C_D = 3 \times 10^{-3} \).

Assuming equation 5.25 represents the main balances in the longitudinal flow for the upper NRE, along channel velocity in the lower layer, \( u_L \), should approximate those values recorded in the upper NRE. The determination of the lower layer velocity, \( u_L \), was completed using data from the EFDC model simulation for the year 2000 that are presented in table 5.2. Here, model data were time averaged over the summer season and values from the middle of the upper NRE (table 5.3) were used to evaluate the terms in equations 5.26 and 5.27. Simulation data showed reasonable correspondence with field data as noted by the mean summer contours of salinity and along channel flow (see figures 5.13 and 5.14 of the model output compared with figures 5.10 and 5.11). Lateral variability was also present in the simulation.

The bottom layer along channel velocity, \( u_L \), was calculated from equations 5.26 and 5.27 to be 2.9 cm s\(^{-1}\) in an upstream direction and is comparable to both the model bottom layer velocity (table 5.3) and field data (figure 5.10). Hence, equation 5.25 is an adequate representation for the main balances in the longitudinal momentum equation. In fact, the dominant components in the balance are the barotropic pressure gradient and the baroclinic pressure gradient. A schematic of the important stresses for the representative two-layer system is presented in figure 5.15. It is important to note that local wind stress itself is not a major contributor to the longitudinal momentum balance. It should also be noted that along channel velocity in the bottom layer as calculated with the use of the layer integrated longitudinal momentum balance is closer to recorded and EFDC modeled velocities than that calculated from the layer integrated lateral momentum balance.
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<th>Units</th>
<th>Value</th>
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</thead>
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<td>( f_{u_1} )</td>
<td>( m^2/s^2 )</td>
<td>( 0.0027 \times 10^{-1} )</td>
</tr>
<tr>
<td>( f_{u_2} )</td>
<td>( m^2/s^2 )</td>
<td>( -0.03 \times 10^{-2} )</td>
</tr>
<tr>
<td>( g(\partial \eta/\partial x)_{u_1} )</td>
<td>( m^2/s^2 )</td>
<td>( -0.684 \times 10^{-1} )</td>
</tr>
<tr>
<td>( g(\partial \eta/\partial x)_{u_2} )</td>
<td>( m^2/s^2 )</td>
<td>( -0.684 \times 10^{-1} )</td>
</tr>
<tr>
<td>( \frac{1}{\rho_0} \left( \int_{z_1}^{z_2} \frac{\partial P}{\partial x} dz \right)_{u_1} )</td>
<td>( m^2/s^2 )</td>
<td>( 0.3329 \times 10^{-5} )</td>
</tr>
<tr>
<td>( \frac{1}{\rho_0} \left( \int_{z_1}^{z_2} \frac{\partial P}{\partial x} dz \right)_{u_2} )</td>
<td>( m^2/s^2 )</td>
<td>( 0.6815 \times 10^{-5} )</td>
</tr>
<tr>
<td>( F_S/\rho_0 )</td>
<td>( m^2/s^2 )</td>
<td>( 0.0236 \times 10^{-1} )</td>
</tr>
<tr>
<td>( F_D/\rho_0 )</td>
<td>( m^2/s^2 )</td>
<td>( 0.177 \times 10^{-1} )</td>
</tr>
<tr>
<td>( F_B/\rho_0 )</td>
<td>( m^2/s^2 )</td>
<td>( -0.377 \times 10^{-1} )</td>
</tr>
<tr>
<td>( U_L )</td>
<td>( m/s )</td>
<td>( -0.029 )</td>
</tr>
</tbody>
</table>

Table 5.2: Components of equations 5.26 and 5.27 evaluated with model data provided that \( h_1 = h_2 = 2 \). Surface stress data provided by Climatology section. \( U_L \) is derived.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Units</th>
<th>North shore</th>
<th>Middle</th>
<th>South shore</th>
</tr>
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<tbody>
<tr>
<td>( dx )</td>
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<td>5000</td>
<td>5000</td>
<td>5000</td>
</tr>
<tr>
<td>( dy )</td>
<td>m</td>
<td>5000</td>
<td>5000</td>
<td>5000</td>
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<tr>
<td>( u_U )</td>
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</tr>
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<td>-2.37</td>
<td>0.75</td>
</tr>
<tr>
<td>( v_U )</td>
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<td>0.0010</td>
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<td>-0.0011</td>
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</tr>
<tr>
<td>( f )</td>
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<td>( 1.3 \times 10^{-6} )</td>
<td>( 1.3 \times 10^{-6} )</td>
</tr>
<tr>
<td>( g )</td>
<td>( m/s^2 )</td>
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<td>9.8</td>
<td>9.8</td>
</tr>
<tr>
<td>( \rho_0 )</td>
<td>( kg \ m^{-3} )</td>
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<td>1004</td>
<td>1004</td>
</tr>
<tr>
<td>( \partial \eta/\partial x )</td>
<td>( m/m )</td>
<td>( -3.3 \times 10^{-7} )</td>
<td>( -3.4 \times 10^{-7} )</td>
<td>( -3.6 \times 10^{-7} )</td>
</tr>
<tr>
<td>( \partial \rho/\partial x_U )</td>
<td>( kg \ m^{-3} \ m^{-1} )</td>
<td>( 2.5 \times 10^{-4} )</td>
<td>( 2.37 \times 10^{-4} )</td>
<td>( 1.92 \times 10^{-4} )</td>
</tr>
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<td>( \partial \rho/\partial x_L )</td>
<td>( kg \ m^{-3} \ m^{-1} )</td>
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<td>( F_{xx} )</td>
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<td>( 2.37 \times 10^{-4} )</td>
<td>( 2.37 \times 10^{-4} )</td>
</tr>
</tbody>
</table>

Table 5.3: Mean seasonal values for the upper NRE obtained from the EFDC model 2000 simulation and divided into an upper and lower layer.
Figure 5.13: Mean flow contours (cm/s) in a transect of the upper NRE as computed with EFDC during the summer of 2000.

Figure 5.14: Mean salinity contours (psu) in a transect of the upper NRE as computed with EFDC during the summer of 2000.
Figure 5.15: Schematic illustrating the important stresses in the longitudinal momentum equation applied to a two layer system. Arrow sizes represent relative strength of each component as well as the direction.

5.4.3 Interpretations for longitudinal mean flow

Local wind stress was shown to have minimal influence on the longitudinal momentum equations. However, it was undetermined what role remote wind stress played in determining mean longitudinal flow as well as how the Coriolis force influenced this flow. Additionally, lateral variability in the components of the longitudinal momentum equation is evident (table 5.3) and may also explain characteristics of the mean longitudinal flow. With that in mind, the EFDC model was used to determine the importance of wind and the Coriolis force on the mean longitudinal flow. The model base case, referred to in the prior section, was established based on the summer time average from the simulation year 2000. Perturbations in this base case were studied by running the same model year 2000 simulation with altered boundary conditions. The elimination of wind stress as a surface boundary condition was one perturbation of the model. It is important to note that wind stress plays a large role in altering water level throughout the estuary as a whole (Luetich et al., 2000a; Luetich et al., 2002) and therefore, the eastern water level boundary condition was set to zero as well. A second perturbation was the removal of Coriolis force in the simulation year 2000. In this perturbation, the winds and their associated water level variability were present.

Summer time averages of surface salinity, bottom salinity, and elevation as a function of distance for the three model cases are presented in figures 5.16a-c, 5.17a-c, and 5.18a-c. Here, lateral variation is presented in terms of north shore and south shore values. As the barotropic and baroclinic components

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in the longitudinal momentum balance are the dominant terms, lateral comparisons between them at approximately 32 km downstream for the three model cases are presented in figures 5.16d, 5.17d, and 5.18d. For completeness, the barotropic and baroclinic components in the lateral momentum balance are also presented in figures 5.16e, 5.17e, and 5.18e.

For the model base case (figure 5.16), surface salinity values increased linearly downstream and approximated four year summer season mean values recorded in the Climatology section (see figure 3.9). Bottom salinity values also increased fairly linearly downstream and still approximated those four year summer season mean values. Elevation decreased linearly downstream and represented a 2 cm decrease along the length of the estuary. Salinity along the south shore in the surface was lower than the north shore value from about 40km downstream onwards. Bottom salinity showed considerable variability, but was generally lower on the south shore. Conversely, water level was higher along the south shore than the north shore from about 40km downstream onwards. It is interesting to note that major deviations in elevation and salinity occur approximately 40km downstream and this is the approximate location for the bend in the estuary (see figure 1.2 as a reference). Accelerations due to the longitudinal barotropic pressure gradient indicate an increase from the south shore to the north shore with the largest values along the north shore. Accelerations due to the longitudinal baroclinic pressure gradient indicate an increase from the south shore to the north shore with the largest values along the north shore. Along the south shore, the longitudinal barotropic pressure gradient is larger than the longitudinal baroclinic pressure gradient in the near surface waters and then this pattern reverses. This would lead to downstream flow in the surface along the south shore with upstream flow near the bottom. Conversely, the longitudinal baroclinic pressure gradient along the north shore becomes larger than the longitudinal barotropic pressure gradient higher in the water column and therefore leads to upstream flowing water throughout most of the water column near the north shore. Accelerations due to the lateral barotropic pressure gradient term are towards the north shore whereas accelerations due to the lateral baroclinic pressure gradient term are directed towards the south shore. In the surface, the lateral barotropic term is larger than the baroclinic term which pushes flow northward. Near the bottom, this pattern reverses and pushes flow southward.

When no winds and no water level variability at the downstream boundary are applied to the model (figure 5.17) surface salinities still linearly increase down the estuary, but the overall values are slightly lower than the base case. The exception is near the head of the estuary (0-20 km) where salinity is higher than in the base case. Bottom salinities are dramatically higher than in the base case. In fact, it appears that the salt wedge has pushed completely upstream towards the head of the estuary. Elevations decreased linearly downstream and were reduced to a 1.5 cm decrease along the length of the estuary. Overall, the elevations were lower than the base case. As with the prior case, salinity along the south shore was lower than salinity along the north shore in the near surface from about 30km downstream onwards. Bottom salinities are very similar. Elevation along the south shore was
Figure 5.16: Surface salinity (a), bottom salinity (b), elevation (c), accelerations due to the longitudinal baroclinic and barotropic pressure gradient terms (d) along the north shore and the south shore of the NRE, and accelerations due to the lateral baroclinic and barotropic pressure gradient terms (e) from the model year 2000 simulation. Vertical lines indicate an upper NRE range (31km-36km) over which the longitudinal barotropic and baroclinic components were computed.

higher than elevation along the north shore. Inspection of the accelerations due to the longitudinal barotropic terms (figure 5.17d) shows an increase from the south shore towards the north shore. The longitudinal barotropic terms themselves have decreased from the base case condition. The longitudinal baroclinic pressure gradient term slightly increases from the north shore towards the south shore in the surface and then are similar at depth. In general, these terms are also smaller than the base case. In the surface along the south shore, the longitudinal baroclinic term is smaller than the longitudinal barotropic term and hence downstream flow should dominate along the south shore. Along the north shore, the longitudinal baroclinic and barotropic terms are roughly the same with the longitudinal baroclinic term increasing with depth. A trend of upstream flow should dominate along the north shore. The longitudinal baroclinic pressure gradient becomes larger than the longitudinal barotropic pressure gradient term along both shores and hence upstream flow should prevail at depth. Both the lateral barotropic and baroclinic pressure gradient terms have increased from the base case. Southward flow should dominate from inspection of these two terms.

The scenarios explained above become modified when Coriolis is removed from the model (figure 5.18). Both surface and bottom salinity increase linearly and are slightly lower than in the model base case. Elevations still decrease linearly at an overall decrease of 2 cm over the estuary length. It is interesting to note, however, that lateral variability in salinity and elevation is significantly reduced.
Figure 5.17: Surface salinity (a), bottom salinity (b), elevation (c), accelerations due to the baroclinic and barotropic pressure gradient terms (d) along the north shore and the south shore of the NRE, and accelerations due to the lateral baroclinic and barotropic pressure gradient terms (e) from the model year 2000 simulation without wind stress and downstream water level variation. Vertical lines indicate an upper NRE range (31km-36km) over which the barotropic and baroclinic components were computed.

Surface salinity and elevations are roughly the same across the estuary. This is also reflected in the fact that the accelerations due to the barotropic pressure gradient are very similar laterally. Lateral variations in accelerations due to the baroclinic pressure gradient are evident and are very similar to the model base case. Interestingly, the lateral barotropic term is directed southward but is in general very small. The lateral baroclinic term is directed northward, which is opposite to that for the base case.

A synthesis of this information indicates that the Coriolis force causes fresh water to deflect towards the southern shore. This creates lateral variability in the water elevation where fresher water is piled up along the southern shore. The combination of the opposing lateral barotropic and baroclinic pressure gradients work to push water towards the north shore. Wind forcing of Pamlico Sound has a significant impact on water level variation in the NRE and on the water level gradient in the upper NRE. This additional component of the longitudinal barotropic pressure gradient plays an important role in determining the average position of the salt wedge in the upper portion of the estuary. A stronger longitudinal barotropic pressure gradient and a reduced longitudinal baroclinic pressure gradient along the south shore and a weaker longitudinal barotropic pressure gradient and a stronger longitudinal baroclinic pressure gradient along the north shore mean longitudinal downstream flow persists along the south shore and upstream flow persists along the north shore.
5.5 Summary

Winds play an important role in the upper NRE. The predominant winds over the NRE are oriented in a NE/SW manner. This orientation may be considered across channel winds in the upper NRE and along channel winds in the lower NRE (east of the bend and into Pamlico Sound). On time scales less than a week, these winds were shown to generate a distinct lateral response in the upper NRE. Specifically, they generate pycnocline excursions which expose upper layers of the water column to higher density, lower DO water. The along channel component of this wind in reference to the upper NRE is relatively small and does not seem to generate an Ekman type upwelling in the lateral direction.

Over a seasonal mean, it was shown that the along channel wind component for the upper NRE does not control mean longitudinal flow. The along channel wind over the lower NRE and Pamlico Sound does affect mean longitudinal flow in the upper NRE, however, by impacting the barotropic pressure gradient. Without these winds, the salt wedge pushes upstream towards the head of the estuary because the longitudinal baroclinic pressure gradient is much larger than the barotropic pressure gradient.
Chapter 6

Application: Fish Kills

Major fish kills occurred within the upper NRE from September 24 through September 27, 2000 and again on October 25, 2000. According to the NC Division of Water Quality (Water Quality Section Environmental Sciences Branch, 2000) 30,000 dead Atlantic Menhaden, *Brevoortia tyrannus*, were found near Flanner’s Beach on the south shore of the upper NRE on September 24, 2000. Official reports stated that the menhaden were dead for 12-36 hrs. That same day, 12,500 dead menhaden were found along Kennel beach on the north shore of the upper NRE. Three days later, 78,200 menhaden were found from Flanner’s Beach to Slocum Creek along the south shore of the upper NRE (see figure 1.1 as a reference). In the October fish kill, 14,000 menhaden were found near Flanner’s Beach on the south shore of the upper NRE. At the time of the investigation the fish had been dead for 24 to 48 hours and investigators suspected an environmental stressor such as a shift in oxygen or salinity.

Unfortunately, none of the P-AVPs, CTDs or ADCPs were deployed during the September fish kills because they were being serviced. Only the bottom mounted CTDs were in position for the October fish kill. This circumstance made it difficult to observationally assess physical conditions that may have contributed to the fish kills. Consequently, the calibrated 3-D model of the upper NRE was used to simulate salinity conditions around the times of the fish kills. The model results together with the bottom mounted CTD data provide an indication of likely upwelling events. As was demonstrated earlier (Chapter 4), high salinity and hypoxic water are often coincident suggesting that the upwelling of high salinity water means the upwelling of low oxygen water as well. The presence of low oxygen water in the middle of the channel is confirmed by the MODMON weekly mid-river observational data.

6.1 September 2000 Fish Kills

Simulated salinity for the seven day period that contained the two fish kills is presented (figure 6.1) from a run that simulates year 2000 as presented in Chapter 4. The times when probable fish kills occurred are colored by gray boxes (Water Quality Section Environmental Sciences Branch, 2000). Approximately two days prior to the September 24, 2000 (JD 267) fish kill, winds were light (< 5 m s⁻¹) and directed towards the S/SW. Salinity was somewhat elevated along the north shore in both the near surface and near bottom waters (figure 6.1).
At approximately JD 266, one day before the fish kills were reported, the wind shifted toward the N/NE. With this wind, salinities along the north shore decreased while salinities along the south shore increased (figure 6.1). The wind began to increase in speed and upwelling appears to occur along the south shore by the beginning of JD 267. Bottom salinity along the north shore dropped 6 psu while bottom salinity along the south shore increased 4 psu. A similar trend was seen in the near surface as salinity dropped 2 psu along the north shore and increased 1.5 psu along the south shore. The combination of this salinity variability generated a LSI much less than -1 and suggests severe upwelling (figure 6.2). In fact, the middle water column LSI was -5, much lower than the severe upwelling threshold. This strong upwelling along the southern shore of the upper NRE clearly occurs within the time window identified for the mortality of approximately 30,000 menhaden along Flanner’s Beach on the southern shore of the upper NRE.

During the time that investigators were on the estuary (> JD 267.5), the wind shifted slightly in both direction and speed and the salinity oscillated. This behavior was documented before (figure 4.10) and is probably a baroclinic seiche. This oscillation reversed the upwelling and caused high salinity water to upwell along the north shore. Approximately 12,500 menhaden were reported dead along Kennel Beach along the north shore of the NRE during this same time.

The wind shifted several times during the next three days. At approximately JD 268.5, the wind turned toward the N/NE at 5 m s\(^{-1}\). These winds persisted for nearly a day and bottom salinity along the north shore dropped 6 psu and rose 3 psu along the south shore. Surface salinity dropped 4 psu along the north shore and rose 2.5 psu along the south shore. The combination of these changes generated severe upwelling along the south shore as the LSI for all depths was close to -10 (figure 6.2). Near bottom and near surface salinities were nearly the same along the north shore (figure 6.1). A MODMON mid-river run was conducted on September 27, 2000 (JD 270) and documented a large expanse of hypoxic water in the region (figure 6.3). Thus it appears as though wind-driven upwelling of high salinity/hypoxic water occurred along the southern shore during the time window of this second fish kill.

### 6.2 October 2000 Fish Kill

Bottom-mounted CTD data were available during this fish kill as well as the salinity from the model simulation. The winds during this time were consistently blowing toward the S/SW. Figure 6.4 graphs the model salinity near the bottom and surface and the near-bottom observed salinity from the CTDs. The fish kill was reported on October 25, 2000 (JD 298), although the fish were estimated to have been dead for 24 to 48 hrs. The gray squares in figure 6.4 represent the estimated time of the fish kill.

At JD 296.5, the wind was directed toward the SW at a speed of approximately 5 m s\(^{-1}\) after several days of relative calm. The SW wind initiated a downwelling of fresher water along the south
Figure 6.1: Wind vectors (direction wind is blowing toward) (a) and model salinity along the north (b) and south (c) shores from September, 2000. Shaded regions represent time periods for fish kill occurrence as given by NCDWQ (Water Quality Section Environmental Sciences Branch, 2000).
Figure 6.2: Wind vectors (a) and model LSI for the near surface (a), middle (b) and near bottom (c) during September, 2000. The shaded areas represent times for possible fish kills as stated by the NCDWQ. LSI values above the horizontal shaded areas in (c) signify strong upwelling events along the north shore whereas LSI values below that area represent strong upwelling events along the south shore.
shore (figure 6.4); this phenomena was present in both the model and observed data as salinity near the bottom dipped 4-6 psu. Bottom salinity along the north shore remained high and near surface salinity increased. Upwelling along the north shore was apparent in the near bottom waters as well as in the middle and near surface as LSI values were higher than 2 (figure 6.5). At JD 297.5, the wind shifted direction and slowed slightly. This caused upwelling along the south shore for half a day as salinity along the north shore decreased by 3 psu and increased by 4-6 psu along the south shore.

MODMON mid-river water quality samples were taken on October 19, 2000 (JD 292) and October 25, 2000 (JD 298) and showed that hypoxia was present in the mid-river region near the reported fish kill (figure 6.6 and 6.7). Thus it appears as though wind-driven upwelling of high salinity/hypoxic water along the southern shore was present during the time window of the October fish kill.

### 6.3 Fish kills: a synthesis

This research has shown that hypoxia is generated in the bottom waters when oxygen consumption is greater than the aeration of the bottom waters due to vertical mixing. Vertical mixing and pycnocline ventilation are controlled by the stratification in the system and the wind stress.

The upper NRE is normally stratified during the summer seasons and this stratification can vary in degree and thickness between density layers. Generally, the density difference between layers is strong enough, however, for the generation of hypoxic condition in the lower layers regardless of this layer thickness. This research has documented the wind-driven upwelling of high salinity/hypoxic water along the shores even when hypoxic bottom water was in a small lower layer.

Fish kills were not recorded at the same time that data were collected, however the wind-driven upwelling of hypoxic water was recorded on multiple occasions. This physical response of the system to across channel winds provides a hypothetical albeit plausible mechanism for fish kills in the upper NRE.
Figure 6.4: Wind vectors (a) and simulated and observed salinity along the north (b) and south (c) shores from October, 2000. Shaded regions represent time periods for fish kill occurrence as recommended by the NCDWQ.
Figure 6.5: Wind vectors (a) and the LSI from simulated salinity data for the near surface (b), middle (c) and bottom (d) as well as for the observed data (d) from October, 2000.
Figure 6.6: Axial distribution of dissolved oxygen from October 19, 2000 (JD 292) determined from MODMON data.

Figure 6.7: Axial distribution of dissolved oxygen from October 25, 2000 (JD 298) determined from MODMON data.
It may be that fish are inadvertently caught in a strong hypoxic upwelling event and are killed. Another potential explanation is that fish are stressed by periodic exposure to weaker hypoxic upwelling. In a stressed state, these organisms could be much more susceptible to death due to further weak hypoxic upwelling.

In general, three elements combine to create conditions that would favor a low DO fish kill in the NRE: 1) the presence of hypoxic and severely hypoxic conditions; 2) the presence of fish; and 3) direct or indirect wind-driven advection of the hypoxic volume.
Chapter 7

Summary and Conclusion

The seasonal longitudinal variability of the Neuse River Estuary was determined from the MODMON four year database of mid-river water quality data. The effect of freshwater discharge was particularly evident as it freshened the estuary and pushed the saltwedge downstream during the spring. The estuary was saltier during periods of low freshwater discharge like the summer. Bottom and surface salinities were well correlated linearly with distance downstream ($R^2_s > 0.93$) for all seasons. Linear models for the seasonal relationship are presented (table 3.1). These seasonal relationships were very similar to those determined for the four year mean salinity vs. downstream distance represented by $S = 0.23x + 0.20$ in the bottom and $S = 0.17x - 0.55$ in the surface. Seasonal variability in longitudinal salinity has been noted in other estuaries for example, the Delaware estuary (Wong, 1995), the Tamar estuary (Uncles and Stephens, 1990) and Mobile Bay (Schroeder et al., 1990).

Seasonal vertical salinity stratification was also well correlated with distance downstream. During the summer and the winter seasons, salinity stratification linearly increased from the head of the estuary to approximately 25 km downstream. Beyond this distance, salinity stratification linearly decreased slightly. That pattern altered during the fall and spring where salinity stratification linearly increased from near the head towards the mouth. These results are summarized in table 3.3. Thermal stratification was weak, but occurred during spring and summer presumably due to differential heating and cooling.

Salinity stratification was strongly correlated with bottom water DO for the spring and summer seasons. Increases in stratification implied decreases in bottom DO toward hypoxic conditions. In fact, bottom water DO generally decreased downstream during both the summer and spring seasons. Hypoxic conditions dominated during the summer season. This behavior is also prevalent in systems like the main stem of the Chesapeake Bay (Officer et al., 1984), tributaries of the Chesapeake Bay like the Rappahannock and York rivers (Kuo and Neilson, 1987; Kuo et al., 1991), and Mobile Bay (Turner et al., 1987; Schroeder and Wiseman, 1988). As was found in the Chesapeake Bay (Boicourt, 1992), a DO balance elucidated the fact that the generation of hypoxia is often due to an imbalance between vertical mixing and biological demand. During the summer, biological demand is high and is of the same magnitude as vertical mixing which leads to the depletion of oxygen in the bottom waters.
The lateral variability of the upper Neuse River Estuary was analyzed for the first time. Bottom-mounted ADCPs recorded velocity profiles along the shores of the estuary and were used to describe mean longitudinal circulation. Continuous bottom salinity data were available and were used to determine the effects of freshwater discharge throughout the year. A novel autonomous vertical profiling system was designed and built for the express purpose of recording the dynamic vertical salinity and DO structures at high frequencies and with high vertical resolution. The profiled data were the first of this type to be recorded in the NRE or in virtually any estuary.

As tidal influence on the NRE was minimal due to its restricted connection with the Atlantic via three small inlets, the main causes for variability in the system were freshwater discharge and atmospheric forcing. Fresher water often flows along the southern shore of the upper NRE with saltier water along the northern shore. Mean along channel flows along the southern shore were oriented downstream with a maximum of 6 cm s\(^{-1}\) near the surface and minimum of 0.5 cm s\(^{-1}\) near the bottom. Along the north shore, along channel flows were oriented upstream with a maximum of 6 cm s\(^{-1}\) and a minimum of 0.5 cm s\(^{-1}\). Lagged correlation between discharge and averaged water level showed a significant positive correlation. Changes in water level lagged discharge by 30 days. Bottom salinity and freshwater discharge were highly coherent at low frequencies (8-40 days) and 180° out of phase indicating that high volumes of discharge decreased bottom salinities. Variability in bottom salinity lagged freshwater discharge from Kinston by 18 days in New Bern and this lag grew progressively downstream. It took 29.5 days for discharge to affect bottom salinity at Marker 11, 31.4 days along the south shore and 33.0 days along the north shore. Lateral variability existed between bottom salinity and discharge as bottom salinity along the south shore was influenced by discharge quicker than bottom salinity along the north shore.

At periods of a week or more, salinity and wind were strongly coherent with wind oriented in a NE/SW direction. Bottom salinities were coherent and in phase at this period and it was concluded that the salt wedge was being pushed along the upper NRE by NE/SW winds. The synoptic weather band controlled variability in both water level and salinity at a period of 2 to 5 days. North-northeastward wind uniformly drove water levels in the NRE down and non-uniformly affected bottom salinity. Bottom salinity along the shores were coherent, but approximately 140° out of phase suggesting that this process is both a longitudinal and lateral one. In this case, fresher water was advected down along the north shore while the salt wedge was being blown out of the estuary. South-southwestward wind uniformly drove water levels up and advected fresher water down along the south shore while advecting the salt wedge up the estuary. At these time scales, significant coherence was found between bottom salinity and wind oriented in a NNE/SSW direction. This wind orientation, which represented an approximate across channel direction in the upper NRE, was also significantly coherent with salinity at a diurnal time scale.

Diurnal variability in across channel bottom salinity was prevalent during the summer months.
Significant coherence between salinity at both shores and between salinity and wind confirmed that diurnal oscillations in salinity were related to diurnal oscillations in the across channel wind speed. Bottom salinities were 180° out of phase suggesting that this is primarily a lateral process. Wavelet analysis was used to identify specific times of high energy in the salinity and across channel wind record at a diurnal period. Generally, peaks in energy in the north shore bottom salinity occurred at a diurnal period as wind blew towards the North-northeast. Similarly, peaks in energy in the south shore bottom salinity occurred at a diurnal period as wind blew towards the South-southwest. This diurnal variability may have masked variability due to wind-driven baroclinic seiches that occur at the same period. However, evidence for a wind-driven baroclinic seiche was documented during cooler months when the wind record was not dominated by a strong diurnal variability.

Spectral analysis revealed significant energy at higher frequencies in both water level and vertically averaged across-channel velocity. Coherence tests between these variables yielded a significant coherence at approximately 1.5-2.0 cph or a 30 min period that corresponded to a wind-driven barotropic seiche in the lateral direction. Flow fields were also influenced by the wind. Generally, winds blowing towards the N-NE created a northward surface flow and southward return flow. As wind shifted to the SW, the two-layer flow was reversed. Variations in wind speed and direction altered this paradigm.

The novel autonomous profilers captured time series profiles of DO and salinity every 15 minutes. There was strong negative correlation between salinity and DO which indicated that at these locations, high salinity/low DO water was advected into the area and not generated locally by microbial degradation and respiration. Comparisons of salinity and DO contours with wind vectors showed how wind advected and upwelled volumes of low DO/high salinity water along the shores. This pattern was very sensitive to wind speed and direction. A period of upwelling sometimes existed for only a 3-hour time frame and therefore the 15 minute sampling schedule provided the necessary rigor to capture this high-frequency variability. Because of the tight relationship between salinity and DO in the upper layers, salinity and its change in the water column was used as an indicator for vertical DO variability.

Periods of upwelling were quantified by monitoring salinity values in the upper water column. When salinity values were higher than the mean at these locations, hypoxia was also present. Downwelling was associated with salinity values less than the mean at these locations. A linear relationship existed between across channel wind speed and variability in salinity in the upper water column. In this case, winds blowing towards the N/NE decreased salinity along the north shore and increased salinity along the south shore (i.e. downwelling along the north shore and upwelling along the south shore). The situation reversed for winds blowing towards the S/SW. A linear, two-layer model was applied to the upper NRE to determine the vertical migration of the pycnocline due to across channel winds. This linear model documented the vertical pycnocline excursion based on different wind speeds and varying density differences between the layers. The model was further applied with a varying wind speed and compared with data collected by the autonomous profilers. The linear model replicated conditions recorded in the
field with correlation coefficients of 0.51 along the north shore and 0.4 along the south shore.

The observational data were used to validate a 3-D baroclinic, finite difference model called the Environmental Fluid Dynamics Code (EFDC). The EFDC model was calibrated to simulation year 1998 using data acquired from the MODMON program and from NCSU. The model’s vertical resolution was increased and validation of its accuracy, precision and correlation with field data was completed for simulation years 1999 and 2000. Model data was used to interpret the lateral variability in mean longitudinal flow. It was determined that lateral variation in the longitudinal barotropic and baroclinic pressure gradients lead to the lateral variation. Specifically, along the north shore, the longitudinal barotropic pressure gradient is weaker than along the south shore, yet the longitudinal baroclinic pressure gradient is stronger. This forces saltier water upstream along the north shore and fresh water downstream along the south shore.

Experimentation with the calibrated EFDC model was done to determine the influence that discharge and wind individually have on lateral variability. The experimental conditions for the model simulated seasonal conditions where either discharge, atmospheric forcing or both dominated as forcing functions. For example, late winter and early spring was dominated by high discharge and strong winds. During the summer, discharge was low and changes in salinity were controlled by the wind. Without wind, the salt wedge pushed upstream towards the head of the estuary since the baroclinic pressure gradient is stronger than the barotropic pressure gradient.

The model reproduced velocity and salinity comparable to recorded values. Peaks in the energy spectra of simulated salinity matched those in the spectra of recorded salinity. Strong coherence existed between the simulated and recorded salinity and were nearly in phase. The model faithfully reproduced salinity during the summer of 2000 and matched conditions in October during a fish kill along the southern shore. Through model output, it was shown that the wind-driven advection and upwelling of high salinity and therefore low DO water was present during the times that fish kills occurred in the summer and early fall of 2000.

Historical accounts of fish kill in the NRE document the concurrence of severely hypoxic bottom water. One possible explanation for the cause of fish kills in the upper NRE is that hypoxic water that was generated along the channel may be large enough to engulf fish within a cross section of the upper NRE. Another possible explanation, suggested by this research, is that this volume of hypoxic water can be driven by across channel winds toward the shores where it will be upwelled. These upwelling events can occur on time scales long enough to stun or kill fish within the area, yet short enough to change entirely before an investigation was conducted.

7.1 Estuary Comparison

The Mobile Bay Estuary is a semi-enclosed basin on the north coast of the Gulf of Mexico (Schroeder and Wiseman, 1988). Similar to the morphology of the NRE, it is a combination of
a drowned river valley and bar-built estuary with a shallow average depth of 3m. The NRE is microtidal with a maximum tidal range of 0.05m and the Mobile Bay Estuary is also microtidal with a maximum tropic tide range of 0.8m and a mean diurnal range of 0.4m. The bathymetry of Mobile Bay estuary is complex and has a dredged ship channel that cuts through it. It can be strongly salinity stratified with a value of 5 psi/m. Density stratification is always stable and is controlled primarily by salinity (Schroeder and Wiseman, 1988) as is true in the NRE as well. Subtidal variability in salinity has been recorded at periods of 5 to 20 days, however it has not been determined if these fluctuations are due to wind stress, river runoff, or fortnightly tides. This research provided evidence for subtidal variability in salinity in the upper NRE for periods of 24 hours to 20 days related to wind stress. Seasonal affects due to discharge were also noted for the NRE.

In the Mobile Bay estuary, there is a strong relationship between salinity stratification and the vertical distribution of dissolved oxygen. In general, hypoxic water generated by strong salinity stratification is advected by tides and wind-driven baroclinic motions. These wind-driven baroclinic motions have been used as an explanation for the occurrence of “Jubilees”, or the appearance of fish along the shores (usually the eastern shore) in a deteriorating or dying state (Schroeder and Wiseman, 1988). The main explanation for why these “Jubilees” occur is that light (5 m s⁻¹) winds over the days preceding the event turn to easterly winds on the day previous to and during the jubilee resulting in surface waters being moved away from the eastern shore. Lower-layer hypoxic water moves toward the eastern shore to maintain continuity. The jubilee ceases if the wind direction changes. As in the Mobile Bay Estuary, this research documented the local wind-driven advection of low DO/high salinity water towards the windward shore.

Pioneering work on the dynamics of estuaries has been done on the Chesapeake Bay and its tributaries. An early study of the oxygen dynamics off Calvert Cliffs in the upper Chesapeake Bay showed a strong correlation between oxycline depth and halocline/ pycnocline depth and a dependence of oxycline depth on wind direction and impulse during the summer (Carter et al., 1978). At the surface, the on/offshore wind impulses were one of the variables responsible for variations in DO. With depth (6m and 12m), alongshore wind impulses became more significant for DO variability. The importance of the alongshore wind in tilting pycnoclines laterally and leading to upwelling of high salinity/low DO water was further elucidated (Sanford and Boicourt, 1990; Sanford et al., 1990).

Similar to the findings for the upper Chesapeake Bay, this research showed that the NRE directly responds to local wind forcing. Data analysis as well as dynamic analysis showed the direct upwelling of high salinity/high density/low DO water on the windward shore of an across channel wind. Instances of a response to remote wind forcing were also noted in increases of higher salinity water in the upper NRE due to an across channel wind of longer duration forcing the salt wedge up the estuary (Chapter 4, Results section). No direct response was seen in the upper NRE to an along channel wind. Although this response is possible, weather systems passing through the NRE are primarily in the across channel
direction. Hence, the longitudinal response seen in the upper Chesapeake (Sanford et al., 1990) was not documented here.

The longitudinal mean flow in the upper NRE is not simply a geostrophic balance as has been assumed for partially mixed estuaries (Pritchard, 1956). In the NRE, remote winds play a large role in maintaining the position of the salt wedge in the estuary and controlling the overall dynamics. Since the NRE is microtidal, the study of the winds on controlling the dynamics has been more conclusive than on other tidally influenced estuaries like the Chesapeake, for example. In other ways, the NRE has dynamic similarities with Chesapeake Bay and its tributaries. This is not surprising considering its geographic location.
Bibliography


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