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Investigation of the Physical Mechanisms Controlling Exchange Between Mount Hope Bay and the Sakonnet River

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INVESTIGATION OF THE PHYSICAL MECHANISMS CONTROLLING
EXCHANGE BETWEEN MOUNT HOPE BAY AND THE SAKONNET RIVER

BY
WILLIAM DELEO

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE
REQUIREMENTS FOR THE DEGREE OF
MASTER OF SCIENCE

IN
OCEANOGRAPHY

DEAN OF THE GRADUATE SCHOOL

UNIVERSITY OF RHODE ISLAND

UNIVERSITY OF RHODE ISLAND

2001

Abstract

The complexities of estuarine circulation control the exchange of physical, chemical and biological materials between oceanic and riverine systems. This exchange is predominantly driven by three distinct forcing mechanisms: tides, winds and density. The influence of each mechanism must be understood in order to interpret and predict the currents. This study investigates the circulation in the upper Sakonnet River (SR) and Mount Hope Bay (MHB), with respect to each forcing mechanism.

This study also investigates how exchange between the lower SR

APPROVED:

Thesis Committee

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Abstract

The complexities of estuarine circulation control the exchange of physical, chemical and biological quantities between oceanic and riverine systems. This exchange is predominantly driven by three distinct forcing mechanisms: tides, winds and density. The influence of each mechanism must be understood in order to interpret and predict the currents. This study investigates the circulation in the upper Sakonnet River (SR) and Mount Hope Bay (MHB), with respect to each forcing mechanism.

This study also investigates how exchange between the lower SR and MHB is affected by the complex geomorphology of the upper SR. Circulation in the upper SR, or the Sakonnet River Narrows (SRN), is complicated by two factors. MHB is connected to the ocean through the SR Passage and through the East Passage (EP), both parts of greater Narragansett Bay (NB). MHB is therefore exposed to two distinct sources of tidal forcing. Furthermore, the SRN has a series of breakwaters and natural coves which have unforeseen effects on the flow. NB in general has highly variable bottom topography, and this is compounded in the SRN by the man-made structures.

The methods utilized in this study of the SRN were tide gauges, deployed for 41 days, and ADCP/CTD surveys, conducted over two spring-tide tidal cycles and two neap tide tidal cycles. These data were

analyzed in conjunction with local wind data to investigate both localized and regional aspects of the circulation of MHB. To compliment the field study, a linear admittance model was developed for the region.

In general, the tides dominate upper SR volume transport. Maximum observed spring tide current velocity values reach 1.5 m/s, corresponding to a volume transport of about 1000 m³/s. The tidal prism for the SRN ranges from 4.02x10⁶ to 1.39x10⁷ m³ for spring and neap conditions, respectively. The tidal currents between the SR and MHB are predominantly semidiurnal, but the currents exhibit a double-peaked flood and a single-peaked ebb indicative of a significant M4 component. Peak ebb current occurs shortly after high water.

The cross channel structure of the velocity field was observed to be consistently and dramatically inhomogeneous. Furthermore, flow in the vicinity of the breakwaters was often turbulent, and regularly mixed the water passing through the SRN. The mixing is believed to have a detrimental effect on the flushing of MHB heat pollution.

An empirical barotropic volume transport model was developed from sea surface height and volume transport data. Model predicted transport agrees with observed values within approximately 25%. Low frequency transport variability was examined by comparing predicted transport data with wind data. Storm events are believed to explain large anomalies in the predicted low frequency transport record; winds

out of the SE/NW were found to be the most influential. Low frequency transport anomalies at times exceed tide-induced transport, suggesting that the wind-driven flow can at times exceed that due to the tides.

Overall results from the study suggest that upper SR geomorphology has a significant influence on the exchange of water between MHB and the SR. In particular, the breakwaters are believed to induce anomalous tidal current phase, and allow for significant wind induced low frequency transport anomalies. Therefore, the influence of the breakwaters should be considered for future physical oceanographic studies conducted in the region.

Dedication

*This work is dedicated to the Moon, for She drives the tides,
and to the Sun, for He drives the winds,
and to the Earth, for She is the Mother of us all.*

WITHIN US

*Let us learn the revelation of all nature and thought;
that the Highest dwells within us, that the sources
of nature are in our own minds. As there is no screen
or ceiling between our heads and the infinite heavens,
so there is no bar or wall in the soul where we, the
effect, cease, and God, the cause begins.*

*Within us is the soul of the whole; the wise silence,
the universal beauty, to which every part and particle
is equally related; the eternal One.*

*When it breaks through our intellect, it is Genius;
when it breaths through our will, it is Virtue; when
it flows through our affections, it is Love.*

~Ralph Waldo Emerson~

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1 Introduction

1.1 Objectives and significance

The characterization of the hydrodynamics of a shallow estuarine system has multiple aspects. Estuarine circulation is determined by the interaction of the system and the various forcing mechanisms. Forcing mechanisms typically dominant in estuaries include the tides, the winds, river inputs and water column density differences. The system investigated in this study is of particular interest due to its complicated geometry and variable bottom topography. This study will investigate the response of the system to each forcing mechanism in terms of the resulting mixing and exchange. Topics addressed will include: tidal prism and residual flow, the relationship between sea surface oscillations and current oscillations, the effects of wind stress, the degree of stratification, the significance of overtides, and water column flow structure.

Circulation patterns in Narragansett Bay (NB) are complicated due to NB's complex geomorphic features (Figure 1.1). NB consists of three main channels that connect the upper bay to Rhode Island Sound (RIS): the East Passage (EP), the West Passage (WP), and the Sakonnet River

Figure 1.1. Narragansett Bay bathymetry. Narragansett Bay may be characterized as a drowned river valley type estuary. Three main channels connect the upper bay to Rhode Island Sound; they are, from left to right: the West Passage, the East Passage and the Sakonnet River Passage. Note the different depths of the channels, especially the difference in depth between the EP and the SR.

Passage (SR). The different mean depths of the three channels suggest distinct tidal phase speeds (Figure 1.1). (Note: The abbreviation “SR” will be used interchangeably for the “Sakonnet River” and the “Sakonnet River Passage” throughout this work.) Mathematical representation of the interaction of these waves in the upper bay is not straightforward. Results are presented here from an observational study and simple analytical models, which aim to characterize the small-scale variability of the tides in the upper SR. The study site is situated at the junction of the SR with Mount Hope Bay (MHB), located in northeastern NB (Figure 1.1). MHB also connects with the EP to the southwest.

Tidal propagation through the upper SR is further complicated by a series of breakwaters and coves (Figure 1.2). This region between the breakwaters is referred to as the Sakonnet River Narrows (SRN). Previous modeling has not included these features when predicting tidal currents in the region (Gordon and Spaulding 1987). This study will focus on the SRN in order to examine the influence of the breakwaters and coves on tidal propagation through the system.

A secondary focus of this study is the effects of wind stress on circulation. Wind stress induced motion in NB has been found to equal or exceed tide-induced motion (Weisberg 1976a). It has also been suggested that mean wind speed be included in residence time calculations (Pilson 1985). In addition, larger wind induced sea surface slopes have been predicted for the SR than in either the EP or the WP

Figure 1.2 MHB tidal flushing. Shallow regions (>20 ft) are in grey, deeper regions in white. The two inlets to MHB, EP (red line) and SR (blue line), were sampled for transport (Figures 1.3 and 1.4) by Kincaid (1996). The SR station is co-located with the northern transport station for this study. The Brayton Point Power Plant is located in the north-east corner of the bay. The region identified as the Sakonnet River Narrows is bracketed by two breakwater type constrictions, both remnants of out-of-use bridges. The northern bridge was a train bridge, and the southern was a foot bridge, referred to as the "stone bridge".

(Gordon and Spaulding 1987). The effects of wind stress will be addressed by comparing low frequency sea surface height data with meteorological data.

While the primary goal is to constrain NB circulation, a number of additional motivations exist for this study, such as the role of turbulent mixing in the SRN, and the relative exchange of water between MHB, the SRN and the lower SR. Results may also provide data for more precise numerical models, and for mass balance calculations. Furthermore, all such data and results may be applied to investigations of pollution in MHB.

This study investigates the role of enhanced friction within the SRN in terms of the production of overtides and the mixing of the water column. Studies have shown that overtides, or fluctuations at higher harmonic frequencies of the dominant constituent, tend to form in regions where friction is important (Parker 1991). Bottom friction is the primary mechanism for energy dissipation, and has a non-linear contribution to the equation of motion. Levine and Kenyon (1975) sited the southern breakwater, originally a stone bridge (Figure 1.2), as being the area of greatest tidal energy dissipation in all of Narragansett Bay. Other studies conducted in NB have observed a double flood (Weisberg and Sturges 1976, Turner 1984), which is indicative of a significant quarter-diurnal component, an overtide of the dominant semi-diurnal component.

Additional motivation for this study came from a report issued by the Rhode Island Department of Environmental Management. This report suggests that there is a strong correlation between a decline in finfish stocks in MHB and NB and an increase in water use from the Brayton Point power plant (BPPP) located in upper MHB (Applied Science Associates 1996). Two reasonable explanations for such a correlation are: the intake of the BPPP may be entraining too large a quantity of fish larvae and eggs, or the thermal discharge from the plant is having a detrimental effect on the biota. Thermal effluent discharge has been reported to have variable effects on the local biota depending on species migration patterns and natural background temperatures (Abbe 1988). This study will address the influence of the SR and the SRN on the mixing and flushing of MHB, and in turn the flushing of the heated plume.

In addition, this study will provide the necessary data to calibrate more precise numerical models of MHB circulation, tidal prism and residual flow. Present models (Gordon and Spaulding 1987) simulate MHB tidal influence with purely semidiurnal "in phase" tides for the two passages, EP and SR. This study will investigate whether the semidiurnal constituent alone can be used to predict the flow regime, or if overtides or other constituents are required. Improved models will better approximate the flushing time of the bay, and improved flushing time estimates are essential to evaluating the relative impact of the BPPP

heated effluent discharge on the overall temperatures in the bay. Furthermore, these estimates are also crucial to evaluating the impact of other types of pollutants, such as nutrients.

An additional curiosity arose from a circulation study conducted by Kincaid (1996). Kincaid used a shipboard acoustic Doppler current profiler (ADCP) to investigate the transport at each of MHB's inlets: EP, SR, and the Taunton River (TR). Results from the study suggest that the tidal currents at the EP inlet (MH Bridge) are sometimes "out of phase" with those at the SR inlet (SR Bridge) (Figures 1.3 and 1.4). Hicks (1959) also noted anomalous semidiurnal tidal phases in this region. The phase relationship of the current in these two areas has a direct impact on the mixing and flushing of MHB. This study is designed to address this phase difference and determine the mechanism that causes it.

Previous tidal prism and residual flow estimates made in this area have utilized current meters and produced questionable results (Spaulding and White 1990). One strong point in favor of the use of the ADCP for this purpose is that it samples a large fraction of the channel cross-section, whereas the current meter estimates net flow from a point measurement or series of point measurements. Therefore, this work will address the effectiveness of a shipboard ADCP survey for the assessment of the tidal prism and residual flow.

Figure 1.3 Neap tide MHB tidal current. Measurements of transport at the EP (a) and the SR (b) inlets to MHB were collected by Kincaid (1996) with a ship mounted ADCP. Predicted tidal height for Newport is plotted in panel (c). Note that current direction was observed to differ between the two inlets at two points over the cycle: at mid-flood and at high water. At these times, inflow to MHB was observed through the EP and outflow through the SR.

Figure 1.4 Spring tide MHB tidal current. Measurements of transport at the EP (a) and the SR (b) inlets to MHB were collected by Kincaid (1996) with a ship mounted ADCP. This figure is similar to Figure 1.3. Predicted tidal height for Newport is plotted in panel (c). Note that current direction was observed to differ at high water, as with the neap tide survey.

1.2 Hypothesis

Tidal currents in the upper SR were found to lead those at the EP inlet to MHB (Hicks 1959, Kincaid 1996). Although the SR inlet passes only approximately 10% of the water volume passed by the EP inlet (Kincaid 1996), a phase difference between the two inlets may alter the influence of the SR on MHB mixing and flushing. These data also provide a glimpse into overall NB circulation systematics.

Possible mechanisms for a phase difference may be either "regional" or "local". Regional explanations might originate from the difference in shallow water wave speed as measured from Rhode Island Sound (RIS) up each passage (Figure 1.1). Since the tides travel from west to east along RI's southern shore (Hicks 1959), the tides at the mouth of the EP should lead those at the mouth of the SR. Furthermore, the average depth of the SR is much less than that of the EP (Pilson 1985), suggesting that the phase speed of the tides in the EP would be greater than the speed in the SR. Therefore, the difference in tidal wave propagation up the two passages cannot explain the tides in the upper SR leading those in the upper EP. A more likely scenario, which is to be tested by this observational study, is that the local geometry of the SRN produces the anomalous tidal current between MHB and the SR (Figure 1.2).

2 Narragansett Bay review

2.1 Setting

NB may be characterized as a drowned river valley (Figure 1.1). Most areas of the bay have central channels (~12-55 meters), with shallow (~3-6 meters) flats to either side, and the upper SR and MHB are no exception. MHB mean low water depth is 6 m or less in over 70% of the area (Spaulding and White 1990), but at the MH Bridge, depths reach 24 m. MHB has a mean tidal range of 1.34 m, with a maximum of 1.68 m during spring tides and a minimum of 1.0 m during neap tides (Spaulding and White 1990). The mean annual fresh water discharge is $17.94 \text{ m}^3/\text{s}$ from the Taunton River (TR) and Three Mile River combined (Spaulding and White 1990).

2.2 Previous work

The physical oceanography of NB is by no means a new topic. Multiple previous studies have addressed the tidal and non-tidal volume transport of NB and MHB. But the methodologies of each study have been different and the results reflect this. In Section 2.2.1, the various

methods utilized to study NB volume transport will be presented. In Section 2.2.2, previous studies that produced results specifically relevant to observations made in this study will be presented.

2.2.1 Narragansett Bay Transport

Both the tidally oscillating volume transport and mean, or non-tidal, volume transport have been investigated within NB. Methodologies adopted for such measurements have often been dictated by the state of the art instrumentation. Previous observational studies have utilized both Eulerian (Hicks 1959, Levine 1975, Weisberg and Sturges 1976, Weisberg 1976a, Turner 1984, Spaulding and White 1990, Kincaid 1996) and Lagrangian (Haight 1938, Binkerd 1972) measurements, resulting in data with a broad range of temporal and spatial characteristics. Eulerian measurements, such as those collected with a current meter, describe the change of current velocity as a function of time at a fixed position. Traditional current meters make point measurements, while other Eulerian measurements sample a wider region; for example, an ADCP samples most of a water column, and geo-electromagnetic measurements integrate an entire passage. Lagrangian measurements, such as those collected with floats, describe the change of position with time due to the flow. Floats have been utilized to sample both surface and sub-surface

currents. In this section, the various methods utilized to measure or approximate Narragansett Bay volume transport will be discussed.

Results produced with numerical modeling (Gordon and Spaulding 1987) will be discussed later in Section 2.2.2.

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Floats provide a Lagrangian measurement of the water movement; such measurements essentially tag a piece of fluid and track its position with time. Various types of floats utilized in NB include: surface floats that provide a point measurement at the water surface, sub-surface floats that provide a point measurement at depth, and pole floats that provide a depth integrated measurement of water column velocity.

Surface floats are perhaps the most convenient to deploy and recover, but the results are often contaminated by winds. Pole floats also ride partially out of the water and therefore are also affected by wind, but not as severely as the simple surface float. A collection of sub-surface floats designed for different depths is useful to obtain a suite of point depth measurements. Some types are attached to a surface float for easy recovery, providing a combined measurement of surface and depth.

iv Haight (1938) conducted an extensive physical oceanographic study of NB and utilized surface floats for tidal current measurements.

Available in a somewhat raw form, these data were also utilized by Hicks (1959) to investigate the non-tidal flow of NB. Although sub-surface current meters were available at the time, Hicks used Haight's float data

because he found the available current meter records to be of inadequate length for non-tidal flow calculations (Hicks 1959). Hicks' method utilizes surface float data, offshore salinity data, river runoff data, and cross-sectional area values to calculate estuarine residual flow. In a similar way, Pilson (1983) calculated NB transport using available fresh water input and mean salinity data.

In Hicks' (1959) study, waters above the halocline were assumed to have a mean down-bay velocity, and those below were assumed to have a mean up-bay velocity. Upper water layer mean velocity was approximated by multiplying the surface velocity value by 0.75. But since lower layer velocities were not available, net non-tidal transport values could not be directly deduced. Therefore, freshwater input information was utilized to infer bottom layer velocity. To assess the accuracy of the calculations, measurements of the offshore salinity were compared with salinity values approximated from calculations of non-tidal flow within the Bay. This method is limited primarily because of the required approximations of surface layer velocity from surface float measurements, and due to the lack of consideration for ground water input, precipitation and evaporation. Nevertheless, results from the Hicks study indicate that predicted offshore salinity values, were found to be similar to measured salinity values.

Results for non-tidal flow for MHB were found to be $385 \text{ ft}^3/\text{s}$ and $165 \text{ ft}^3/\text{s}$ for the upper and lower layers, respectively. Tidal currents

were found to exhibit standing wave type motion with a slight progressive component. The flood current exhibited a double peak in current velocity, which increased in prominence in the upper bay. In addition, the combined M2, M4 and M6 tidal constituents were found to approximate the observed tidal current well.

Binkerd (1972) investigated the utility of "pole floats" to obtain a more integrated measurement of current velocity than the surface floats used by Haight (Binkerd also attempted to calibrate an electrokinetograph with these floats. A description of the electrokinetograph technique is discussed below.) Pole floats can be fabricated of various lengths and stand upright in the water. Binkerd also used "droag floats" in his study; droag floats have a surface float attached to a sub-surface "sail". Both of these float types are prone to wind interference. Results from the study indicate that the velocity of the pole float does not closely match the average velocity over its length.

Current meters provide an Eulerian measurement of water movement. Instruments are typically mounted on a pier or dock, but can also be mounted on a buoy/anchor line for measurements off the coastline. Many units take measurements mechanically, some with a rotating propeller directed into the flow. Others may use acoustics or electromagnetics. (The electromagnetic current meter is not to be confused with the passive geo-electromagnetic method discussed below.) Mechanical and electromagnetic current meters make point

measurements, and therefore multiple instruments are often deployed at various depths to estimate the vertical velocity profile. Acoustic instruments, such as the ADCP, are mounted either on the surface or bottom and sample the majority of the water column at great resolution. One limitation to the ADCP is that it cannot sample very near the surface or bottom due to instrument housing ringing and side lobe contamination, respectively.

Weisberg and Sturges (1976) conducted a study of the net circulation of the WP of NB. They utilized two sets of mechanical current meters. One set was mounted on a rigid mooring deployed for the study, and the other off the shore on a buoy/anchor line. To obtain transport information, these investigators had to extrapolate vertically between points at the mooring location, and then horizontally across the channel. Furthermore, to obtain the net or non-tidal circulation, these data had to be low pass filtered.

Results from Weisberg and Sturges (1976) indicate that the instantaneous current can be characterized by the semi-diurnal tide. Non-tidal flow was an order of magnitude less, and was found to be well correlated (0.7) with the longitudinal component of the winds. Furthermore, the net transport for the WP was observed to be either landward or seaward for several days duration. Winds were therefore concluded to be the dominant driving mechanism of net transport in the WP.

A second current meter study focused on the non-tidal flow in the Providence River, a major tributary to NB (Weisberg 1976a). Two current meters located two and four meters off the bottom on a buoy/anchor line were deployed for 51-days duration. The record of current velocity was correlated with wind data. Results indicated that 48% of the variability in the current was at sub-tidal frequencies. Furthermore, the variability in the sub-tidal current correlated very well (97%) with the longitudinal winds. It was concluded that in comparison to the tides, the winds have an equal or greater influence on the circulation of the Providence River.

The most recent NB study using current meters focused on the circulation dynamics of MHB and the Lower TR (Spaulding and White 1990). Point-sampling current meters were deployed at the surface and bottom at three locations. In addition, current data were compared with tide height data and wind data. Volume conservation calculations utilizing fresh water input estimates and transport data (extrapolated from velocity point measurements) were not found to balance. The authors suspected that the spatial variability in the current induced the disagreement. Results from this study will be discussed further in Section 2.2.2.

“Geo electromagnetic” methods have also been used to measure currents in NB. The movement of charged particles across magnetic field lines results in a voltage, detailed by Faraday’s Law of Induction. In 1832, Michael Faraday first proposed that this “motionally induced”

voltage (MIV) could be measured to infer the volume transport of seawater, because it carries charged ions through the Earth's magnetic field. The first measurement of the small voltage produced by the tidal current was not observed until 1881. But instrumentation quality has improved over the years, allowing the technique to be refined.

Krabach (1970) investigated the feasibility of measuring the MIV in the West Passage of NB. Three electrodes were placed across the channel. The potential difference between the middle electrode and the adjacent electrodes produced two records of voltage, one for each side of the channel. Results for each side were compared with tidal records. When comparing signal frequency components, the study found a close relation between tidal height and observed voltage. Further West Passage studies were conducted in 1999 (Krezan 1999) using a new type of electrode (Ag/AgCl) that provided more accurate voltage readings. Results from the study, although qualitative, illustrate a close relationship between observed voltage oscillations and the tides. In both studies, the relationship between voltage amplitude and tidal current amplitude remained uncertain, and therefore the volume transport could not be determined without calibration with direct observations.

and Many current meters produced today, such as the ADCP, utilize acoustics to measure water velocity. The technique is based upon the Doppler shift of acoustic waves (see Section 3.1.2). A moored ADCP can sample the majority of the water column by range gating the return

signal, and therefore yielding better spatial coverage than one or even a series of point sampling current meters.

Kincaid (1996) utilized an ADCP to conduct a transport study of MHB. But the instrument was not deployed on a fixed mooring, as most current meters have been in the past. By mounting it on a research vessel, it was used to measure channel velocity cross-sections. This type of sampling produces a relatively complete velocity field and requires far less extrapolation to calculate transport data. Velocity data were collected in this way at four locations in MHB and the lower TR, sampling each station periodically over a full tidal cycle on multiple days. Although this sampling scheme produces far better spatial coverage than that produced by a moored instrument (Kincaid sampled four channel cross-sections with just one instrument and one vessel), one significant limitation is low sampling frequency (Kincaid sampled once every 1-2 hours).

The importance of sampling the majority of the channel cross-section was clearly illustrated in the data; significant vertical and horizontal variability in the velocity field was consistently observed. In fact, channel velocities were often observed to exhibit inflow on one side and outflow on the other. Further results from Kincaid (1996) will be discussed in Section 2.2.2.

2.2.2 Mount Hope Bay Circulation

Previously, the SRN has received attention due to its often-anomalous behavior. Although no other study was specifically designed to address the same questions as this study (namely the influence of SRN geometry on SR/MHB exchange), many observations made in previous studies lend insight into the observations presented here. Such observations will be presented in this section and referenced later in the work as results are drawn upon. There are five previous studies in particular that offer information regarding the curious behavior of the SRN: Hicks (1959), Levine and Kenyon (1975), Gordon and Spaulding (1987), Spaulding and White (1990) and Kincaid (1996).

The most notable aspect of the SRN physical oceanography is the tidal current. In terms of current amplitude, the SR exhibits velocities higher than any other region of NB, 130 cm/s (Levine and Kenyon 1975). In fact, Levine and Kenyon estimated that the upper SR was the region of highest frictional tidal energy dissipation in all of NB, comprising about $\frac{1}{4}$ of the total. Both Spaulding and White (1990) and Kincaid (1996) observed lower current velocity values, approximately 50 cm/s and 5 cm/s respectively, but they both sampled in a lower velocity region of the channel, north of the breakwaters.

In terms of tidal phase, Hicks (1959) noted that the current in the upper SR was found to lead the current at adjacent stations by 1-1.5 hours. In addition, when Kincaid (1996) investigated the tidal current between MH Bridge and the SR, he too found approximately a 1-hour phase lead in tidal current at the SR. Neither work attempts to provide an explanation for the anomaly.

The SR has also previously exhibited anomalous non-tidal flow values, in other words, low frequency variability. The non-tidal flow, or low frequency transport variability, has also been found to be anomalous. From mid-depth current meter data, Spaulding and White (1990) calculated mean flow for the SR passage to be eight times that of the entire TR fresh water input to MHB. Since the calculation of mean transport was dependent on one current meter at one depth and one location in the channel, the authors suspected lateral velocity variability as the explanation for this anomaly. A few years later, ADCP measurements collected by Kincaid (1996) confirmed the existence of extreme velocity variability in the same region. But, an additional factor controlling low frequency transport is wind forcing. Although Spaulding and White (1999) found little correlation between wind data and current data, Weisberg (1976a) and Pilson (1985) suggest that winds play a major role in the non-tidal transport of NB water. As an additional point of interest, Gordon and Spaulding (1987) conducted a modeling study of NB

wind driven circulation, and found the SR to exhibit the highest wind supported sea surface gradient of the three NB passages.

Over In the following Chapters, observations and analysis of upper SR/MHB circulation will be presented. Results from this study will be used in conjunction with those previously collected to explain tidal patterns observed in the SRN.

3 System Hydrodynamics

Overview

Three channels contribute to the flushing and residence time of Mount Hope Bay (MHB) water (Figure 1.1): the Taunton River (TR), and the East and Sakonnet River passages (EP and SR) of Narragansett Bay (NB). The TR is the primary fresh water source, while the EP and SR provide marine water and tidal forcing. This study investigates the exchange of water between MHB and the SR. Observations were focused in the upper SR, where there exist two man-made constrictions similar to breakwaters which influence the response of the system to tide and wind forcing (Figure 1.2).

The field study, conducted in the spring of 1997, was specifically designed to investigate the influence of the breakwaters on the circulation and exchange of MHB water. Transport measurements were made with a 1200 kHz ADCP, and sea surface height measurements were collected with three strain gauge tide gauges. Section 3.1 provides details of the data collection methodology. Section 3.2 present general observations of tidal components (3.2.1), transport (3.2.2), and their relation (3.2.3). The influence of wind forcing will also be investigated (3.2.4). In Section 3.3, study observations are combined with previously

collected data (Kincaid 1996) to discuss system function. Conclusions from the field study are presented in Section 3.4.

3.1 Field methods

Even though the SRN system is small with respect to a tidal wavelength (Figure 1.2 and Figure 3.3), tide height and current data collection sites were chosen based on the assumption that significant tidal variability exists within the SRN. Sea surface height measurements were collected to the north and south of the breakwaters and within the SRN. Water transport measurements were collected at the northern and southern entrances to the SRN and at the junction between the SRN and its interior cove.

3.1.1 *Tide height*

Three tide height measuring instruments of two different types were on loan for the duration of the study (Figure 3.1). Each instrument records a barometrically compensated water pressure at a fixed depth; however, they have different methods for handling surface-wave contamination. The instruments at the north and middle stations were equipped with stilling wells (Figure 3.2), and recorded discrete measurements at ten-minute intervals. The third gauge, situated at the

Figure 3.1 Field study time line. The field study was conducted in the spring of 1997. Three tide gauges were each deployed for over 41 days, in excess of a full lunar cycle (~28 days). ADCP transport measurements (vertical bars) cover 4 semi-diurnal cycles (12.4 hours), 2 high amplitude tides and 2 low amplitude tides over 1/2 the lunar spring-neap cycle (14 days). The study period (grey box) is defined as the 41 days between March 7 and April 18. Wind data were also obtained for the same period.

Figure 3.2 Stilling well design. A stilling well can assist with height measurements of the tidally oscillating sea surface. With vents only at the top and bottom of the well, surface wave induced pressure oscillations do not affect the height (H) of the water column inside. The wells were mounted on private docks, in water depths of about one meter. Surface waves in the region seldom exceed 0.3 m.

south station, had no stilling well, but recorded a one-minute mean value for each ten-minute sampling interval. Both methods proved to be effective in limiting surface wave induced noise in the records. In addition, each instrument recorded the difference between atmospheric pressure and bottom pressure to remove the influence of atmospheric oscillations.

Three tide height measurement stations (Figure 3.3) were chosen with consideration of the processes to be studied, site availability and water depth. Water depths need to be such that the instrument is submerged at all times, and preferably deep enough to be below the surface-wave pressure field. In each case, permission was granted from local property owners to mount the instruments on private docks.

The gauges were deployed for 41 days, running from March 8, 1997 to April 16, 1997, well in excess of the minimum ~28 day sampling period required for a full lunar cycle (Figure 3.1). A data backup procedure was conducted once during the collection period to confirm system functionality. The ten-minute sampling interval allowed enough time to upload the data without collection interruption. All instruments functioned continuously for the entire period (Figure 3.4).

Figure 3.3 Observation stations. The SRN is located in the northeast corner of NB (inset). Observations of sea surface height were made north (N), between (M), and south (S) of the SRN constrictions. Transport observations were made at each junction between the SRN and another basin: MHB, the lower SR, and the SRN interior cove (not marked). Each transport datum constitutes the average velocity for a channel section (dashed lines), multiplied by the cross sectional area. Perpendicular transects depicted in Figure 3.10, panels e and f, were collected at each breakwater (also not marked). In addition, density profiles were collected at three locations across the channel, at each transport station.

Figure 3.4 Tide gauge time series. The three records are of tide height vs. time for the north (a), middle (b) and south (c) stations. Water depth observations (blue) lack surface wave induced high frequency noise, and clearly illustrate the semi-diurnal tide in each record. Tidal amplitude periodicity over the spring neap cycle (14 days) is apparent in the signal envelope, as is additional low frequency height variability. The mean for each data set (horizontal line) is later subtracted for station inter-comparison. Two gauges (a and b) also measured temperature (red), and show an overall increase in bottom temperature over the study period.

3.1.2 Circulation

A 1200 kHz ADCP manufactured by RD Instruments (RDI) was used for measuring transport and circulation patterns. It records current velocity by measuring the Doppler shift of sound pulses scattered off small particles carried in the flow field. The sound is transmitted and received by four differently directed beams, all angled at 20° to the vertical. The instrument listens to returns from progressively greater times, which correspond to energy scattered from progressively greater depths when aimed downward from a boat. Measurements are then averaged into prescribed depth cells, or "bins". In this study bins were 0.5 m, resulting in 0.5 m velocity depth resolution.

The ADCP may be deployed from a small boat for either time series or underway measurements (Figure 3.5). Sampling was primarily conducted while underway at a ship speed of 2 to 2.5 knots. A velocity profile is collected approximately every seven seconds, over a horizontal distance of about 10 m. Within each seven-second period, ten pings were averaged to obtain each current velocity profile datum. ADCP sampling lines typically run from one shore to the other perpendicular to the channel axis, and are called transects. For each transect, a volume transport datum is calculated from velocity, and cross-channel distance and depth measurements, all obtained from the ADCP (Figure 3.6). In

Figure 3.5 ADCP deployment. The ADCP was deployed from a 20' skiff with a specifically designed instrument harness and mount. The mount holds the ADCP out of the water for transit (pictured here), and the transducers can be rotated into the water for sampling. A 12 volt DC battery provided power, and data were logged on a portable computer. Pictured from left to right are: Diana Stram, Jim Andrews, Dr. Chris Kincaid, and Paul Hall.

Figure 3.6 Water velocity cross section. Data obtained from the ADCP constitutes the 7-second average velocity within each depth cell. For each transport datum, a section like this is used to calculate the mean velocity and the cross sectional area. The channel bottom (marked "bad") lies about 1 meter below the last pictured depth cell. This particular profile was collected at the N station just past peak flood. Positive velocities (warm colors) are northward, and negative velocities (cool colors) are southward. All profiles such as this are displayed from the viewpoint of an observer looking north.

addition, a limited number of transects were driven parallel to the channel axis to profile the bottom topography, and to sample the along channel extent of observed flow features.

CE ADCP transect lines were in general positioned at boundaries between the four basins examined: MHB, the SR, the SRN, and its interior cove (referred to as "the cove") (Figure 3.3). The cove just south of the southern breakwater was not sampled. When technical problems arose, the main channel transects were given sampling preference over the cove transect. These transects may be referred to as "primary" and "secondary", respectively. In choosing specific transect locations, consideration was given to bottom topography, travel time across the channel and between stations, and water depth. Channel cross-sections with steep sides reduce the risk of hitting rocks, and allow for the instrument to sample close to shore. Short, close together cross-sections allow for higher sampling frequencies.

rese The ability to calculate accurate volume transport is sensitive to water depth. Regions with relatively deep water provide better transect locations because the ADCP cannot accurately sample a fixed layer at the surface and bottom. Therefore, a larger fraction of the water column is sampled in deeper water. The surface is missed for two reasons. First, the transducers are typically submerged about 1 m, to ensure that they do not surface. Also, a small layer (approximately 0.5 m) is lost due to instrument-generated noise. When the instrument transmits, or "pings,"

the instrument housing vibrates for a short duration. During that time, the received signal is too noisy for accurate measurements. Since returns from the closest scatterers arrive first, the duration of this so-called "blanking period" translates into a distance from the transducer face. The bottom layer is missed due to side-lobe contamination of the main-lobe signal. The signal is contaminated once the first side-lobe generated bottom return arrives at the instrument; this duration translates into 1-2 meters.

The resolution and sampling limitations of the ADCP mentioned above are a function of the frequency at which it transmits sound. Lower frequency instruments can resolve larger depth bins, and have larger surface and bottom data gaps. But, for a given signal strength, a lower frequency instrument has better range. The 1200 kHz utilized in this study was the highest frequency ADCP produced by RDI at the time. Therefore, the instrument utilized provided the best available spatial resolution; the depths did not challenge the range of the instrument.

ADCP field days were chosen based on the stage of the spring-neap cycle, and wind conditions. Since tidal volume transport is larger during a spring tide, and smaller during a neap, data from both periods provide end-member results. Four full tidal cycles were sampled with the ADCP: two spring tide cycles (April 7, 8) and two neap tide cycles (April 15, 16), each on consecutive days (Table 3.1). Days with significant winds were

avoided as high winds agitate the sea state, and 1+ ft waves make data collection difficult and often lower data quality.

Date	April 7, 1997		April 8, 1997		April 15, 1997		April 16, 1997		Total	
	Spring Tide		Spring Tide		Neap Tide		Neap Tide		4 full tidal cycles	
Station	# visits	period (min)	# visits	period (min)	# visits	period (min)	# visits	period (min)	# visits	period (min)
North	17	43.9	19	39.3	22	33.9	21	35.5	79	37.8
South	17	43.9	19	39.3	21	35.5	21	35.5	78	38.3
Cove	2	-	18	41.5	18	41.5	16	46.7	54	-

Table 3.1 ADCP Sampling Summary. Primary stations were sampled from 17-22 times in each 12.4-hour tidal cycle, or about once every 40 minutes. The Cove station was occasionally sacrificed (April 7 and 16) in order to maintain this sampling interval.

3.2 Results

The observations made in this study consist of various data types (e.g. ADCP and tide gauge) from multiple instrument stations. Results are presented for individual data sets and from comparisons made between data sets. First order results are those reliant solely upon individual instrument measurements (Sections 3.2.1 and 3.2.2). Second order results either require multiple data types (Sections 3.2.3 and 3.2.4), or data from multiple instrument stations (Section 3.3.1). In addition, some results require previously collected data (Section 3.3.2), or are dependent on lengthy or complicated computations (Chapters 4 and 5).

3.2.1 Tidal Component analysis

The goal of this aspect of the study is twofold: to characterize tidally driven flow by analyzing the relative amplitudes and phases of the various tidal constituents from the sea surface height data, and to remove them from the record in order to investigate non-tidal aspects of the system. To remove the tidal constituents, a sinusoid was generated for each prescribed frequency, amplitude, and phase, and was subtracted from the record. Once all constituents are removed in this manner, the resulting data set is then considered “non-tidal”. The non-tidal record may then be used to address other mechanisms for sea surface variability, such as wind stress (Section 3.2.4).

Primary tidal constituents were calculated from sea-surface height data with the response method (Appendix). Admittances from the response method were used to compute an amplitude and phase for each known tidal forcing frequency. With an amplitude (A) greater than that of any other component by at least a factor of 4, the “semi-diurnal” moon tide (M_2) is the dominant oscillation (Table 3.2). In addition, its amplitude measured at the northern tide gauge station (A_N) was found to be greater than that at the other two stations (A_M and A_S), which were themselves of similar amplitude,

Tidal Component	Frequency (cycles/day)	Analysis Method	NORTH		MIDDLE		SOUTH	
			Amp. (m)	Phase (deg)	Amp. (m)	Phase (deg)	Amp. (m)	Phase (deg)
M2	1.9323	Response	0.57754	198.835	0.49403	197.525	0.49452	188.847
N2	1.8960	Response	0.13661	174.938	0.11999	169.078	0.11510	161.839
S2	2.0000	Response	0.12554	241.350	0.11687	243.519	0.11253	228.711
M4	3.8646	Harmonic	0.1012	15.550	0.0750	-4.500	0.0624	-27.799
K1	1.0027	Response	0.05694	74.087	0.05653	86.885	0.05664	69.552
O1	0.9295	Response	0.05639	90.532	0.04993	93.935	0.05394	84.800
K2	2.0055	Response	0.02984	245.901	0.02797	247.696	0.02650	232.207
P1	0.9973	Response	0.01888	86.292	0.01877	97.650	0.01868	81.821
Q1	0.8932	Response	0.01799	39.717	0.01746	42.985	0.01807	32.924

Table 3.2 Tidal constituents. Amplitudes and Greenwich phases of the principal tide constituents determined from sea surface height records in the SRN. Tidal amplitudes for the largest constituents are greater in the lower MHB (North) than the SRN (Middle) or the upper SR (South). The amplitude of the dominant semi-diurnal constituent (M2) is 16% greater in MHB. The phase of the MHB M2 oscillation differs by 10 degrees between lower MHB and the upper SR, SR high/low water occurring before that in MHB.

$$(A_N^{M2} = 0.58 \text{ m}, A_M^{M2} \cong A_S^{M2} = 0.49 \text{ m}).$$

The phase at the northern station (Θ_N) lagged that at the southern station (Θ_S) by ten degrees, and the middle station (Θ_M) by 1 degree. Figure 3.7 shows this important result is apparent in the raw data, where the north station exhibits higher amplitude oscillation, but the southern station changes direction sooner.

The next largest components were N2 and S2, with amplitudes ranging from 0.11 m to 0.14 m. In both cases, the amplitude of the northern station exceeded the other stations, which were themselves of similar magnitude, (e.g. $A_N^{N2} > A_M^{N2} \cong A_S^{N2}$). This pattern is not apparent for the lower amplitude linear components such as K1 and O1, which in general are similar for each station.

Overtides, or sea surface oscillations at frequencies harmonic to the forcing, are often observed in coastal regions (Parker 1991). Since the response method does not address overtides, the harmonic method of tidal analysis (Appendix) was utilized to investigate their influence. The harmonic method produces an amplitude and phase for each specifically selected frequency. Overtides at frequencies 2, 3, and 4 times that of the M2 were investigated, namely M4, M6, and M8 respectively. Of these, only the M4 had a significant contribution

($A_N^{M4} = 0.10 \text{ m}, A_M^{M4} = 0.08 \text{ m}, A_S^{M4} = 0.06 \text{ m}$). In addition, the phase of the M4

Figure 3.7 The semi-diurnal tide. The difference in M2 amplitude and phase illustrated by the tidal component analysis is also evident in the raw data. High water for the N station (blue) clearly exceeds the M (red) and S (green) stations. Also, high water clearly occurs at the S station first.

was found to differ amongst stations in 20-degree increments, where

$$\Theta_S^{M4} < \Theta_M^{M4} < \Theta_N^{M4}.$$

It should be noted that this process does not produce *entirely* non-tidal records. Even though the combination of the two methods, response and harmonic, is very effective in identifying the qualities of most constituents, the procedure contains some error, and is not failsafe. Non-tidal sea surface variability will introduce error, and frequency resolution is a function of record length. Also, oscillations at harmonic frequencies (Table 3.2) other than those specifically removed remain in the record. In particular, significant energy was later found at the M5 frequency (Figure 3.8).

3.2.2 Currents

Results from the ADCP surveys are presented in this section. These data are used to characterize various aspects of the flow, including: transport oscillation frequency, transport magnitude, cross-channel and vertical flow structure, and water column mixing. Since constituent analysis requires at least a 14-day record, the tidal current records were not examined for constituents. But, the current is expected to oscillate at the same frequencies as the free surface (Section 3.2.1). In this section results of the analysis are presented. A discussion of

Figure 3.8 Sea surface height power spectral density. Peaks in power spectral density identify the dominant frequency components. Presented here (solid line) is the power spectral density for the three tide height stations: north (a), middle (b), and south (c). Note that these data are plotted log-linearly. In each case, the majority of the energy is at the M2 frequency (1.94 cycles/day). The power spectral density for the tide-removed signal is also presented (dotted line). Although specifically addressed in the tide removal process, significant energy may still be found at the M4 and M6 frequencies. In addition, residual energy may be found close to the M3 and M5 frequencies, which were not addressed.

system behavior, which requires comparison between measurement stations, will be presented in Section 3.3.1.

Tidal currents between the SR and MHB are predominantly semidiurnal; there are two cycles each lunar day, ~24.8 hours (Figure 3.9). Peak magnitudes for ebb tide current velocity and volume transport were found to exceed those during a flood tide by up to a factor of two. However, the duration of the flood tide exceeded the ebb by ~2 hours. Maximum spring tide volume transport reached $1000 \text{ m}^3/\text{s}$, and current velocity exceeded 1.5 m/s in the narrowest regions.

In addition, the currents exhibit a double peak in volume transport, during each flood tide. Similar flood tides have been observed throughout NB (Turner 1984, Weisberg and Sturges 1976). But in the upper SR, the mid-flood trough is remarkably exaggerated. Mid-flood currents at all stations on all four days, came to a halt, or actually reversed for a short period. A strong double-peaked flood tide such as that observed in the upper SR suggests a highly frictional environment (Parker 1991).

Volume transport measurements based on single current meters require spatial extrapolation, usually with some assumption of velocity flow structure homogeneity. Results of this study clearly show significant variations in both current velocity and direction across the channel. During each flood, currents off the channel axis were observed

Figure 3.9 North Station transport. Transport at the north station from ADCP data taken during one tidal period (12.4 hours) on April 7, 1997 (spring tide). The sampling frequency for each station was approximately once every 40 minutes; 18-21 values over the cycle. All data sets contain the pronounced double-peaked flood current.

to flow in the opposite direction from current in the main channel flow (Figure 3.10). Ebb currents were in general more laterally and vertically homogeneous, although the turning of the tide often exhibited stratified flow (Figure 3.10). Such flow inhomogeneity would make it difficult, if not impossible, to accurately estimate net flow from a discrete velocity measurement, such as that collected with a standard current meter (see Chapter 5).

Vertical velocity measurements at each constriction suggest that the water column becomes well mixed as it passes into and out of the SRN. Vertical velocity measurements have exceeded 0.5 m/s in 1.5 m/s flow (Figure 3.10). Turbulence in the region of the breakwaters was also observed by direct visual inspection; boils of upwelled water notably damped surface waves (Figure 3.11). In addition, at times of peak flood, stationary waves often formed at the southern breakwater (Figure 3.11). Such a phenomenon is also an indication of a highly frictional environment. Effects of the turbulence on density profiles are discussed in section 4.1.1.

3.2.3 Height and current

In this section, the relationship between sea surface height and transport will be developed in order to extend our knowledge of currents

Figure 3.10 Cross-channel sections of northward velocity (a-d) and along-channel sections of vertical velocity (e-f) in the SRN. Tidal currents through the SRN were consistently inhomogeneous across the channel, and data presented here are intended to illustrate the patterns most commonly observed: the cross-channel flood current velocity field (a and b), the cross-channel ebb-flood transition velocity field (c), the cross-channel ebb current velocity field (d), and the along-channel flood current vertical velocity field (e and f). Panels to the left (a, c, e) present data from the N station, and panels to the right (b, d, f) present data from the S station. The along-channel profile at the southern breakwater (f) intersects the cross-channel profiles (b and d) at their deepest points, and the data in panels b and f were collected at the same stage of the tide. The N station profiles do not intersect (a and c with e) because the cross-channel transects at the N station (a and c) lie further away from the breakwater than at the S station (see Figure 3.3). At each breakwater, flow is restricted to the center of the channel (the middle 1/3 of the sections b and d). Core velocities mid-channel may reach 3 knots, or about 150 cm/s, often with back flow to

e.

f.

either side (b). Bottom scouring in the vicinity seems to be limited by the presence of some sort of bridge footing (the shallow regions to the right of e and f). Consequently, core vertical velocities down-stream of the constrictions have a strong negative component (e and f), as much as 50 cm/s (f) in 150 cm/s flow (b). At the N station, further downstream from a breakwater (a), core flow often hugs one shore with back flow along the opposite shore.

Figure 3.11 Surface features. At times, the state of the sea surface lends insight into the behavior of the current below. When surface features were observed in the SRN, photographs were taken for record. The most pronounced feature was localized surface wave damping caused by turbulence (a) and upwelling (b). A trail of surface boils was often observed on the down-stream side of each breakwater. The southern breakwater produced a trail at times extending 200 m down-stream (b). The remnants of a train bridge at the northern break-water (a) also produced a similar, though not as extensive, turbulent eddy field. Furthermore, 2-foot high stationary waves were observed between the southern breakwaters at peak flood on a spring tide. Stationary waves are indicative of flow over sharp changes in bottom depth. ADCP bottom profiles along the channel axis indicate a sharp decrease in depth at the breakwater, with deeper areas to either side (Figure 3.10).

beyond the limited sampling period, and also to gain additional insight into the nature of SRN tide induced wave motion. This relationship is often used to describe estuarine tidal wave motion, using progressive and standing wave types as end member possibilities. With a progressive wave, peak current velocities, and therefore peak transports, occur at peak sea surface heights. An example of this type of wave motion is a surface water wave approaching a beach; peak positive shoreward velocities occur at the crest of the wave, and peak seaward horizontal velocities occur at the troughs. With a standing wave, peak horizontal velocities are found midway between crest and trough. Standing waves are formed by the interaction of two progressive waves traveling in opposite directions. Opposing horizontal currents cancel at the crest and trough, resulting in no horizontal motion there; at these locations motion is purely vertical. Examples of this type of wave motion are found in the sloshing of water in a tub, and the seiching of a lake.

Between these end member possibilities lie intermediate stages of standing and progressive wave motions, often observed in embayments. Such wave motion results from the frictional attenuation of the tidal wave as it travels up and back down the bay. For example, at a given location in a bay, the amplitude of the landward traveling progressive wave may be greater than the amplitude of the reflected seaward traveling progressive wave. The combination of these two waves will

cause the horizontal currents to cancel sometime before or after high and low water.

Data from the north station are used as a representative measure of the system because tide gauge and ADCP transects are closely situated. The type of wave motion observed at the north station was similar to that of a progressive wave traveling from MHB south through the narrows. The first flood peak occurs just after low water, and the peak ebb currents occur just after high water; the second flood peak coincides with mid-flood (Figure 3.12).

The seemingly progressive nature of the tide likely results from the difference in tidal amplitude between MHB and the SR. Recall that the amplitude of the north station exceeded that of the others by over 8 cm (Table 3.2). Similar wave motion would be produced from the interaction of two progressive waves, traveling in opposite directions, of the same frequency but of different amplitude (i.e. a smaller amplitude wave traveling north from the SR, and a larger amplitude wave traveling south from MHB). In such a case, the resultant wave motion would be dominated by the larger amplitude wave. Since the MHB M2 tide is so much greater than that in the SR, when the two interact, the MHB M2 tide dominates. This effectively results in a progressive wave originating in MHB, traveling south through the SRN.

a.

b.

Figure 3.12 Transport (circles joined by lines) and sea surface height data (line) for spring tide (a) and neap tide (b) conditions. These data were collected at the north station. The transport between MHB and the SR oscillates as a progressive wave traveling from MHB south into the SR; peak ebb current (negative) and the first flood current (positive) peak approximately coincide with high and low water, respectively. The second flood peak occurs at mid-tide with sea level rising.

3.2.4 Wind

The wind can have a dramatic effect on transport and circulation in an estuary. When winds drive water toward land, coastal sea level increases. The change in sea level is often exaggerated at the head of an estuary where water is funneled into shallow, narrow regions. This phenomenon, often referred to as a "storm surge", is easily detectable with sea surface height measurements, but must also be understood to represent significant volume transport. The Sakonnet River, and Narragansett Bay in general, is approximately rectangular and has been known to experience such surges with the passing of a storm. In this section, the relationship between regional wind stress and sea level in upper NB is investigated by comparing observations of SRN sea surface height with regional meteorological data. Data from both T.F. Green airport and the URI GSO dock were utilized for this analysis; due to data gaps, a merging of the two data sets was required. Section 4.4 addresses the effects of specific storm events.

Wind speed and direction data were used to calculate wind stress (τ). A time series was calculated for each angular component of τ in 5° increments around the compass ($\tau_\Theta = \tau_0, \tau_5, \tau_{10}, \dots, \tau_{355}$). (Θ represents the angular direction of the wind vector, not the direction the wind is coming from.) This procedure produces the time history of wind stress

magnitude in a particular direction, such as due east (τ_{90}) (Figure 3.13).

Wind stress may be approximated as,

$$\tau_{\Theta} = \rho C_D W W_{\Theta},$$

where ρ is the density of the water, C_d is a non-dimensional drag coefficient, W is the wind speed and W_{Θ} is the wind velocity component in a particular direction Θ (Pond and Pickard 1983). For the drag coefficient, a value of 0.002 was used, which is approximately what is cited in the literature (Pond and Pickard 1983).

We choose to correlate τ_{Θ} with dH/dt , or H' , instead of H (where H represents the de-tided record) with the following reasoning. Sea surface height should relate to wind stress given sustained winds for an extended period, beyond the set-up time scale of the bay. In that circumstance the bay height would achieve a steady state, wind stress balanced against water pressure gradients (Sturges and Weisberg (1971) estimate a ½ foot increase in sea surface height over the length of the bay given sustained 20 knot winds). Shorter periods of sustained wind stress, due to variable wind speed and direction, will not produce a steady state profile and so better correlate with H' (H' was approximated with a Matlab difference function).

Cross-correlations between τ_{Θ} and H' values were generated (r_{Θ}). Highest r_{Θ} values indicate the direction (Θ) of wind stress that shows the strongest correlation between stress and height change. A positive r_{Θ}

Figure 3.13 Wind stress and wind direction. Wind direction (a) and wind stress (b-c) data. Wind stress force vectors were calculated from regional wind speed and direction data for the duration of the study. The data shown here represent wind stress directed toward 315° (b) and 45° (c), representing orthogonal components. Negative data represent directions of 135° and 225° respectively. The passing of a storm produces strong winds that rotate in direction as the storm passes. The strongest storm events during the study period are apparent on days 2, 6, 17, and 23. Wind direction during the study period is presented as a stack histogram (a). The angle represents the direction that the wind is coming from, and the magnitude represents normalized time as a percentage of the study period. Predominant wind directions are from the north and south-southwest, constituting about 7% and 5% respectively.

corresponds to a positive τ_Θ , causing an increase in sea level.

Results from this analysis indicate that wind stress directed toward 315° (Table 3.3) had the most influence on sea surface height rise during the study period, and similarly, the opposite direction, 135° , had the strongest influence on sea level fall. Each station provided similar $r_{\max}$ directions, reflecting their similar low frequency variability.

	NORTH	MIDDLE	SOUTH
Greatest (+) correlation (r_Θ)	0.65	0.55	0.45
direction (Θ)	315	315	315

Table 3.3 Wind Stress (τ_Θ) and Height Change (dH/dt) Correlation Results. Results indicate that winds directed toward 315° have the strongest influence on sea level rise.

When calculations involving wind induced embayment set-up and set-down are done, winds oriented with the up bay axis are typically considered most influential (Weisberg 1976a) if the bay is sufficiently small that Coriolis effects are insignificant. The maximum response of the SR system is not to winds aligned with the channel (0° or 180°), but at a 45° angle to the channel. This result suggests that in NB, off shore winds also play a significant role. By Ekman theory, wind induced transport in the ocean off the Rhode Island coast will be at right angles to the wind direction, easterly winds producing northward transport. The observed value of 315° is likely the cumulative wind effects over NB and RIS.

3.3 Discussion

3.3.1 Sakonnet River Narrows system

SRN current and sea surface height data from all stations were compared and contrasted in order to describe the overall systematics of SRN tidal circulation. As indicated in Section 3.2.3, a comparison of N station current time series and sea surface height suggest primarily progressive wave type motion. In addition, a comparison of current oscillations between the N and S transport stations also suggest a progressive component to the tide (Figure 3.14), with currents at the north station often leading those at the south station. But, in contrast, height difference data between stations ($\Delta H_{SM} = H_S - H_M$) illustrate that peak velocities coincide with peak height differences (Figure 3.15), often a signature of standing wave type motion.

These conflicting observations suggest that the simple progressive and standing wave models are not adequate to describe flow within the SRN. This is likely the result of friction-induced turbulence related to the presence of the breakwaters (Figure 3.10 e and f, and Figure 3.11). Turbulent flow has an increased eddy viscosity over laminar flow, and the standing and progressive wave motion models are only for inviscid fluids. Although values of eddy viscosity were not calculated from the

Figure 3.14 Sakonnet River Narrows transport. Transport curves for spring tides (a and b) and neap tides (c and d) are presented to examine the relative timing of the transport between stations. Transport between MHB and the SRN (circles) leads transport between the SR and the SRN (squares) in most cases.

Figure 3.15. Height difference and transport.

The difference (SRN-MHB) in water height (dashed curve) between MHB (solid curve) and SRN (dotted curve) is remarkably similar in waveform (a) to the observed transport (open circles). The largest differences occur at high and low water, and reflect the difference in tidal amplitude between the two regions. A similar height difference curve may be created with generated sinusoids at the prescribed amplitudes and frequencies (b-e). The M2 and M4 constituents (b) were used to generate curves for the N station (solid curve) and M station (dashed curve) tides (c). The difference in height from these two curves (d) also exhibits a double peaked flood. Increasing the M4 amplitude by a factor of two further accentuates the inflection (e).

data, vertical velocity data and direct observation clearly illustrate the turbulent flow. (Additional effects of the friction-induced turbulence will be presented in Section 4.1.1.) Disregarding the progressive and standing wave models, a descriptive conceptual model of flow between MHB, the SRN and the lower SR over a tidal cycle will be presented. The model reflects our observation that tidal circulation within the SRN is dominated by the MHB M2 tide and the presence of the SRN breakwaters.

Tidal currents observed between the north and south stations are about 10 minutes out of phase, or about 5° at the M2 frequency (Figure 3.14) (estimated shallow water wave travel time is about 4 minutes). Currents turn from flood to ebb and ebb to flood first at the north station, then at the south station (tidal currents at the cove's mouth turn approximately one hour later). In other words, the tidal current between MHB and the SRN "leads" the current between the SRN and the SR.

Figure 3.16 contains three illustrations to help describe the direction and duration of the observed tidal current under both spring tide (Figure 3.16 j) and neap tide (Figure 3.16 k) conditions. The period of time represented by each illustration is one M2 cycle, or 12.4 hours. "Full flood" and "full ebb" describe periods where all observed current directions are in agreement, either all in or all out, respectively. The time span between the turning of the current at the north station and that at the cove station is referred to as a "transition period". The approximate

Figure 3.16 SRN flow patterns over a tidal cycle. SRN tidal flow direction (a-i) and duration (j-k) are presented here. Panels (a-i) proceed from conditions of full ebb (a) through a full tidal cycle and back to full ebb (i), illustrating the direction of flow. Charts (j-k) illustrate the duration associated with each panel for spring (j) and neap (k) conditions; the full pie represents one semidiurnal cycle. Note that the current turns between MHB and the SRN (b and g) before turning between the SR and the SRN (c and h). Transport in and out of the interior cove follows (d and i). Panel (e) represents the "mid-flood slack", where flow either stopped or reversed slightly at each observation station.

duration of the lull in transport observed at mid flood is referred to as a “mid-flood slack”. Figures 3.16 b-d depict the flow observed as the tide turns from ebb to flood; the current is shown to turn first at the northern breakwater, then at the southern breakwater, and then finally at the mouth of the cove. Figures 3.16 g-i depict a similar flow pattern for the flood to ebb transition.

Some physical characteristics of the SRN breakwaters not directly obvious in map view may lend insight into the cause of the anomalous flow patterns described above. It is clearly obvious from Figures 1.1 and 1.2 that the SRN is a narrow region between two basins, and that the breakwaters further restrict the passage. However, ADCP bottom mapping shows that it is even more restricted in cross section, both laterally and vertically, below the water line. Figure 3.10 b and d illustrate the shape of the bottom in the cross-channel direction at the southern breakwater, and Figure 3.10 e and f illustrate the shape of the bottom in the along channel direction for the north and south breakwaters, respectively (a cross channel bottom profile for the northern breakwater was not obtainable due to obstructions mid channel, see Figure 3.11 a). Evident to the right side in each along channel profile is a very shallow region (~10 m). This feature lies in line with the breakwater to either side, and is believed to be some sort of footing associated with the now out-of-use bridges. Due to this so-called footing and to additional breakwater below the water line, the resulting cross-

section area allowing flow between the lower SR and the SRN is approximately 3000 m^2 (300 m wide and 10 m deep); the cross section between MHB and the SRN is believed to be of similar magnitude.

An argument can be made that if a natural system is constricted laterally, it will tend to maintain cross-sectional area due to scouring of the bottom. In the SRN, scouring seems to be limited by the so-called bridge footings. While scouring is evident just upstream and downstream of the southern breakwater footing (Figure 3.11f), it does not serve to change the cross-sectional area of the narrowest region. Therefore, it is believed that the SRN breakwaters indeed restrict flow and are a large source of friction.

Perhaps some insight might be gained by the following thought experiment. Consider a dam, which separates a canal from the sea. When the dam is closed, there is no flow, and there is a large height difference across it; the water rises and falls with the tide on one side and on the other side it remains dry. If we open up a thin section of the dam from top to bottom, there would be flow, but it would be too small to have a significant effect on water level in the canal. In addition, the sea surface would be nearly discontinuous, as water cascades between levels. If we open the dam further, the flow increases, and slope of the sea surface decreases. If we open it up all the way, there is no dam, flow is uninhibited, and the sea surface slope becomes small. In the case of the SRN, the dam is opened up about 25%; note that this value of 25% is

a *further* restriction solely due to the presence of the breakwaters (Figure 3.3).

Observations of surface height difference across the SRN breakwaters are also an indication of high friction and restricted flow. Between the north and middle stations a difference in height of 15 cm was commonly observed, and at times reached 30-40 cm (Figure 3.17). These constriction-induced sea level elevation gradients are in excess of those expected in NB from wind events (e.g. ~15 cm height difference over entire length of NB for a 20 knot wind, Weisberg and Sturges 1976). Furthermore, the SRN is particularly dramatic in that a sharp difference in sea surface height (~1-3 cm) was visually apparent across each breakwater (>1 m length scale).

In summary, turbulent flow, tidal amplitude differences, tidal current phase data, bottom scouring information, and instantaneous sea surface height difference data, all indicate that flow constriction by the SRN breakwaters causes high current shears resulting in strong frictional effects. Let us now address the original hypothesis: anomalous circulation patterns previously observed within the upper SR are the result of local as opposed to regional factors. As shown here, the phase difference of the tide between the EP and SR inlets to MHB is directly related to the presence of the SRN breakwaters. Therefore, the observations and deductions of this work support the original hypothesis.

Figure 3.17 Height difference over the study period. MHB height was subtracted from SRN height revealing semi-diurnal oscillations, and also lower frequency amplitude oscillations; spring tide height differences (around day 32) exceed neap tide values (around day 39) by a factor of two. Data for days 2, 6, and 23 illustrate periods of prolonged height difference. Note that these same days were identified as days with storm events (Figure 3.13).

3.3.2 Mount Hope Bay system

In recent years, concerns have been raised about the influence of power plant effluent emissions into MHB. The Brayton Point Power Plant utilizes a large volume of MHB water to cool its machinery, 2-8 times the Taunton river fresh water input (Gibson 1996), and discharges the heated water back into the bay. The ecological impact of this effluent plume is dependent on the residence time and mixing of the heated waters. Although this study was not specifically designed to address the issue of MHB flushing, some insight may be gained by examining data from this study combined with previously collected data.

Mount Hope Bay is flushed with seawater through two channels, the EP inlet and the SR. Previous spring tide volume transport data collected at the entrance to each channel (Kincaid 1996), and spring tide data collected for this study (4/7 and 4/8), are plotted together in Figure 3.18. Transport between the two inlets is clearly dissimilar. The magnitude of the EP transport is 6-10 times that of the SR. This would suggest that the EP provides the dominant tidal influence in MHB. Furthermore, transport through the EP inlet oscillates as a standing wave, with peak ebb flow about three hours ($1/4$ cycle) after high water. But, as discussed previously, in the SR inlet peak ebb flow occurs at

Figure 3.18 Mount Hope Bay tidal flushing. Spring tide volume transport data from this study (squares 4/7 and triangles 4/8) and from Kincaid (1996) (circles). All SR data presented here were collected at the same location (the North station of this study). Time is given in units of 1/2-lunar day (i.e., 12.4 hours), with times 0 and 1 being high water. Note that these data are wrapped to span more than one tidal cycle. All SR data (open symbols) are combined and fit with a weighted curve, and EP data (closed circles) are connected with straight lines. The SR transport data cluster tightly around the fitted line. The SR volume transport leads that in the EP by almost 2 hours. EP transport exceeds SR transport by almost a factor of 10.

least two hours before it occurs in the EP inlet and more resembles the motion of a progressive wave.

The effects of the phase difference between the EP inlet and the SR inlet on the flushing of MHB remains unclear. Future investigators are advised to consider the interaction of the two inlets for MHB residence time estimates.

3.4 Conclusions

The tide has the strongest influence on circulation and exchange in the upper SR. Sea surface oscillations are dominated by the semi-diurnal moon tide (M2). The M2 amplitude is about 0.5 m, and exceeds all other components by at least a factor of four. The next largest tidal components were the N2, S2, and M4, with amplitudes ranging from 0.06 m to 0.14 m. Although tidal currents between the SR and MHB are also predominantly semidiurnal, the currents exhibit a double-peaked flood and a single-peaked ebb indicative of a significant M4 component. Therefore, consideration of the influence of overtides is required for correct interpretation or prediction of the tidal currents in MHB and the upper SR.

The combined ADCP and tide gauge measurements illustrate that peak ebb occurs shortly after high water. Peak observed ebb current

velocity exceeds peak flood velocity, and the duration of the ebb is shorter. Maximum observed spring tide velocity exceeds 1.5 m/s in the narrowest regions, corresponding to a volume transport of about 1000 m³/s. Furthermore, the observed tidal current leads from north to south through the system: flood and ebb currents are observed earlier between MHB and the SRN than between the SR and the SRN.

At any given cross-section of the channel, velocity magnitude and direction were often found to be inhomogeneous. Vertically stratified flow was observed at transitions between tides, and horizontally stratified flow was observed along the shore during the flood. Based on these observations, point current measurements taken in this region would be difficult to interpret.

The SRN breakwaters have highly frictional effects and restrict flow between MHB and the lower SR. Flow within the channel in the vicinity of the breakwaters is turbulent. The turbulence is believed to result in increased eddy viscosity, which causes the flow to deviate from simple standing and progressive wave models.

The tidal flushing of MHB is dominated by EP standing wave motion. Furthermore, exchange between MHB and the SR passage is mainly controlled by the oscillation of the MHB free surface, rather than the oscillation of the lower SR free surface. This results from the larger tidal amplitude in MHB and the segregation of the lower SR from MHB due to the presence of the SRN breakwaters.

The winds also seem to have a strong influence on circulation in the study area. Results show that the wind stress component directed 315° (135°) had the most influence on sea surface height rise (fall) during the study period. This direction is 45° off the channel axis, possibly indicating the influence of off shore winds on NB set-up.

4 Empirical model

Overview

This chapter details an empirical volume transport model derived from observations made during this study. An empirical equation for transport as a function of pressure excess is derived from four tidal cycles of coincident sea surface height and transport measurements. With the establishment of a model such as this, short ADCP surveys may be used effectively to “calibrate” tide gauges. Tide gauge data may then be utilized to predict flow. Section 4.1 presents support for the general approach, including results from a diagnostic test developed by Vennell (1998). Section 4.2 presents results from both linear and polynomial models, and further develops equations for a weakly non-linear model. In Section 4.3, model predicted transport is compared with observed transport, and a 41-day predicted transport record is created. In Section 4.4, storm events are investigated as a cause for low frequency variability apparent in the 41-day record. Conclusions drawn from the study are summarized in Section 4.5.

4.1 Introduction

Types of estuarine circulation models include those that solve the equations of motion, often referred to as computational or numerical models, and empirical models. Numerical models require estimates of many parameters, some of which are difficult to measure directly. The spatial and temporal variability of density, and the influence of friction are two examples. An empirical model utilizes equations derived directly from observations. There are strengths and weaknesses to both approaches. An empirical model will often replicate the system well under the conditions observed when it was calibrated, but will not when conditions are variable. In turn, the numerical model may be accurate for a broad range of conditions, but lack the precision of an empirical model under some conditions. An empirical model is by no means void of assumptions though, and in this section, the assumptions surrounding an empirical volume transport model for the SRN region will be investigated. Two primary assumptions were made; the first relates to density homogeneity within the system (Section 4.1.1), and the second relates to transport homogeneity (Section 4.1.2).

4.1.1 Baroclinic effects

The empirical model developed here utilizes two point measurements of bottom pressure to estimate the along channel pressure gradient. Pressure excess is then used to predict volume transport. It is assumed that there is no significant height difference across the channel. Justification for this lies in the very small ratio of channel width to tidal wavelength. But, there must also be no significant density variability. The flow must be considered primarily barotropic; that is, contours of constant pressure are parallel with the sea surface. Consequently, density information is required.

Mixing of the water column serves to minimize baroclinic effects within the system. Strong vertical mixing is recorded in the SRN system with both the CTD and the ADCP. Water column density profiles were collected at each ADCP station. Due to instrument malfunction and noise, "clean" or uncontaminated data were successfully collected only during ebb currents of one tidal cycle. Nevertheless, results from the survey were sufficient to show that water passing from MHB through the SRN becomes well mixed. Figure 4.1 shows that water moving from MHB towards the SRN is vertically stratified, with a roughly 1.5° C vertical temperature difference. The warm surface water is characteristic of MHB and the Brayton Point

Figure 4.1 Ebb tide temperature and salinity profiles. These graphs contain CTD temperature (squares) and salinity (circles) profiles collected in the upper SR. Each set of figures, upper and lower, constitutes a progression of profiles over the course of an ebb tide: early ebb (a, d), mid ebb (b, e), and late ebb (c, f). The upper set (a-c) was collected at the N ADCP station, and represents water moving from MHB into the SRN. The lower set (d-f) was collected at the S ADCP station, and represents water moving from the SRN into the SR passage proper. Evident in the upper set is the development of a fresh warm surface layer. The lower set illustrates that the stratified water column becomes well mixed by the time it exits the SRN.

thermal plume. After passing into the SRN, the water column has been mixed such that both temperature and salinity profiles are vertically uniform; this mixing is the result of extreme vertical velocities (25-50 cm/s) observed with the ADCP at each breakwater (Figures 3.10 and 3.11). An important implication of this result is that thermal energy from the Brayton Point plume is mixed deeper into the water column. Under conditions when the net non-tidal flow of SRN bottom water is northward into MHB, this will contribute to the trapping of thermal energy within MHB.

4.1.2 Channel Shortness

The second assumption is that the volume transport at each cross section within the channel is similar. In other words, tidal currents are approximately in phase and of similar magnitude. As with the pressure data, this assumption is required due to the logistics of the measurements: a sea surface gradient is established based on the height difference between *two* locations, and the transport is measured at *one* section between. The transport at that one section must therefore be representative of the transport within the channel, requiring the transport at the ends of the channel to be similar.

The transport will vary between sections if the cross-sectional area varies significantly with movement of the free surface. To account for this, the section between the N and M stations was chosen due to the lack of coves and tidal flats in the region. The transport will also vary if the phase of the tidal current varies significantly within the channel; given a short enough channel section, it is reasonable to assume that the currents are “in phase”. Vennell (1998) has established criteria for the shortness of a channel necessary to satisfy this requirement. With Vennell’s method, the equations of motion are scaled to establish which terms dominate the flow. The greatest term is then used to assess the scale of variability, characterized by a shortness parameter (ϵ).

The following equation represents the along channel shallow water momentum balance:

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} - fv = -g \frac{\partial \eta}{\partial x} + \frac{1}{\rho} \frac{\partial \tau}{\partial z} \quad (4.1).$$

Terms on the left of (4.1) represent the acceleration, the advective acceleration in three dimensions, and Coriolis acceleration respectively. Variables u , v and w are velocity components in x , y , and z respectively, f is the Coriolis parameter, and t is time. Terms on the right of (4.1) represent pressure gradient and friction. Variable η is free surface displacement, τ is the horizontal shear stress, and g and ρ are gravity and density, respectively.

Scales are derived primarily from channel geometry, oscillation frequency, and velocity magnitude. Scales based on data obtained in this study include velocity amplitude (u_0), and sea surface displacement amplitude (η_0) and phase (ϕ) at the channel ends. The velocity amplitude is estimated from peak observed current velocity, approximately 1 m/s. Sea surface amplitude ratios ($a_0 = \eta_{0A} / \eta_{0B} \leq 1$) and phase differences ($\phi_0 = \phi_A - \phi_B \geq 0$) are calculated for the observed semi-diurnal (M2) tide.

For a “short” channel, the shortness parameter (ε) must be very small ($\varepsilon \ll 1$). For tidal oscillations within a channel, this value is effectively the ratio of tidal prism in the channel (m^3) to net transport integrated over half a tidal cycle (m^3),

$$\varepsilon = \frac{W_0 L_0 2\eta_0}{W_0 H_0 u_0 (T_0 / \pi)} \quad (4.2).$$

The tidal prism represents the water volume change resulting from the change in surface height between low and high water. It is calculated from mean channel width (W_0), channel length (L_0), and the surface height difference between low and high water ($2\eta_0$). The along channel transport is calculated from mean channel width, mean channel depth (H_0), velocity amplitude and tidal period (T_0). With a short channel (L_0), ε is very small and most of the volume flux contributes to transport rather than surface elevation increase. When most of the volume is being transported downstream, transport at the channel ends will be similar.

Results from the analysis using data from station N and M indicate that the SRN system may be considered short. All calculated values of ε for the SRN system (0.015, 0.025, and 0.026) satisfy the requirement of being much less than unity (Table 4.1). The phase of the current ($2 \tan^{-1} \varepsilon$) is expected to vary less than 3 degrees between the S and N stations. In particular, the NM section has the smallest ε value, and currents at section ends are expected to be about 1.72 degrees out of phase. Values for two short hypothetical systems, as presented by Vennell (1998), are provided for comparative purposes.

A “short” channel indicates that transport at one section is representative of the transport throughout the channel. In addition, as previously illustrated, the flow is considered primarily barotropic. These two criteria being met, the phase of the transport is dependent only on the relative amplitude and phase of the surface oscillation at the ends of the channel (Vennell 1998). In the next section (4.2), empirical equations relating the transport and surface oscillations will be developed.

4.2 Methods

The model developed in this study uses four tidal cycles of tide gauge and ADCP data to establish a relationship between pressure excess and volume transport. Data from two gauges are used to

	A	B	NM	MS	NS
Scales					
Period (days)	0.52	0.52	0.52	0.52	0.52
Channel length, L_0 (m)	1000	1000	1500	2100	3600
Channel width, W_0 (m)	500	500	500	1000	700
Channel depth, H_0 (m)	20	20	10	8	9
Drag coefficient, c_d	0.0025	0.0025	0.0025	0.0025	0.0025
Coriolis parameter, f_0 (s^{-1})	1.0E-04	1.0E-04	1.0E-04	1.0E-04	1.0E-04
Velocity amplitude, u_0 ($m\ s^{-1}$)	1	1	1	1	1
Surface amplitude ratio, α_0	1	0.98	0.85	1.00	0.86
Surface phase difference, ϕ_0 ($^\circ$)	1	0.0	1.3	8.7	10.0
Calculated values					
Angular frequency, ω_0 (s^{-1})	1.4E-04	1.4E-04	1.4E-04	1.4E-04	1.4E-04
Wavenumber, k_0 (m^{-1})	1.0E-05	1.0E-05	1.4E-05	1.6E-05	1.5E-05
$k_0 L_0$	0.01	0.01	0.02	0.03	0.05
Size of terms					
Acceleration term (1)	0.006	0.005	0.003	0.007	0.013
Advection term ($u_0 L_0^{-1} \omega_0^{-1}$)	0.041	0.036	0.015	0.025	0.026
Rotation term ($f_0 W_0 L_0^{-1} \omega_0^{-1}$)	0.002	0.002	0.001	0.002	0.002
Bottom friction term ($c_d u_0 H_0^{-1} \omega_0^{-1}$)	0.005	0.004	0.005	0.016	0.026
ε	0.041	0.036	0.015	0.025	0.026
$2 \tan^{-1}(\varepsilon)$, ($^\circ$)	4.70	4.13	1.72	2.87	2.98

Table 4.1 Characteristics of channels determined short. Columns A and B contain examples reproduced from Vennel (1998) for comparison. Column labels NM, MS, and NS signify channel sections between each pair of tide gauges deployed within the SRN. Surface amplitude ratios were calculated for the M2 tide (see Table 3.2). In computing ε , a value of $\eta_0 = 0.5$ was used (Table 3.2). Listed equation-of-motion terms are unitless, each divided by the $\partial u / \partial t$ scale, $\omega_0 u_0$. Results indicate that each channel section within the SRN qualifies as short ($\varepsilon < 1$), and the difference in tidal current phase, or phase lag, between channel ends ($2 \tan^{-1} \varepsilon$) is expected to be less than 3 degrees. Although the observed phase lag was greater, 4.8° (Section 3.3.1), the observed lag is not much in excess of Vennel's A and B examples, 4.13° and 4.70° .

calculate a lateral pressure gradient between the ends of a channel. Once the relationship between pressure gradient and volume transport is established, volume transport through the channel is then predicted from observed pressure gradients for the duration of the tide gauge deployment, 41 days. This section first addresses the relationship between pressure and transport, and the degree to which the system behaves linearly (4.2.1). The transport-to-pressure relationship is then presented as a function of tidal range, and termed “weakly non-linear” (4.2.2). Model agreement with observed transport is addressed in Section 4.3.

4.2.1 Pressure and transport

Although the relationship between pressure excess and induced current velocity is inherently non-linear, the non-linearity may not be discernible above the noise level. In this section, the degree of non-linearity observed in the system is investigated by fitting the data with various models. The best model will be further developed in Section 4.3.2.

The observed difference in sea surface height (ΔH_0) between two tide gauge stations is used as a proxy for the pressure gradient through the hydrostatic relationship ($\Delta p = \rho g(\Delta H)$), and is plotted against volume

transport (Q) (Figure 4.2). Height difference ($\Delta H_o = H_{SRN} - H_{MHB}$) is defined such that $\Delta H_o > 1$ produces positive, or northerly, volume transport. Since H and Q are both functions of time, this plot requires coincident data. To this end, height data collected every ten minutes were interpolated at one-minute intervals. The interpolated H value which was closest to the mid-point of an ADCP transect was chosen for each volume transport collection time (~40 minute intervals). Timing errors in this analysis are controlled by the duration of individual ADCP transects (≤ 10 minutes).

A linear and polynomial curve fit for all data, and linear regressions for each individual field day's data, are also illustrated in Figure 4.2. For the linear regressions, the equations take the form of

$$\Delta H_o = mQ + b \quad (4.3),$$

with slope m (m^2s) and y -intercept b (m). The polynomial curve is of degree three, and takes the form $\Delta H_o = M_0 + M_1Q + M_2Q^2 + M_3Q^3$. Results in Figure 4.2 show that each case exhibits a non-zero y -intercept, such that $\Delta H_o = 0$ does not coincide with $Q = 0$. For zero transport, with both models there is a positive height difference between stations of approximately 5 cm. This unrealistic relationship could result if the demeaning of the sea surface height data did not serve to level the tide gauges to a common geopotential surface. If this is in fact true, the

Figure 4.2 Pressure excess to transport models. Each figure plots coincident transport (Q) and surface height difference (ΔH_0) data, where ΔH is a proxy for pressure excess. The complete data set (a) was fitted with a linear regression and a third order polynomial, producing similar results. Each tidal cycle of data is plotted separately (b-e) with a linear regression. Non-zero intercepts evident in each plot are believed to result from a de-meaning error in the height difference data.

deviation of b from zero actually provides a correction to the ΔH_o data set.

The ΔH_o data represent the time varying relative height of the sea surface between stations. The height of the sea surface is calculated from pressure (p) measured at depth. The pressure data set may be represented as $p(t) = p_d + p_\eta(t)$, where p_η is the tidally oscillating pressure, and p_d is a constant derived from the depth of the sensor. The constant p_d has a different value for each station. Therefore, in order to compare the tidal signal between stations, p_d must first be removed. Since there is no simple way to establish the exact depth of the instrument sensor, p_d was estimated based on the mean value of the data. The oscillating pressure field is then $p_\eta(t) = p(t) - \bar{p}$.

If removing the mean from each data set does not effectively level the data sets to the same geopotential surface, then p_η will be in error by a constant. The y -intercept values are likely a reflection of this error for the following reasons. The values represent a small fraction of the observed variability in water depth, making them a reasonable range for error. In addition, the y -intercept values have a tight distribution (ranging from 3.5-5 cm). If their non-zero values were purely the result of the de-meaning error, they should all have the same value. Under this assumption, the average intercept, $\bar{b}$, will be used as an adjustment for the ΔH_o data set ($\Delta H = \Delta H_o - \bar{b}$).

The quality of the models may be quantified by the Pearson's R^2 (KaleidaGraph 1996). For the combined data set, both linear and polynomial models produce similar results (Figure 4.2a), accounting for about 75% of the variability ($R^2 \cong .75$). For the individual days, three of the four linear regressions do a better job at representing the data ($R^2 > .80$), while the fourth is rather poor ($R^2 \cong .40$). The similarity between the linear and polynomial models indicate that the relationship is only "weakly" non-linear. Noise is to a large extent hiding the nonlinear component and making the distribution seem almost linear. Since a simple linear model can produce comparable results to the higher order polynomial model, we focus our analysis on the linear model.

4.2.2 A weakly non-linear model

In this section, the individual, corrected, linear trends (Figure 4.2, b-e) will be further investigated, focusing on the slopes of these trends. The slope of the lines (m_n , where $n=1:4$ for the four field days) is the ratio of ΔH_O to Q . Some factors influencing the relationship, such as friction, are a function of water speed, which varies over a tidal cycle and over the spring neap cycle. This implies that the best model for these data would have coefficients that vary with each of these factors. Since the polynomial fit to the complete data set has only a 2% larger value of R^2 ,

only linear fits are used for the individual tidal cycles to investigate spring-neap variability.

In an attempt to account for effects of the spring-neap cycle, we utilize the relationship between slope values (m_n) and tidal range (r_n). Values for m_n with high R^2 values, 4/7, 4/8, and 4/16, were plotted against tidal range (Figure 4.3). A linear regression was calculated, where tidal range (r_n) is related to m_n with a slope (m') and y-intercept (b'), $r_n = m'm_n + b'$. Tidal range for the entire study period (r) was calculated from sea surface height data, and smoothed (Figure 4.4). Substituting r for r_n and $m(r)$ for m_n , and solving for $m(r)$, yields, $m(r) = (r - b')/m'$. By rearranging (4.3), the transport for a given day may be represented as $Q_n = (\Delta H_o - b_n) * m_n^{-1}$. Substituting $\bar{b}$ for b_n , then $\Delta H(t)$ for $\Delta H_o - \bar{b}$ and $m(r)$ for m_n , yields transport as a function of time and tidal range, $Q(r, t) = \Delta H * m^{-1}$. Or more simply,

$$Q(r, t) = \Delta H(t) * C(r) \quad (4.4),$$

where $C(r) \equiv m^{-1}(r)$, and has units of m^2s^{-1} . This equation describes a linear relationship between ΔH and Q , which oscillates with tidal range

Figure 4.3 Tidal range dependence. The tidal range values (r_n) are plotted against m_n ($m_n = \Delta H_O/Q$ value for a particular day, n). The linear trend (for days 1, 2 and 3), $r_n = m'm_n + b'$, is used to calculate a relationship between ΔH and Q that varies with r . If no discernment is made between the spring tide ($n = 1, 2$) and neap tide ($n = 3, 4$) results, the m value for the combined data could be used (Figure 4.2), representing tidal ranges from approximately 0.75 to 2 meters. Note that the value for April 15 (open square) is not included in the linear trend but is included in the combined data set.

Figure 4.4 Tidal range. Tidal range (r) was calculated for the entire tide gauge deployment period of 41 days (a) from sea surface height data ($r = H_{\text{MAX}} - H_{\text{MIN}}$ for each cycle). These data were passed through a low pass filter (b), and the smoothed tidal range curve is utilized as input to the empirical model.

(τ), and hence lunar phase (Figure 4.4). The two sets of field days, April 7-8 and April 15-16, provide data coverage for extremes in tidal amplitude (Section 3.2.2). By presenting the relationship in this way (4.4), the model takes full advantage of the spring-neap transport measurements coverage.

4.3 Results

Model precision may be quantified by the root mean square difference (σ) between observed (Q_o) and predicted (Q_p) transport:

$$\sigma = \sqrt{\frac{\sum(Q_p - Q_o)^2}{n}} \quad (4.5).$$

Values of σ for the four field days were all between 241 and 268 m³/s, approximately 25% of peak observed transport.

The model was used to predict volume transport for the entire tide gauge deployment period of 41 days. Tidal oscillations are apparent in the signal, with a double peaked flood especially pronounced between days 25 and 35 (Figure 4.5a). But there is also significant low frequency variability. The dominant low frequency feature predicted by the model occurs between days 20 and 25. During this period, the model predicts that water will flow from MHB into the SRN continuously for a duration

Figure 4.5 Model predicted and observed northward transport through the SRN. Tidal oscillations are obvious in the model predicted transport record (a). In addition, a double peaked flood is often apparent, especially between days 25 and 35. The average of the rms differences for the four volume transport collection days ($\sim 250 \text{ m}^3/\text{s}$) is plotted as error bars for the daily low-pass filtered predicted transport (b). Observed spring (c) and neap (d) tide transport (dashed line) is also presented for comparison, with the four rms differences ranging from 241 to $268 \text{ m}^3/\text{s}$. Anomalies in the filtered transport record often exceed the standard deviation, most clearly observed on day 23, although most days exhibit negligible or slightly southerly net flow.

of three full tidal cycles; although the speed of the current still oscillates with the tide, flow direction is persistently south. Also note that the magnitude of this feature (up to $1000 \text{ m}^3/\text{s}$) exceeds the root mean square (rms) difference σ between the model and observation ($\sim 250 \text{ m}^3/\text{s}$) by a factor of four.

The data were passed through a low pass filter with a cutoff at the lowest measured tidal frequency (Q1), about 1 cycle/day. Therefore, variability remaining in the record may be considered primarily non-tidal (Figure 4.5b). Again, the period between days 21 and 24 indicates prolonged negative net transport, or transport from MHB into the SRN. Deviations from zero in non-tidal transport signify net transport over a tidal cycle, sometimes referred to as “residual flow”. The term residual flow is often used to refer to fresh water input, but here it is not specifically process dependent. Since the magnitude of the anomalous flow is comparable to observed tide induced flow ($>1000 \text{ m}^3/\text{s}$) and 50 times larger than the normal river input to MHB (Section 2.1), and the duration spans multiple tidal cycles, fresh water input is not a reasonable explanation.

4.4 Discussion: Storm events

Previous studies in NB have stressed the importance of the winds on NB low frequency transport. Model results are therefore used to

estimate the magnitude of wind-induced transport for the SRN. It has already been shown that the wind stress (τ) correlates with sea surface height change (dH/dt) (Section 3.2.4). In particular, it was shown that winds oriented $315^\circ/135^\circ$ have the most influence on sea surface height change in the SRN system. Therefore, the same wind directions are expected to produce the most wind-induced transport.

Peaks in the wind stress record identify storm events (Figure 4.6). As a storm passes through the region, winds often swing from southerly to northerly, as seen on days 2, 6, 17, and 23. The onset of each of these storm events is accompanied by a negative (southward) transport anomaly. This implies that winds out of the south force water through the EP into MHB, and from there it then flows into the SRN. The tail ends of these storms, where winds are out of the north, coincide with a rebound in the transport signal. The most prominent deviation from this pattern occurs on day 36, where an inverse relationship is evident. Here a negative peak in wind stress accompanies a negative transport anomaly.

The observed relationship between southerly winds and transport is counterintuitive in that the *localized* effect of winds out of the south should be to force water northward. But, as suggested in section 3.2.4, the winds may have a more *regional* effect on transport in the bay. Surface height data in general agree with the concept of water being

Figure 4.6 SRN wind induced set-up and predicted transport.

Panel (a) exhibits wind stress (toward 315 degrees) vs. time, (b-d) are sea surface height vs. time data for the north (b), middle (c), and south (d) stations, and (e) is the predicted transport related to the height difference between the north and middle stations. Storm events are clearly apparent on days 6, 17 and 23 (a). Peaks in non-tidal sea surface height are observed at each station (b-d) following each of these events, suggesting that the storm events induced NB set-up. In addition, predicted transport (e) illustrates negative peaks in transport following each event. Negative transport describes transport from MHB into the SRN.

Figure 4.7 Rhode Island Sound wind induced transport.

Wind stress causes the sea surface height of NB to vary at sub-tidal frequencies. Winds out of the SE (black vector in a) were found to have the greatest influence on sea surface rise rates. Such winds cause the sea surface at all stations to rise, implying

that water is being transported from RIS into NB (green vector in a). For each set of stations, MN (d), SN (e), and SM (f), height difference curves were averaged over a tidal cycle to examine the sub-tidal variability. The SN curve (e) is relatively flat, indicating that the SR and MHB vary similarly at sub-tidal frequencies. The other two curves (d and f) vary wildly, and are inverses of each other, implying that the M station introduced the variability. For example, given a set-up event, the SRN is not set up as much as the SR or MHB, which are set-up similarly. Therefore, wind induced setup of NB is believed to cause net transport into the narrows from both sides (c). A similar phenomenon was observed given a wind related reduction in NB surface height; net transport is out of the SRN from both sides. Model predicted transport values associated with these sub-tidal height difference anomalies are of the same magnitude as observed tide induced transport values (Figure 4.6).

pushed north by southerly winds, causing the sea surface at the head of NB to rise (Sturges and Weisberg 1971). Storm events on days 6, 17, and 23 accompany positive sea surface height anomalies ($H > \bar{H}$) at each tide gauge station (Figure 4.6); the day-2 event shows no change. The flow direction anomaly only appears in the ΔH record. Storms with strong southerly winds seem to cause NB to be set-up, as expected, but the response of the SRN is damped relative to MHB and the SR (Figure 4.7). This seems to explain the high magnitude ($>1000 \text{ m}^3/\text{s}$) anomalous flow predicted during periods of strong southerly winds.

4.5 Conclusions

Observations from a 40-day field experiment, including both tide gauge and ADCP measurements, were used to establish a relationship between volume transport and height difference between MHB and the SRN. A linear relationship accounted for at least 75% ($R^2 > 0.75$) of the variability, matching results produced with a third order polynomial model.

A transport model was developed. It consists of a linear relationship between height difference and transport, which is modified by a coefficient that varies with tidal amplitude. Model predicted volume

transport agrees with observed values to within approximately 25%

($\sigma \cong 250 \text{ m}^3 \text{ s}^{-1}$, and $Q_{\max} \cong 1000 \text{ m}^3 \text{ s}^{-1}$).

By calibrating the model during periods when both ΔH and Q data are available, we can extend the model to predict a 40-day time series of Q within the SRN system. A primary goal is to characterize how the SRN responds to wind forcing. Normally low frequency flow between MHB and the SRN predicted with the model is southerly and of low magnitude. But at times significant low frequency variability exists. The largest of these anomalies predicts a period where tidal influence in the SRN is overridden for multiple consecutive tidal cycles.

In most cases, strong southerly winds correspond to NB set-up events. But, results from this study indicate that net transport between MHB and the SRN during NB set-up events is southward. This result is in disagreement with small-scale wind stress forcing, which would push water from the SRN north into MHB. Clearly regional wind stresses, rather than local effects, are controlling the flow.

5 Residual Flow

Overview

Transport data collected with a shipboard ADCP were utilized to estimate SRN residual flow and tidal prism. Four transport time series (two spring tides and two neap tides) were produced for each of two channel cross-sections. For this study, “residual flow” (Q_R) is defined as mean transport (m^3/s) over one full semi-diurnal cycle, and “tidal prism” (V_P) as the semi-diurnal flood transport (m^3) in one such cycle. Section 5.2 details two methods by which residual flow was calculated from the data; one method also produces tidal prism values. Results from the two methods are also presented in Section 5.2. Section 5.3 addresses the effectiveness of a shipboard ADCP field study for such measurements, and presents possible explanations for anomalous results. Conclusions drawn from the exercise are presented in Section 5.4.

5.1 Introduction

The ecological impact of pollutant loading to an estuary is often directly related to the residence time of the pollutant. Information concerning the circulation and fresh water input can assist with

estimates of residence time. Heat pollution is of particular concern in MHB. A local power plant utilizes bay water to cool its machinery, and then exports the water at an elevated temperature back into the bay (Figure 5.1). The residence time of the effluent plume is one of the factors determining its impact on the MHB ecology. Chapters 3 and 4 have discussed the tidal and wind driven circulation of MHB and the upper SR. This chapter presents information concerning net transport, or residual flow. Residual flow in an estuary is often directly related to the fresh water input, but with MHB the fresh water input should relate to the net of the two outlets (EP and SR). Since this study can only present results for the SR, the fresh water input to MHB will not be addressed. Furthermore, in order to discount the possibility that the residual transport is induced by local wind fluctuations, a much longer data set is required (Weisberg 1976a). Rather, the residual flow for the SRN simply lends further insight into the low frequency exchange between MHB and the SR.

Because it was the existing technology, current velocity data from current meters have traditionally been used to calculate residual flow. One or more point measurements must be extrapolated across a channel section to obtain transport. For this reason, cross channel flow inhomogeneity introduces error in such calculations; recall that flow in this region has been shown to be consistently non-uniform across a channel section (Section 3.2.2), both horizontally and vertically. With

Figure 5.1 The SRN influence on MHB heat pollution removal.

The Brayton Point Power Plant, located in the NE corner of MHB utilizes bay water to cool its machinery, and exports the heated water back into the bay. Combined with fresh water from the Taunton River, this heat travels down bay as a density plume. This figure hypothetically depicts the path of the density plume as it travels from the TR south within the dredge channel. Results from this study indicate that part of the plume reaches the SRN and gets mixed down into the water column, diminishing its release rate to the atmosphere. In addition, the mean flow is at times up bay, from the SRN into MHB, further increasing the residence time of the heat pollution.

presently existing technology a moored ADCP may be used to sample the entire water column, but this method still requires the water column to be representative of the channel section.

In this study, a shipboard ADCP is used to sample current velocity across channel sections. Obvious advantages include sampling a large fraction of the channel cross sectional area, for calculating transport from mean velocity. The primary disadvantage of this method over a moored system is the required manpower and the limited duration of the time series: only four individual semi-diurnal cycles (at two cross sections) are available. Another limitation with the shipboard measurements is sampling frequency. In this study, sampling was relatively infrequent, only once every 40 minutes. Because of the short record and low sampling frequency, two calculation methods are used to estimate residual flow from these transport data.

5.2 Methods and Results

Residual flow was calculated from the transport data (Q) by two methods (Table 5.1). With method A, a linear interpolation of the data is created. The residual flow (Q_R), or net transport, may be understood as the integral of the curve. The integral may be calculated geometrically as the difference between net inflow (positive transport) and net outflow

Date	Tidal Prism m^3		Residual Flow A (m^3/s)		Residual Flow B (m^3/s)	
	North	South	North	South	North	South
4/7/97	1.13E+07	1.05E+07	7.4	-100.8	4.3	-101.4
4/8/97	1.39E+07	1.21E+07	71.7	-40.3	65.9	-52.4
4/15/97	4.02E+06	4.34E+06	-85.1	-71.7	-86.9	-73.6
4/16/97	6.80E+06	5.64E+06	-49.3	-94.1	-54.5	-96.2

Table 5.1 Tidal prism and residual flow values for the SRN.

Transport data were collected with a shipboard ADCP. Each tidal prism value represents the volume of water (m^3) transported through each channel cross section (north and south) over one semi-diurnal flood tide. The residual flow values represent mean transport over one semi-diurnal cycle. Spring tide days are the 7th and 8th, and neap tide days are the 15th and 16th, representing end member system conditions. Residual flow was calculated with two independent methods (A and B), both providing similar results. The most notable anomaly relates to the spring tide residual flow at the north station; values are positive rather than the expected negative values found in every other case.

(negative transport) over one semi-diurnal tide cycle with period T (5.1).

Similarly, net inflow provides an estimate for V_P (5.2).

$$Q_R = \int_0^T Q(t) dt * \frac{1}{T} \quad (5.1)$$

$$V_P = \int_{t_1}^{t_2} Q(t) dt \quad (5.2)$$

In equation 5.2, the integral spans one flood tide, or the time of slack low water (t_1) to the time of slack high water (t_2).

The second method (method B) fits an ensemble of sinusoids at known tidal frequencies (f_n) to the transport data (Q). The algorithm chooses the optimal least squares fit for various amplitudes (A_n in units of transport) and phases (Φ_n) of each component (5.3). The nine components chosen for this computation are those identified by the sea surface oscillation tide component analysis (Section 3.2.1). The method also allows for a constant Q_R to be added to the curve. When integrated over a whole number of periods, each individual sinusoid has zero net contribution, and this constant (Q_R) represents the residual flow.

$$Q(t) = Q_R + A_1 \sin(f_1 t + \Phi_1) + A_2 \sin(f_2 t + \Phi_2) + \dots A_9 \sin(f_9 t + \Phi_9) \quad (5.3)$$

Results from the analysis are presented in Table 5.1.

Tidal prism magnitudes for the spring tide days (1.13 - 1.39×10^7 m^3) are about 2-3 times the values for the neap tide days (4.02 - 6.8×10^6 m^3), and seem to be relatively consistent between the north and south

stations. For the residual flow, methods A and B provide similar results, indicating residual flows as large as 10% of the peak observed transport ($\sim 1000 \text{ m}^3/\text{s}$). But, values vary between stations, and between spring/neap conditions. Most notably, the residual flow over half a day at the north station under spring tide conditions was positive, or northward, while all other values were negative, or southward; residual flow is often expected to be consistently down bay.

5.3 Discussion

As mentioned previously, heat contamination in MHB has been a topic of concern. Results from this study may lend insight into the fate of heat pollution in MHB, and facilitate estimates of the pollutant residence time. In this section, the inconsistency of the residual flow data will be addressed (Section 5.3.1), and the implications of such flow on MHB flushing will be discussed (Section 5.3.2).

5.3.1 Residual flow variability

The observed positive or northward flow seems anomalous. This result was first observed using the interpolation method (A). Because of concerns that the sampling frequency (1/40 minutes) might not be

adequate to approximate the waveform, the second method (B) was also developed. Since method B utilizes an entirely different approach yet produces similar values, the residual flow results are believed to be valid.

Residual flow in an estuary is often related to fresh water input, which should produce net outflow, or in our coordinate system, negative flow. In this study, residual flow was sometimes observed to be up bay. In addition, it would be expected that net flow through the northern cross-section would closely resemble net flow through the southern cross-section. Residual flow between stations was also observed to be inconsistent. In addition, the data has a high magnitude and broad range (about +70 to -100 m³/s); residual flow was often up to 10% (100 m³/s) of peak observed transport (1000 m³/s). Spaulding and White (1990) calculated a 150 m³/s residual flow southward through the SRN.

But the residual flow through the SRN is not constrained to be seaward due to fresh water input. Given the existence of two inlets to MHB, it is feasible that one inlet could have net inflow and the other net outflow. A more appropriate term might be *mean flow*, where the general concept does not suggest fresh water input. Once this expectation is removed, the data seem much less anomalous.

The results suggest the tides as the most likely mechanism for the anomalous residual flow. Excluding fresh water input as an explanation, winds and low frequency tides could affect the results. Winds during the study period were light and can most likely be excluded also. Tidal

explanations could result either from the influence of tidal components at sub semi-diurnal frequencies, or from tidal exchange between the SR and the EP. One semidiurnal tidal cycle was skipped between transport measurements made on consecutive days. As a result the same half of the diurnal cycle was sampled on these two days in each case (spring and neap), which could account for the anomaly. In principle, the same similarity could be attributed to common tidal amplitude; high amplitude tides (4/7 and 4/8) exhibit mean flow out of the SRN from both sides, while low amplitude tides (4/15 and 4/16) exhibit mean flow seaward at both stations, from MHB into the lower SR. But this would result in complete draining of the SRN within a day or two during spring tides, and this does not occur. Therefore, the results suggest the diurnal tides as the most likely mechanism for significant tidal contributions to the residual flow values. But, an explanation for the opposing directions of flow observed at the two stations is not clearly evident.

5.3.2 MHB flushing

The way in which water is exchanged between MHB and its two inlets, the EP and SR, has direct implications on the flushing of MHB pollutants. As mentioned previously, one pollution concern for MHB is heat. The Brayton Point Power Plant is located in the upper reaches of

MHB, close to the Taunton river inlet. Effluent emissions are conducted in bursts coincident with high water. Fresh water from the river combined with the heated plant effluent results in a low density plume (Figure 5.1). The fate of this plume will determine the impact of the heat on the ecology. Recall that water passing from MHB into the SRN was found to develop a warm fresh surface layer (Section 4.1.1). Therefore it is believed that at least part of the plume reaches the SRN.

Based on results from this study, the SRN system serves to prolong the residence time of heat pollution within MHB. Heat is removed from the system most easily when confined to the surface layer, as it may freely escape to the atmosphere. If the heat were distributed throughout the water column, it would not escape as easily, and therefore the residence time of the pollutant would increase. Recall that turbulence induced by the SRN breakwaters causes the water to become well mixed vertically. The heat once carried at the surface becomes distributed throughout the water column, lowering the surface temperature and raising it slightly at depth. With each ebb tide, thermal energy gets mixed from the surface down into the water column. Furthermore, mean flow is at times up-bay, causing some of the well-mixed heated water to return to MHB, further increasing its residence time, and raising mean temperatures at intermediate and deep depth levels. Even without mean flow up the Bay, some of this mixed water is

returned to MHB on each flood tide and horizontally advected and mixed there.

5.4 Conclusions

The tidal prism for the SRN ranges from $4.02\text{E}+06$ to $1.39\text{E}+07$ m^3 for spring and neap conditions, respectively. Residual flow values for the south station were consistently negative, or seaward, and range from 40-101 m^3/s . Residual flow values for the north station were negative (49-87 m^3/s seaward) for the neap tide case, but positive (4-72 m^3/s up bay) for the spring tide case.

SRN residual flow values, or more appropriately mean flow values, seem to vary with the tides. The influence of the diurnal tide is the most likely explanation for the inconsistent results obtained. A study period of a day or longer would better constrain residual flow estimates. To obtain truly sub-tidal residual flow values, a sampling duration of days to weeks would be required. But such a long duration would be unreasonable for an underway ADCP survey, and a moored system would have difficulty with the lateral flow variability observed, unless an array of ADCPs was used. Therefore it is recommended that future studies utilize a day-long shipboard ADCP survey for SRN residual flow estimates. Note that the 40-minute transport sampling frequency was found to be adequate to

approximate the transport between samples. Alternatively, underway ADCP data could be used to calibrate a moored system.

Residual flow estimates are of primary interest to the flushing of MHB due to concerns of power plant effluent heat contamination. These data combined with other study observations suggest that the SRN has a detrimental effect on MHB heat pollution flushing. Tidal circulation and constriction-induced water column mixing cause the residence time of the heat in MHB to increase. In conclusion, the SRN constrictions serve to prolong the residence time of heat contamination in MHB, and the increased residence time increases the probability of detrimental impacts to MHB ecology.

6 Summary and Conclusions

In this Chapter, results from the study are summarized and conclusions are drawn. Section 6.1 summarizes the results and presents a comparison with data from previous studies. Section 6.2 presents the main conclusions, since previous chapters have discussed the results in detail.

6.1 Study Results Summary

- SRN maximum observed spring tide current velocity values reach 1.5 m/s, corresponding to a peak volume transport of about 1000 m³/s.
- The tidal prism for the SRN ranges from 4.02×10^6 to 1.39×10^7 m³ for neap and spring conditions, respectively.
- The tidal current between the SR and MHB is predominantly semidiurnal, but it exhibits a double-peaked flood and a single-peaked ebb indicative of a significant M4 component.
- Peak ebb current occurs shortly after high water.

- The tidal current leads from north to south through the system; flood and ebb currents are observed earlier between MHB and the SRN than between the SR and the SRN.
- The cross channel structure of the velocity field was observed to be consistently and dramatically inhomogeneous.
- Flow in the vicinity of the breakwaters was turbulent, and regularly mixed the water passing through the SRN.
- The tidal flushing of MHB is dominated by standing wave motion through the EP inlet. Furthermore, exchange between MHB and the SR passage is also controlled by the oscillation of the MHB free surface, rather than the oscillation of the lower SR free surface.
- A weakly non-linear tidal transport model was developed. Model predicted tidal volume transport agrees with observed values within approximately 25%.
- Winds out of the SE (NW) had the most influence on the sea level rise (fall) rate, with winds from the SE (NW) causing flow into (out of) the SRN from (to) both the south and the north. In addition, this off channel axis

wind direction likely reflects the cumulative wind effects over NB and RIS.

- Storm events are believed to explain large anomalies in the predicted low frequency transport record. Anomalies at times exceed tidal influence.

Note: See Table 6.1 for a comparison of data from this study with previous observations.

6.2 Conclusions

Study results clearly illustrate many curious aspects of MHB/SR water volume exchange. The breakwaters located just north and south of the SRN constrict the channel to 25% of its nearby width. These constrictions restrict volume transport between MHB and the SR, producing effects with time scales from minutes to days and beyond. Both tide and wind induced volume transport and sea surface height observations exhibit small-scale spatial variability within the SRN region. In order to interpret the response of the system to the various forcing mechanisms, close inspection of this variability was required.

	DeLeo (2000)			Hicks (1959)	Spaulding and White (1990)	Kincaid (1996)
	N	M	S			
Peak Observed Tidal Current Velocity (cm/s)	100	-	150	135	25	100
Peak Observed Tidal Transport (m ³ /s)	1300	-	1000	-	-	1200
Residual Flow (m ³ /s)	-15.8	-	-78.8	-1.8	-150	-
M2 Amplitude (cm)	57.8	49.4	49.5	-	57.6	-
M4 Amplitude (cm)	10.1	7.5	6.2	-	10.1	-

Table 6.1 Sakonnet River Narrows summary. Residual flow values for this study are the average value of the four tidal cycles for comparison. The DeLeo North Station (N) is co-located with both Spaulding’s and Kincaid’s SR velocity measurements. Hicks’ residual flow was derived from reported non-tidal transport for the mouth of the SR, Spaulding’s residual flow was estimated from mean point velocity measurements, and Kincaid’s were derived from ADCP transects. Spaulding M2 tide was measured at the MH Bridge, and M4 tide was measured in the lower Taunton River.

Given these results, it is concluded that the SRN breakwaters are responsible for a multitude of anomalous observations collected to date, in this work and in previous works. Therefore, consideration for the influence of these constructions is essential for a comprehensive understanding of MHB/SR circulation and exchange. Future investigators conducting research in the region are strongly urged to consider the influence of the constrictions.

Appendix

Tidal analysis methods

The tidal constituents listed in Table 3.1 were obtained through a two-step process. The “response” method was first used to obtain the amplitudes and phases of the primary constituents. The residual signal from the response method was then analyzed for components at overtide frequencies with the “harmonic” method. The techniques, benefits and limitations of each of these methods will be presented in this section.

With the response method, ocean topography is treated as a time invariant linear filter for the astronomical tidal forcing function. For any time invariant linear filter, the properties of the filter may be specified by the impulse response of the filter ($l(t)$) to a delta function ($\delta(t)$) or “impulse”. Given this “impulse response”, the response of any given input function may be calculated by convolution of the impulse response with the input function,

$$x(t) \rightarrow \eta(t) = \int_{-\infty}^{\infty} l(\tau)x(t - \tau)d\tau .$$

In this example, the input signal ($x(t)$) is the oscillating forcing potential from the moon's and sun's gravitational fields, and the output signal ($\eta(t)$) is the tidal component of the observed sea surface oscillation.

The accuracy of this method may be increased by replacing the localized forcing potential with a spherical harmonic representation of the global equilibrium tide (Munk and Cartwright 1966). This refinement allows for the inclusion of large-scale effects on the local signal. Coefficients for the spherical harmonic components are represented as weighting functions, which are determined through a least-squares fit to the observed signal.

With the harmonic method, one prescribes particular frequency components of the astronomical forcing potential, and this function is solved with a least-squares fit to the observed signal,

$$\eta(t) = \sum_{i=1}^N (A_{ci} \cos a_i t + A_{si} \sin a_i t).$$

Variables A_{ci} and A_{si} are solved for, giving the amplitude $(A_{ci}^2 + A_{si}^2)^{1/2}$ and phase $\arctan(A_{si} / A_{ci})$ of each component, and $f_i = \frac{a_i}{2\pi}$ are the known frequency components.

The response method was utilized for the primary constituents because it tends to be more accurate with short records (Munk and Cartwright 1966); it requires a shorter sampling period than the

harmonic method to resolve the same frequencies. The response method cannot, however, determine overtides since they are produced by nonlinear dynamics. For example, the M4 signal is generated from nonlinear response to forcing at the M2 frequency. Therefore M4 (the strongest overtide) was analyzed with the harmonic method.

To obtain a record of the non-tidal sea surface height variability, a generated curve at each frequency, calculated amplitude, and calculated phase was subtracted from the observed signal. The resulting signal still contained a small tidal component due to additional overtides. The record also contained contributions from other environmental factors, such as wind.

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