PATTERNS AND RATES OF HISTORICAL SHORELINE CHANGE IN THE DELAWARE ESTUARY

by

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A thesis submitted to the Faculty of the University of Delaware in partial fulfillment of the requirements for the degree of Master of Science in Marine Studies

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ABSTRACT

Shoreline change in coastal and estuarine systems is a result of both natural and anthropogenic factors that influence sediment accumulation and erosion within the intertidal zone. The Delaware River and Bay estuary, a 215-km long coastal plain estuary on the U.S. Atlantic coast, is a submerging estuarine basin consisting of a tidal freshwater river (upper estuary), a stratified estuary (lower estuary), and a weakly stratified bay (Delaware Bay) at its mouth. Beginning in the late Holocene and continuing to present, marine transgression has caused the bay and estuary to broaden, its shores to retreat landward and upward, and its coastal environments to transition from tidal wetlands and tidal flats to sandy, barrier beaches. Superimposed on this natural change are anthropogenic influences on estuarine sedimentation, beginning in the 19th Century, such as construction of a shipping channel, maintenance dredging, shoreline hardening, and modification of tidal wetlands. Although previous research has shown that much of the estuary shoreline is retreating because of transgressive erosion, the nature of shoreline change in the estuary–bay as a whole has never been established.

In this study, patterns and rates of shoreline change in the estuary from 1879 to 2012 were characterized using five shoreline datasets (1879, 1948, 1991, 2007, 2012) and the USGS Digital Shoreline Analysis System (DSAS) extension for ArcGIS. Linear rates of shoreline change were computed using both linear regression and endpoint methods to investigate temporal variations in shoreline extension and retreat. Volumetric rates of shoreline change where determined using DSAS and sediment bulk density data, to estimate the mass of sediment associated with shore erosion and accretion. Given that wind waves are a known agent of coastal change in the estuary, archived wave data (2007–2015) were examined to identify potential relationships between wave parameters (significant wave height, wave period, wave power) and rates of shoreline retreat. Results indicate that coasts of the lower estuary–bay have been in a state of net retreat during historical times. From 1879 to 2012 the long-term rate of shoreline change for the entire lower estuary–bay was -1.1 ± 0.13 m/yr. This rate of retreat equates to -1.5 ± 0.18 x10^{8} kg/yr, assuming retreat is due to erosion of the shoreface. By comparison, the short-term (2007–2012) rate of shoreline retreat for the lower estuary–bay system was higher at -2.13 ± 0.47 m/yr. Long-term rates of shoreline change for the lower estuary region alone were -0.64 ± 0.13 m/yr and -1.3 ± 0.13 m/yr on the Delaware and New Jersey sides, respectively. In the bay region long-term rates on the Delaware and New Jersey sides were respectively -0.73 ± 0.13 m/yr and -1.7± 0.13 m/yr. Among the four different types of coasts classified for this study (barrier beach, tidal wetland, transitional wetland–barrier, and hardened), transitional and wetland coasts had higher rates of shoreline retreat than the barrier beaches. In sum, both long-term and short-term rates of shoreline retreat are higher on the New Jersey side of the lower estuary–bay.

Comparison of modeled wave properties and shoreline change data indicates a general correlation between wave power and shoreline retreat, presumably due to wave erosion of the shoreface. However, further research is needed to identify the actual mechanisms and time-dependence of shoreface erosion. By documenting historical shoreline change in the estuary, the findings of this study can help identify vulnerabilities associated with sea-level rise, climate variability, and human pressures.

Chapter 1

INTRODUCTION

Shoreline change in estuarine and coastal systems is a consequence of natural and anthropogenic processes that take place over a wide range of temporal and spatial scales. The position of the shoreline, defined by the mean high water line, falls at the top of the shoreface, the intertidal-to-shallow subtidal zone influenced by wave-produced currents. Despite decades of research, our understanding of shoreline change in the context of shoreface dynamics is limited. This is particularly true for large estuarine systems forced by complex interactions among freshwater outflows, tides, and wind-generated waves, and whose coasts range in type from vegetated tidal wetlands to sandy beaches. Because all estuarine systems are unique in terms of hydrodynamic processes, geomorphology, surficial geology, and human uses, simple models of shoreline change such as the Brunn Rule (reviewed by Rosati et al., 2013) frequently fail to predict observed rates of shoreline change.

The Delaware River and Bay estuary (Delaware Estuary), a submerging coastal embayment within the Atlantic Coastal Plain, has been subjected to significant changes in morphology on both geological and recent timescales. Delaware Estuary is 215-km long coastal plain estuary consisting of a tidal freshwater river (upper estuary), a stratified estuary (lower estuary), and a weakly stratified bay (Delaware Bay) at its mouth (Figure 1). Since late Holocene times and continuing to present, marine transgression forced Delaware Bay to broaden and its coastal environments to evolve from tidal wetlands to barrier beaches (Weil, 1977; Knebel et al., 1988;

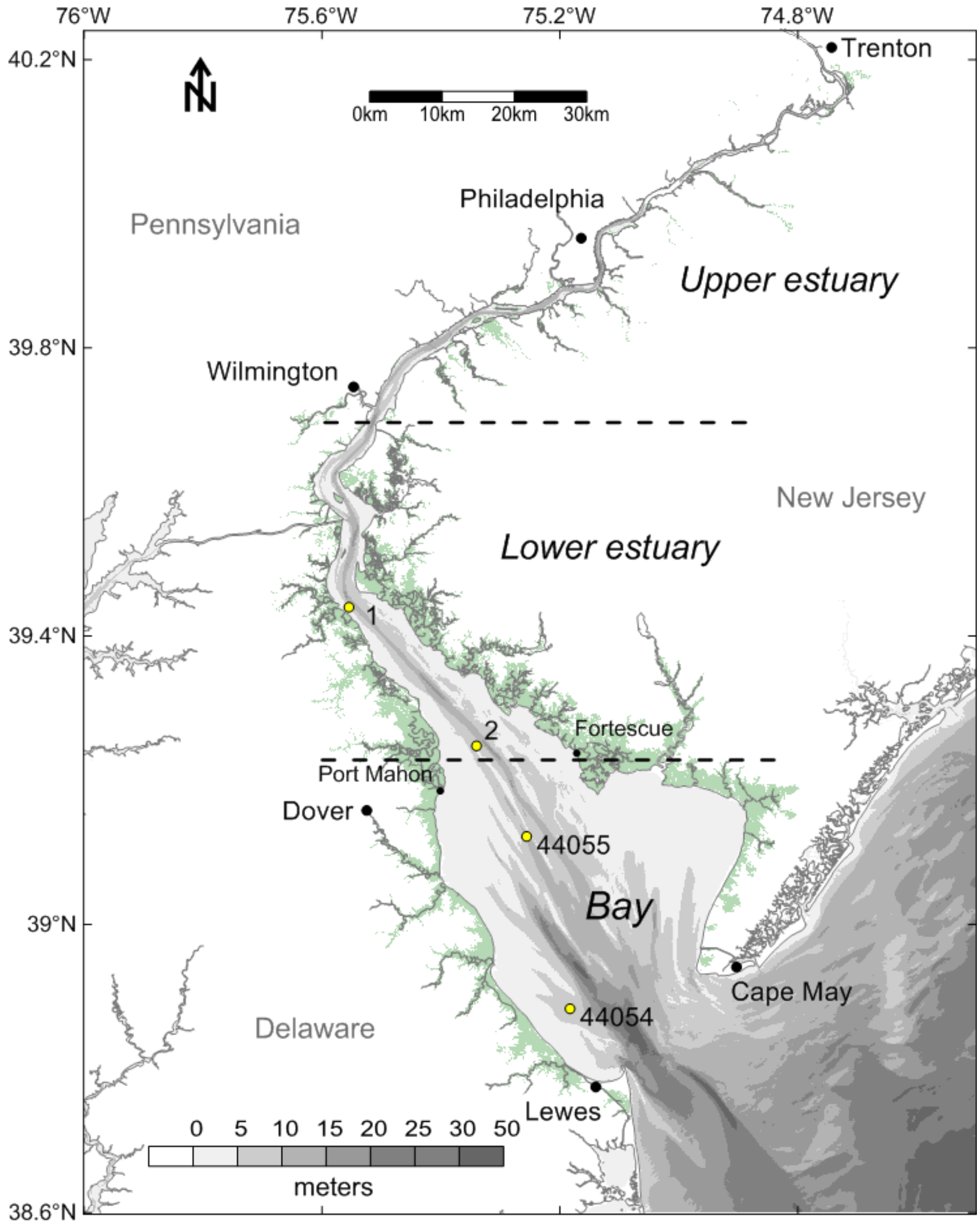


Figure 1. Map of the Delaware River and Bay estuary showing geographic features referred to in the text.

Fletcher et al., 1990; Fletcher et al., 1992; Kraft et al., 1992). Transgressive erosion of the bay shoreface by waves superimposed on continual sea-level rise caused the bay coast to migrate landward and upward over time. Associated with bay widening was an increase in fetch and a transition from tide- to wave-dominated sediment transport, which led to a change from a mud- to sand-dominated coast (Weil, 1977; Fletcher et al., 1990). Superimposed on this natural change are anthropogenic influences on estuarine sedimentation and coastal change, starting in the 19th Century, including shipping channel construction, maintenance dredging, shoreline hardening, and widespread modification of tidal wetlands.

Shoreline change analysis based on coastal mapping information is a useful tool for quantifying shoreline migration during historical times. Previous studies of mapped shoreline change in Delaware Bay have established that much of the bay coast is retreating under the influence of transgressive erosion (Maurmeyer, 1978; Phillips, 1986; French, 1990; Kraft et al., 1992). However, because these studies focused on specific segments of the coast, and considering that mapped rates of change are highly variable, the results are not applicable to the greater estuary–bay system. Moreover, there has never been a study of historical shoreline change in the lower estuary, where rates of marsh edge erosion ranging from 2 to 7 m/yr have been reported (Phillips, 1986; Kraft et al., 1992).

The goal of this thesis research was to quantify patterns and rates of historical shoreline change in Delaware Estuary as a whole, in order to address the knowledge gap described above. To meet this goal, synoptic shoreline data available from government agencies were analyzed using the Digital Shoreline Analysis System (DSAS) developed by the U.S. Geological Survey (Thieler and Danforth, 1994;

Thieler et al., 2009). The DSAS has been applied to studies of shoreline change worldwide, and is considered best practices in shoreline change analysis. Additionally, to gain insight on relationships between wave energy in the estuary and mapped rates of shoreline change, wave data available from four wave buoys deployed between 2007 and 2015 were analyzed. After the background sections provided below, specific objectives of this research are described in Section 1.3.

1.1 Topical Background

1.1.1 Natural Processes and Conceptual Models

The vast majority of research on shoreline change has focused on sandy beaches of fetch-unlimited oceanic coasts, where shoreface erosion and accretion under the influence of breaking waves and sand transport is the main agent of change. By comparison, mechanisms of shoreline change in estuaries have received considerably less attention among researchers (Jackson et al., 2002). In the U.S. Mid-Atlantic region, short-term coastal change is associated with extratropical storms known as "Nor'easters" (Morton and Sallenger, 2003), which by producing higher than average wind speeds for extended periods of time produce large waves and storm surge (Dolan and Davis, 1994). During storm events, waves erode the shoreface by transporting sand to offshore bars, where it is stored until being transported back to beach under fair-weather waves (Sallenger et al., 1985; Hoelfel and Elgar, 2003). Provided with an interrupted supply of sand, oceanic beaches are capable of recovering to pre-storm conditions as the beach heals by sand deposition and accretion (Morton et al., 1994; Sallenger, 2000). Much of the retreat of the Mid-Atlantic oceanic coast during historical times is associated with coastal storms (Figure 2).

Although the rates of change are considerably lower, fetch-limited estuarine beaches display similar patterns of erosion and accretion attributable to storm processes (Jackson et al., 2002).

Shoreline change on tidal wetland coasts has received less than sandy shoreline, although in recent years there has been a surge in publications on the related topic of salt marsh morphodynamics (reviewed by Fagherazzi et al., 2012). Tidal marshes are located in the mid-high latitudes and usually on sheltered coasts, which dampens wave energy (Davidson-Arnott, 2010). The formation of tidal marshes is dependent on variables such as the rate of relative sea-level rise, local sediment supply, vegetation, and the amount of energy in the system. Development of salt marshes begins on vegetated mudflats, generally between mean tide level and mean high water, and continues with development of soil volume created by belowground root biomass produced in-situ along with fine-grained sediment delivered by the tides. Plant growth and sediment trapping by the marsh canopy cause the marsh surface to accrete vertically at a rate approximating the rate of local relative sea-level rise. With continued sea-level rise, salt marshes accrete vertically and migrate landward and upward with coastal transgression. Natural processes associated with tides, waves, vegetation, and morphology of the surface underlying the marsh influence the elevation of the marsh surface relative to mean sea level, as well as the landward and seaward boundaries of the marsh.

The presence of wetland plants has been suggested by some authors to protect the marsh edge from erosion (Francalanci et al., 2013; Möller et al., 2014), but others have reported that vegetation has no influence on marsh-edge stability (Feagin et al.,

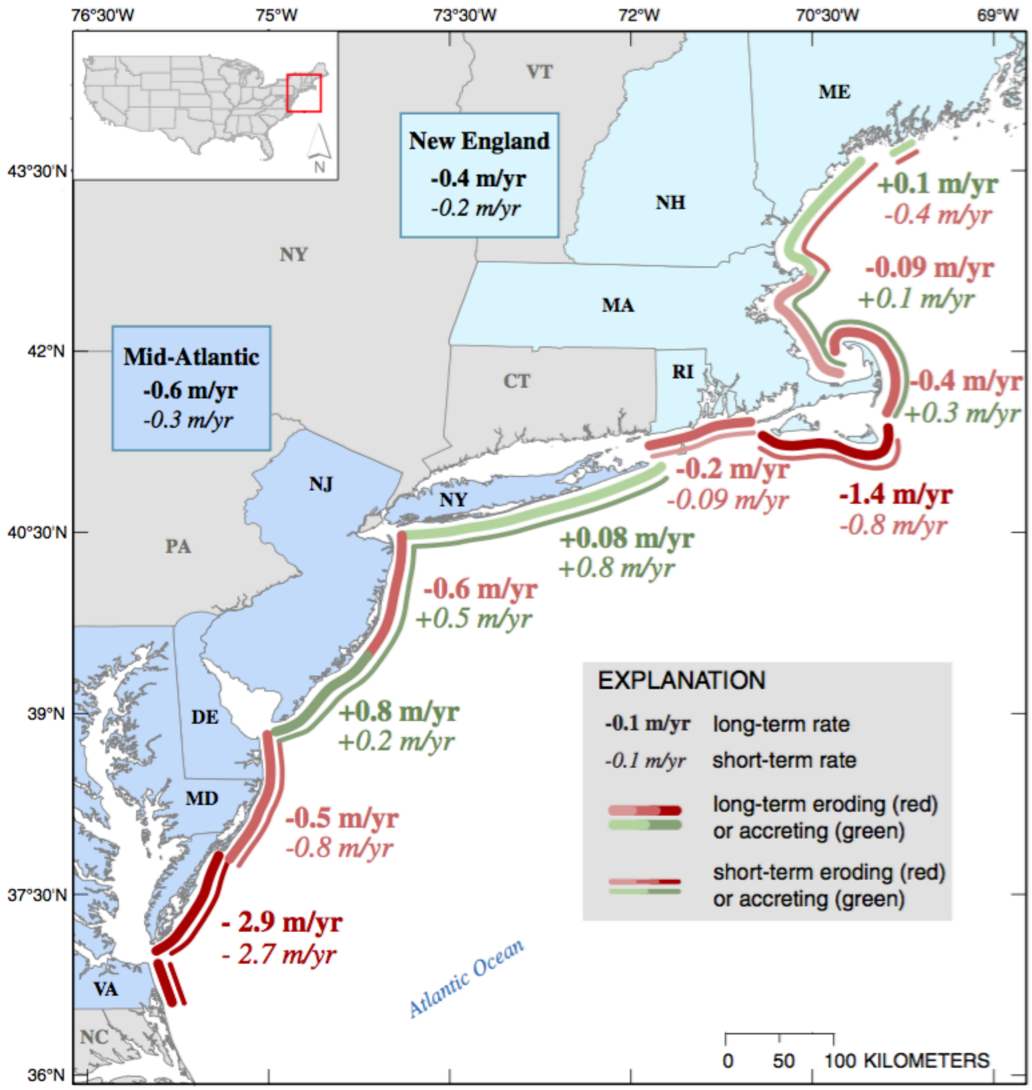


Figure 2. Figure from Hapke et al. (2013) showing patterns and rates of shoreline change on the Atlantic coast. Note that estuarine systems such as the Delaware Bay and Chesapeake Bay were not included in this regional assessment.

2009). Although there is some dispute in the literature as to whether marsh vegetation contributes to the stability of the seaward boundary of the marsh, it is well established that the salt marsh canopy contributes to the entrapment of fine-grained sediment delivered by the tides (Leonard and Luther, 1995; Fagherazzi et al., 2012).

Wind-wave erosion and tidal deposition of sediment are an overarching control on the seaward boundary of tidal marshes and adjacent flats. Whereas tides deliver sediment to tidal flats, waves have a tendency to resuspend sediment and disperse it to other locations. In addition to wave-orbital resuspension of bed sediment, the impact of shoaling or breaking waves on the marsh edge causes undercutting, cliffing, and eventually failure of the marsh edge in large blocks of muddy peat (e.g, Allen, 1989; Schwimmer, 2001). This form of mass erosion, which is distinct from the grain-by-grain erosion of sandy beaches, leads to implications for recovery of the marsh shore after disturbance. A number of researchers have observed a direct relationship between wave power and measured rates of marsh edge erosion and shoreline retreat (Schwimmer, 2001; Roland and Douglass, 2005; Marani et al., 2011; McLoughlin et al., 2015). Wave power (P in kW/m) is described by:

$$P = \frac{\rho g^2}{64\pi} H^2 T \quad (1)$$

As given by Equation 1, the significant height (H) and period (T) of the wave determine its power when it makes contact with the shoreline. Wave height and period in turn are influenced by factors including wind speed and duration, fetch, and local water depth.

When waves make contact with the shoreline, the substrate plays a large role in its erodibility. Shoreline composition, including rock type, sedimentology, and stratigraphy, influence the amount of erosion that takes place per unit wave power applied (Rosen, 1980; Cowart et al., 2010; Cowart et al., 2011; Currin et al., 2015). Unlike sandy beaches, when waves impact the edge of a marsh, there is not always an immediate response in the position of the mean high tide shoreline. Through continuous wave action and undercutting, marsh blocks off into the water and lead to a scarped marsh edge, which does not always recover to its initial state.

Conceptual models of marsh shoreface dynamics described in the literature are useful for understanding shoreline change on tidal wetland coasts (Allen, 1989; Schwimmer and Pizzuto, 2000; van de Koppel et al., 2005; van der Wal et al., 2008; Chauhan, 2009). Although these models were developed from observations in different types of estuaries, they are similar in that they describe erosion and accretion cycles driven by interactions among tides and waves, substrate erodibility, sediment supply, and plant growth. Erosion-accretion cycles start when the marsh surface reaches an elevation sufficient to limit landward propagation of fair-weather waves. This causes waves to impact the shore abruptly at or below the marsh rootmat rather than dissipate continuously over the tidal flat and marsh surface. Because the root mat has a higher yield strength than that of the underlying strata, wave impact and scour lead to scarp formation and undercutting of the marsh edge. When severe, the scarp fails and blocks of marsh fall into the water. The marsh shoreface will recover from such mass erosion only if (1) sediment accumulation in the adjacent intertidal zone is sufficient to raise the bed elevation to mean tide level, and (2) new plant growth by rapidly colonizes and stabilizes the tidal flat. Otherwise, the marsh edge will continue to retreat by mass erosion by wave processes superimposed on relative sea-level rise.

To describe retreating marsh shorelines in coastal Louisiana, Wilson and Allison (2008) proposed an "equilibrium model" of shoreface erosion under relative sea-level rise driven mostly by local subsidence (Figure 3). As a segment of coast subsides within the tide frame, shoaling wind waves scour the shallow intertidal zone, creating an erosional surface extending from wave base landward to the marsh edge. With continued subsidence, the erosional surface deepens and is covered by sediment resuspended by wave scour at the shoreface. Importantly, some of the scoured sediment is transported landward to the marsh and contributes to vertical accretion. In this model, which is applicable to tidal wetland coasts of Delaware Estuary, vertical accretion of the marsh can proceed even with horizontal erosion of its seaward boundary as long as tidal flooding persists and the sediment supply is uninterrupted.

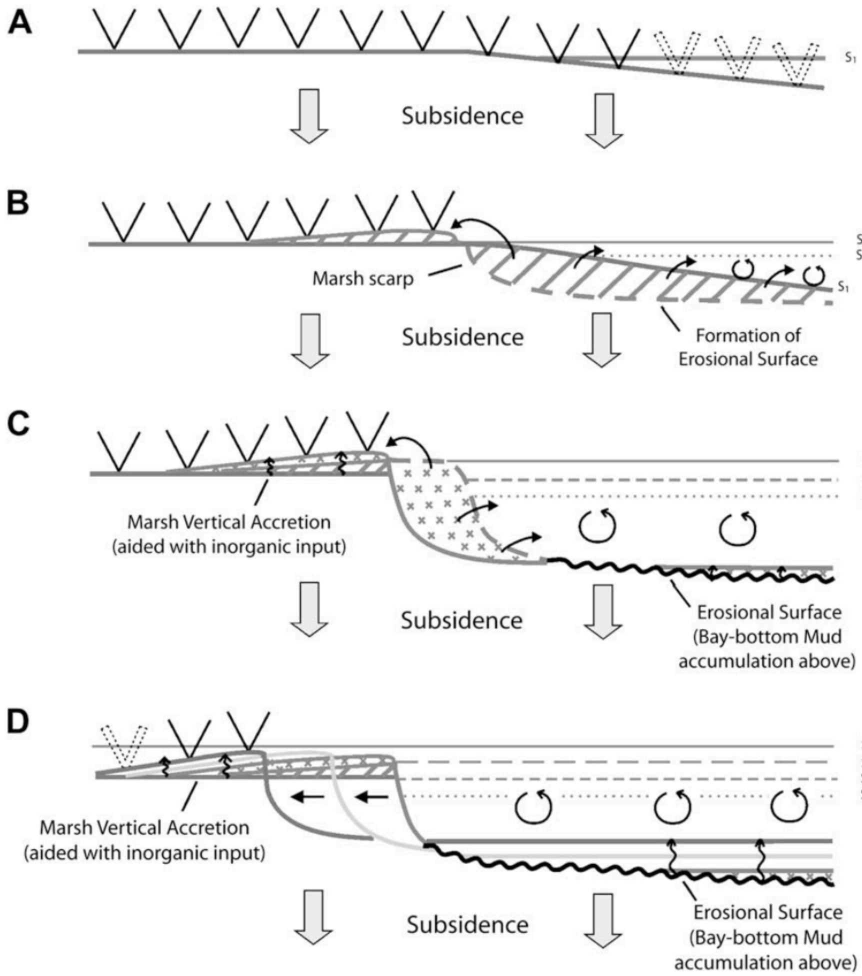


Figure 3. Conceptual model of estuarine shoreface dynamics under the influence of relative sea level rise. Note that the marsh is supplied sediment produced by shoreface erosion. Figure from Wilson and Allison (2008).

1.1.2 Shoreline Change Analysis

Repeat shoreline mapping of coasts is a useful tool for quantifying shoreline change during historical times. In an attempt to understand the mechanisms of shoreline change, it is necessary to first to identify the rates at which the shoreline is moving and how these rates vary in space and time. As reviewed by several authors, there has been a significant improvement in the accuracy of shoreline mapping techniques and change analysis (Crowell et al., 1991; Thieler and Danforth, 1994; Moore, 2000; Boak and Turner, 2005). Early studies of shoreline change relied on National Ocean Service (NOS) T-sheets (topographic sheets), navigation charts, beach surveys, and aerial photography. Aside from positional inaccuracy and differences in projections, sources of error in early shoreline change studies included failure to account for tides at the time of mapping, differences in the shoreline definition (i.e., high water versus low water, overlapping photos, and, in the case of aerial photography, failure to correct for pitch, roll, and differences in altitude (Boak and Turner, 2005). Additionally, calculations of shoreline change in early studies were performed by hand, making it tedious to compute shoreline change data at the highest level of resolution.

Advances in satellite technology and remote sensing techniques have significantly improved the accuracy of topographic maps and navigational charts produced by survey agencies. Positional accuracy of the shoreline has improved with the development of LiDAR (Light Detection and Ranging) and high-resolution orthophotography (Leatherman, 2003; Boak and Turner, 2005). Significantly, the Digital Shoreline Analysis System (DSAS) developed by the U.S. Geological Survey has provided a standardized, automated methodology for analyzing shoreline change (Thieler and Danforth, 1994; Thieler et al., 2009). In this system, rates of shoreline change are based on endpoint and (or) regression analysis of shoreline positions along equally spaced transects at a resolution selected by the user.

Although there have been improvements, there remain limitations in shoreline change analysis when it comes to tidal wetland coasts. For example, on sandy coasts the shoreline is typically identified as the high water line (HWL), a visible wet/dry boundary created by the difference in coloration of the sand (Crowell et al., 1991; Boak and Turner, 2005). However, on wetland coasts the distinction becomes more difficult to make due to the dense vegetation, dark coloration of the sediment, and marsh scarps.

1.2 Regional Background: The Delaware Estuary

1.2.1 Physical Setting

The area of Delaware Estuary investigated in this study included both coasts of the lower estuary and bay (Figure 1). The upper estuary is primarily tidal freshwater between Wilmington and Trenton, the lower estuary ranges from oligohaline (0–5 ppt) to mesohaline (5–18 ppt), and the bay is polyhaline (18–30 ppt). Tidal wetland coasts of the lower estuary and bay are characterized by channel-flat-marsh complexes that fringe numerous subestuaries. The vegetated brackish and salt marshes are dominated by *Spartina alterniflora*, *Spartina patens*, and *Phragmites australis*.

The Delaware Estuary formed in the drowned valley of the ancestral Delaware River, and has undergone two full transgressions and regressions during late Pleistocene times (Weil, 1977, Knebel et al., 1988). Coasts of the Holocene–modern estuary have been migrating landward since approximately 9 ka (Fletcher et al., 1990; Fletcher et al., 1992). A conceptual model for the Holocene evolutionary history of the estuary proposed by Fletcher et al. (1990) is shown in Figure 4. According to this model, the bay has been in a destructive phase since about 5 ka with retreating coasts characterized by sandy, barrier beach deposits atop older tidal wetland mud and peat strata. These older strata were deposited during middle to late Holocene times when the wave-dominated bay of present was a relatively narrow, tide-dominated estuary. Barrier beaches of the modern bay give way to tidal wetland coasts about 60 km up-estuary of the mouth in the vicinity of Port Mahon and Fortescue on the Delaware and New Jersey sides, respectively. As transgression continues under the influence of relative sea-level rise, and as the width (and fetch) of the lower estuary increases, sand derived from erosion of Pleistocene strata underlying the Holocene sediments will begin to accumulate along the tidal wetland coast of the lower estuary. As detailed later, results of the present study indicate that virtually the entire lower estuary shoreline has been retreating since the late 1800s, challenging the notion that this region is "constructive" as proposed by Fletcher et al., (1990).

Rates of relative sea-level rise in the Delaware Estuary based on tide gauge records range from 3.5 mm/yr (1956–2015) at Reedy Point, DE, to 3.4 mm/yr (1919–2015) at Lewes, DE, to 4.5 mm/yr (1965–2015) at Cape May, NJ (NOAA, 2015). These rates approximate long-term rates of marsh vertical accretion on both sides of the estuary (Kraft et al., 1992; Nikitina et al., 2000), suggesting that the marshes are accreting at pace with rising sea level. Averaged over the past two millennia, the rate of relative sea-level rise determined by ^{14}C dating of marsh peat deposits is ~1.3 mm/yr (Nikitina et al., 2015).

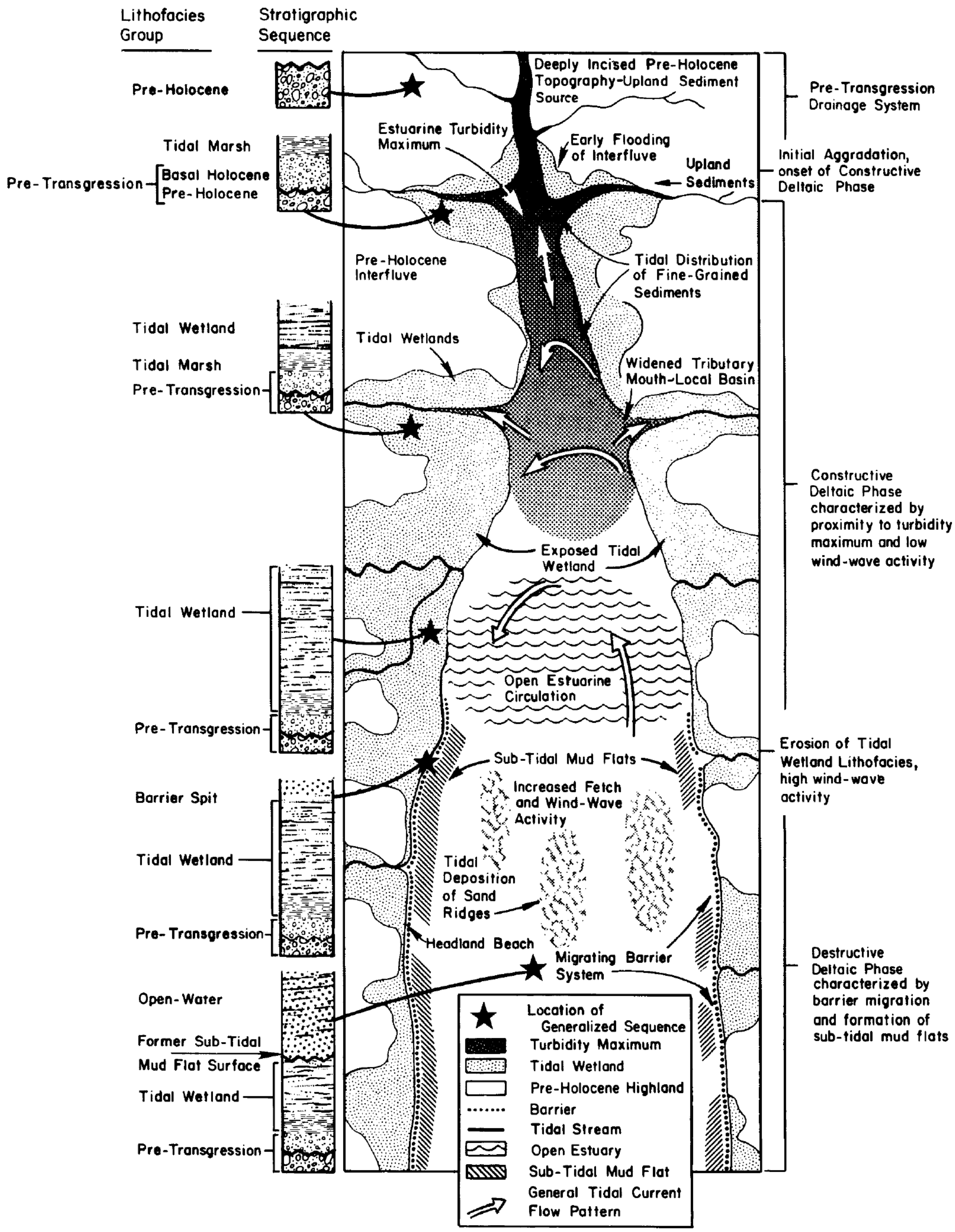


Figure 4. Conceptual model of the Holocene transgression in Delaware Estuary from Fletcher et al. (1992).

Nikitina et al. (2015) attribute most of this rise to land subsidence resulting from glacio-isostatic adjustment, driven by collapse of the proglacial forebulge associated with the former Laurentide Ice Sheet.

1.2.2 Anthropogenic Influences

In additional to the natural factors described above, shoreline change in estuaries is influenced by anthropogenic factors. For example, hardening of the coastline by bulkheads and other engineered structures changes the local sediment transport pathways and is known to limit or prevent natural transgression of the coast. In some estuaries, the Delaware Estuary included, shoreline hardening has greatly reduced the horizontal extent of tidal marshlands (Kraft et al., 1992; Mattheus et al., 2010; Fontolan et al., 2012). In upper Delaware Estuary, most of the natural intertidal zone has been filled and shoreline bulkheaded during historical times to accommodate the needs of the port complex, industry, and municipal infrastructure. Elsewhere in the estuary there has been localized construction of jetties to stabilize inlets, such as at the mouth of the Mispillion River, Delaware, and breakwaters to dissipate wave energy, such as the one near the bay mouth off Lewes, DE.

The U.S. Army Corps of Engineers has been modifying the Delaware shipping channel since 1885 and is currently deepening the shipping channel to 13.7 m (45 ft) to accommodate deep draft ships transiting the estuary to the Wilmington-Philadelphia port complex. Previous channel dredging in the Delaware Estuary involved creation of dredge spoil sites, which have modified the shape of the shoreline at two locations on New Jersey side of lower estuary. In some European estuaries it has been hypothesized that channel deepening changes the tidal prism and hydrodynamics of the estuary with implications to tidal marshlands (e.g., Cox et al., 2003; Liria et al., 2009). Among other factors, estuarine deepening has potential to decrease wind-wave and ship-wake dissipation by bottom friction as waves approach the intertidal zone (Kraft et al., 1992; Cox et al., 2003; Fontolan et al., 2012). Because locally produced wind waves are the most dominant cause of shoreline erosion in estuaries, changing the nature of wave propagation as potential lead to higher rates of shoreface erosion and shoreline retreat.

1.2.3 Previous Work on Shoreline Change in the Delaware Estuary

Phillips (1986) published one of the first studies of shoreline change in the lower estuary (Cumberland County, NJ), determining that the coast eroded at an average rate of -3.21 m/yr averaged from 1940 to 1978. Maurmeyer (1978) and later Kraft et al. (1992) and reported localized rates of retreat as high as -6.9 m/yr along segments of the Delaware coast of the bay. In a study of the Delaware coast between Lewes and Port Mahon, French (1990) computed an overall average rate of retreat of -1.37 m/yr (1842–1977). It is important to point out that rates of shoreline change are highly variable along even a short length of coast, and that rates vary with the period of averaging, generally decreasing with increasing timespan due a "filtering effect" of short-term variability in erosion and accretion (Crowell et al., 1993). For this reason it is important to use synoptic shoreline datasets for the entire region of study, and average shoreline change rates over consistent timespans of interest.

Early shoreline change rates for the Atlantic coast of Delaware have been reported to range from -0.89 m/yr (Galgano, 1989) to as high as -3 m/yr (Dolan et al., 1979). A more recent and comprehensive study of U.S. Mid-Atlantic shoreline change by Hapke et al. (2013) reports long-term (1845–2000) and short-term (1980–2000) retreat rates of -0.5 m/yr and -0.8 m/yr, respectively, for the Atlantic coast of Delaware (see Figure 2). Interestingly, these rates of retreat are significantly lower than those previously reported for Delaware Estuary, which are on the order of meters per year. Understanding the reason for this difference was one of the motivating factors of the present study.

1.3 Research Objectives

To advance our understanding of shoreline change in the Delaware Estuary in the context of shoreface processes, this study aimed to quantify patterns and rates of historical shoreline extension and retreat, focusing on the lower estuary and bay (Figure 1). The specific objectives of this study were as follows:

1. characterize shoreline types throughout the lower estuary and bay using high-resolution orthophotography;
2. compute rates of shoreline change (retreat and extension) based on synoptic shoreline data available for the estuary;
3. estimate volumetric change associated with migrating shorelines; and
4. relate measured and modeled wave properties to the patterns and rates of shoreline retreat identified through shoreline change analysis.

These research objectives were met following methodologies described in Chapter 2.

Chapter 2

METHODS

2.1 Shoreline Data Collection

2.1.1 Data Selection

To document shoreline change in the study area during historical times, data from several sources and covering specific time periods were used. Although high-quality shoreline datasets extending back to the mid-1800s are available for the region, only a few are complete and cover the entire estuary without gaps, and synoptic, (constructed from surveys conducted within a period of about ten years). For this reason not all of the shoreline data available for the estuary from survey agencies were used for this study. Because virtually the entire shoreline of upper Delaware Estuary above Wilmington is stabilized by bulkheads and walls, this area was not included in the shoreline change analysis.

Shoreline data selected for this study is listed in Table 1. For the lower estuary and bay, shorelines representative for 1879−1885 (hereafter 1879) and 1943−1948 (1948) were digitized from NOAA topographic sheets (T-sheets) for analysis in a GIS. The associated positional uncertainty for each T-Sheet ranges between ± 8−10 m depending on the scale of the original T-Sheet (Shalowitz, 1964). The 1948 shoreline was missing a small area of coverage near Port Mahon on the Delaware side of the estuary. To fill this gap, four 7.5-minute topographic quadrangle maps from the USGS) were georeferenced and digitized. These quadrangle maps were part of a

Table 1. Summary of shoreline data used in this study and the associated uncertainty.

Shoreline	Source	Digitization error (m)	1:20,000 error (m)	1:10,000 error (m)	Georeferencing error (m)	Total positional error (m)
1879	NOAA T-Sheet	-	-	8.00	-	8.00
1948	NOAA T-Sheet	-	10.00	-	-	10.00
1946	USGS Quadrangle	0.50	-	-	3.13	3.17
1991	NOAA Digital	-	10.00	-	-	10.00
2007	USGS Orthophotography	0.50	-	-	1.83	1.90
2012	USGS Orthophotography	0.50	-	-	1.24	1.34

series published in 1949 using hydrography and topography from 1946 aerial photographs. The quadrangles were georeferenced together, and the high water line (HWL) shoreline digitized in reference to NAD 1983 Delaware State Plane at a scale of 1:24,000. The average RMS error associated with georeferencing the quadrangles was ±3.13 m and the digitization error was ±0.50 m.

A digital shoreline product was available for the period 1991−1992 (1992), NOAA's Medium Resolution shoreline. This continuous shoreline is composed of shorelines digitized from nautical charts published in 1991−1992 and is the same shoreline product used to construct the location map shown in Figure 1. The scale of NOAA charts for the Delaware Estuary region is generally 1:20,000, though some are scaled at 1:10,000. At these scales the associated error of positional uncertainty is between ±10m.

A shoreline from 2007 was digitized from Delaware orthophotography at 0.25 m resolution and a horizontal accuracy of 1.52 m. These photos were taken from a series of flights taken between 3/19/2007 and 4/20/2007 by the USGS, State of Delaware, Delaware Environmental Monitoring Analysis Center (DEMAC), and Delaware Geological Survey (DGS). The New Jersey Office of Information Technology (NJOIT) took orthophotography at the same resolution during the same time period to obtain imagery in New Jersey. The horizontal accuracy of this flight was 1.21 m. The imagery from both states was digitized and used by Walsh (2004). The positional error of the photos corresponds to a georeferencing RMS error of ±1.83 m. The error associated with digitization of the shoreline is ±0.50 m.

The most recent Delaware shoreline used for this study (2012) was constructed from high-resolution USGS orthophotography. These photos were taken at a resolution of 0.3 m, a horizontally accuracy of 1.52 m and a root-mean-square error (RMS) of 0.88 m. High-resolution orthophotography for the New Jersey portion of the study area was also obtained through the USGS and has the same pixel resolution as the Delaware imagery. The RMS was 0.88 m and had a horizontal accuracy of 1.22 m. The positional error of the photos corresponds to a georeferencing RMS error of ±1.24 m. The error associated with digitization of the shoreline is ±0.50 m. These photographs were taken in April 2012, before Hurricane Sandy. Following the methods described in the literature (Boak and Turner, 2005; Crowell et al., 1991), shorelines were traced digitally following the HWL, represented as discoloration of beach sand or line of debris marking the previous high-tide line. In areas where the HWL was not visible, such on vegetated wetland coast, the marsh edge was taken as the shoreline.

2.1.2 Rates of Shoreline Change

The Digital Shoreline Analysis System (DSAS) was used to quantify shoreline change over periods of interest. Developed by the USGS, DSAS automatically calculated the rates of shoreline change in ArcMap 10.2 (Thieler and Danforth, 1994; Thieler et al., 2009). DSAS calculates a rate of change for a user-defined number of transects perpendicular to an arbitrary baseline roughly parallel to a series of shorelines to be analyzed. Both end-point rates (EPR) and regression-based rates (LRR) of shoreline change can be computed by DSAS for each transect. EPR rates are determined by taking the distance between any two shorelines and dividing by time difference between the years of survey. LRR rates are computed from the slope of a least-squares regression line fit to a plot of several shoreline positions, relative to the baseline, versus the year of survey (see inset in Figure 5). Although the EPR

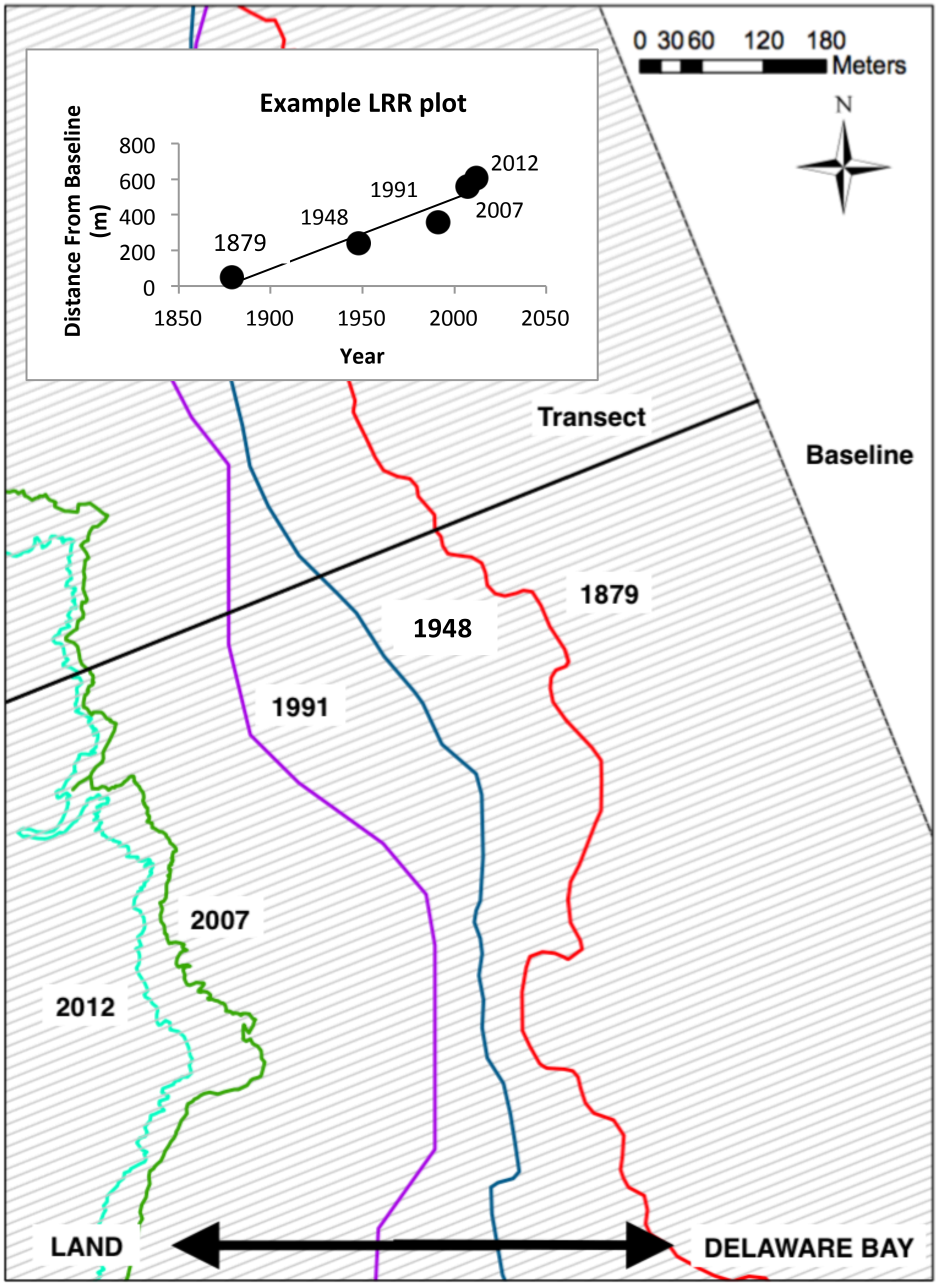


Figure 5. Schematic of the DSAS method described in the text. Shown are the five shorelines and baseline-perpendicular transects (grey lines) spaced at 10 m. The time-distance relationship for one transect (bold black line) is plotted in the inset. The slope of the regression line gives the LLR-based rate of shoreline change.

method has an advantage of being simple, shoreline change by LRR is considered more robust because rates are based on more than two data points, minimizing random error and short-term variability in shoreline position (Thieler et al., 2009).

For the present study both EPR and LRR rates were computed. The LRR method was used to determine long-term (1879–2012) rates of change using all six of the shorelines (Table 1). The EPR method was used to compute short-term rates between each of shoreline endpoint years (1879–1948, 1948–1991, 1991–2007, 2007–2012), as well as long-term rates (1948–2012, 1879–2012). Comparison of shoreline change rates determined using the LRR and EPR methods was used to assess the potential for temporal trends in rate.

Shore-perpendicular transects were spaced 10 meters apart and 3,000 m in length, sufficient to cover the distance between the 1879 and 2012 shorelines. This amounted to a total of 26,108 transects for the lower estuary–bay study area. A landward shift in shoreline position was interpreted to represent "retreat" with negative rates of change, whereas a seaward shift was taken to signify "extension" with positive rates. Although in many shoreline change studies landward and seaward shifts are ascribed to "erosion" and "accretion", these terms invoke processes of sedimentation that were not validated in this study. Erosion was to describe the nature of shoreline change only in the case of retreating shores that displayed erosional morphologies, based on inspection of aerial photographs or field observations.

An example of the DSAS method is shown in Figure 5. For each EPR and LRR calculation of shoreline change less than 15% of transects were removed from the analysis. Transects were removed because of failure to intersect more than one shoreline, transect crisscrossing, failure to calculate rate of change based on the most seaward shoreline, or intersection where the shoreline was not perpendicular to the transect.

2.1.3 Error Analysis

Uncertainties in shoreline change analysis associated with high water line (HWL) position and shoreline change rate are detailed by Hapke et al. (2010) and references cited therein. Uncertainties in HWL shoreline data are generally related to mapping methodology, conversion of shoreline position to geographic coordinates, and map digitizing. The sources and magnitudes of position uncertainties have been addressed by a number of authors (Crowell et al., 1991; Thieler and Danforth, 1994; Moore, 2000; Shalowitz, 1964), and for the present study established values were used to compute the total positional and shoreline change rate uncertainties.

The positional uncertainty for each shoreline (E_{sp}) is given by Equation 2 where E_g is the georeferencing error (±4 m), E_d is digitization error (±1 m), and E_t is the T-sheet error (3–10 m). Due to different sources of data for each shoreline, not all of these terms were used to calculate the positional uncertainty for each shoreline.

$$E_{sp} = \sqrt{E_g^2 + E_d^2 + E_t^2} \tag{2}$$

The uncertainty for EPR and LRR shoreline change rates for each transect is given by Equation 3 where the total positional error (E_a) is the square root of the sum of squares of E_{sp} for the two shorelines, divided by the number of years between the surveys. The uncertainty for each transect per analysis is the same because uniform datasets were used.

$$E_a = \frac{\sqrt{E_{sp1}^2 + E_{sp2}^2}}{time} \tag{3}$$

Table 2. Total change rate uncertainties from EPR analysis with the exception of 1879-2012[1] by LRR analysis given by Equation 3.

Time period	Change rate uncertainties (m/yr)
1879 to 2012	0.06
1879 to 1948	0.19
1948 to 1991	0.34
1991 to 2007	0.64
2007 to 2012	0.47
1948 to 2012	0.17
1879 to 2012[1]	0.13

[1]Regression-based analysis

2.1.4 Shoreline Characterization

To help interpret shoreline change rates, shores of the lower estuary and bay were classified based on visual examination of the 2012 series of orthophotographs available from the USGS. The following shoreline classes were selected: barrier beach; tidal wetland; transitional (between beach and wetland); and hardened (Figure 6a-d). These classes were chosen based on prior descriptions of the estuary coast in the literature (e.g., Fletcher et al., 1990), and maps published by the USGS, the state of Delaware Geological Survey, and the state of New Jersey Geological survey. These shoreline types are readily distinguishable in the aerial photographs used for this study. Barrier beaches were identified based on presence of continuous sandy coasts composed of sand, whereas wetland shorelines were usually highly vegetated with a darker sediment substrate in the fronting intertidal area. Transitional shorelines were characterized by discontinuous patches of beach sand between more continuous segments wetlands—these shorelines are likely to evolve into barrier beach coasts with continued transgression. Hardened shorelines were categorized based on engineered structures such as docks, jetties, groins, and bulkheads.

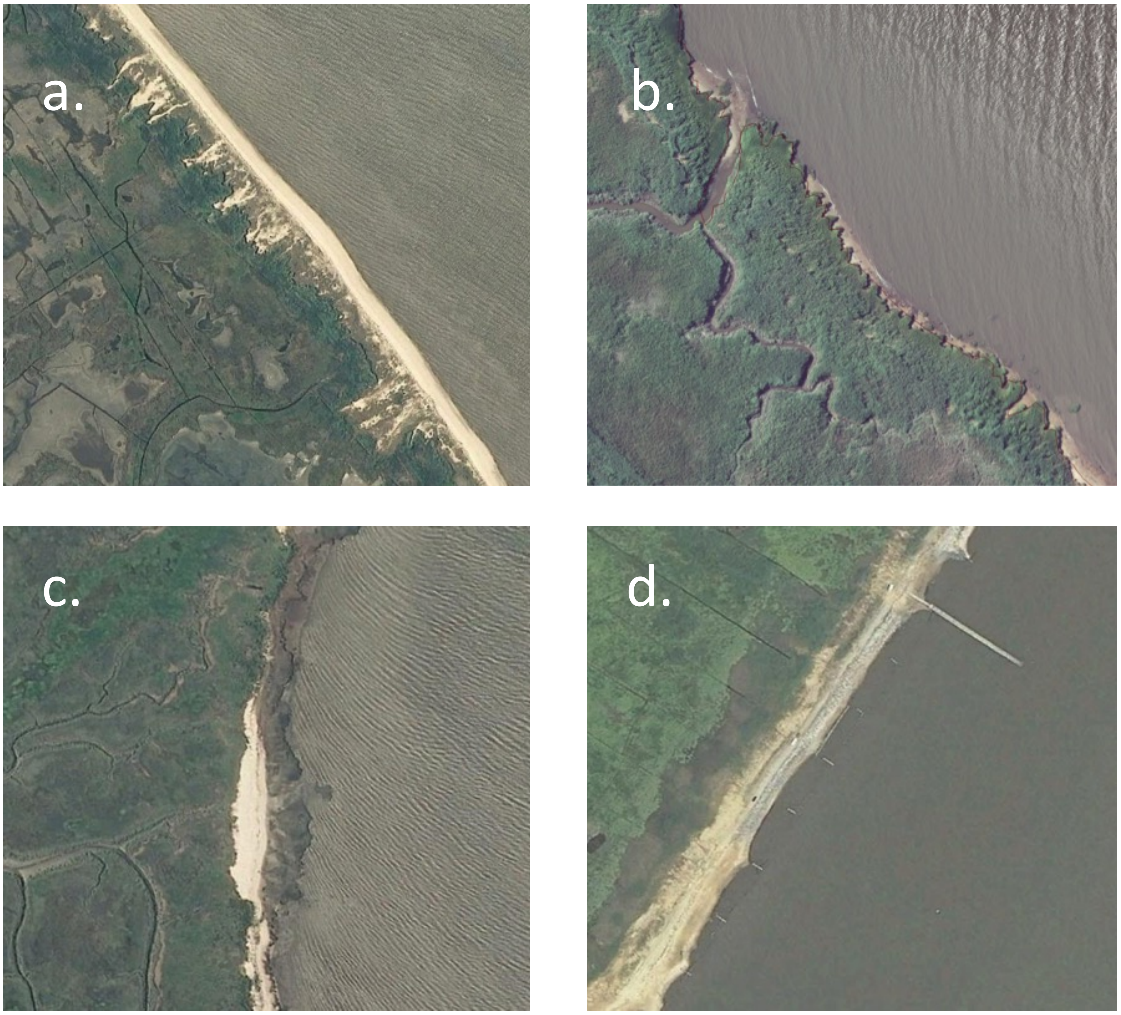


Figure 6. Selected aerial photographs showing examples of the four shoreline classes defined for this study: a) barrier beach; b) tidal wetland; c) transitional; and d) hardened.

2.1.5 Volumetric Change Analysis

The change in shoreline volume associated with shoreline movement was estimated for study as means to determine the mass of sediment lost and gained through retreat and extension if by erosion and accretion, respectively. Volumetric change was computed from the distance between shoreline endpoints (length), 10-m transect spacing (width), and spring tide range (height). Total volumetric change (per meter of shoreline) can be derived from shoreline change as $\Delta V = H \Delta S$, where ΔV is total volume change, H is the height of the shore profile, and ΔS is the change in the MHW shoreline position. Estimating total volume change in this way is common practice when direct measurements of volume change are unavailable (Farris and List, 2007).

NOAA tide stations (8557380, 8536110, 8537121, 8551910, 8551762, and 8545240) were used to calculate the average tidal range for the lower estuary and bay regions. Using methods described by Hobbs et al. (1992), volumetric change was converted to sediment mass using dry-bulk density data available from numerous cores collected in prior research in the lower estuary and bay (Sommerfield, unpublished data). Because the error associated with dry-bulk density measurements is negligible compared to other possible sources of error, the percent error determined for the shoreline uncertainty was used to estimate the error associated with the volumetric (and mass) change.

2.2 Wave Data and Analysis

To characterize wind-wave characteristics in the lower estuary and bay, data obtained by two Triaxys Directional Wave Buoys by the Department of Natural Resources and Environmental Control (DNREC) and University of Delaware were analyzed. This was undertaken to identify along-estuary variations in wave parameters (significant wave height, wave period, and mean wave direction) that could convey information on spatial patterns of wave power (Equation 1) at the coast. The locations of these buoys are shown in Figure 1. During the first deployment (May 2007–February 2012), the buoys were maintained by DNREC in cooperation with NOAA and designated NDBC44054 "Lower Bay" and NDBC44055 "Upper Bay" (Brown and Leathers, 2013; Jenkins et al., in review). The buoys collected spectral wave data in 20-minute hourly bursts from which wave parameters were computed in 1-hr means. Although the wave data time series were not continuous over the five-year deployment, as discussed by Brown and Leathers (2013), sufficient data were available to develop seasonal wave climatology for both buoy stations.

The same two wave buoys were deployed by the University of Delaware in the lower estuary in March 2014, serviced in August 2014, and redeployed until March 2015. Unfortunately, Buoy 1 was lost in late February 2015 due to extensive icing in the estuary, thus data for only the March–August period were available for analysis. The time series for Buoy 2 includes full seasonal cycle for 2014–2015. For the University of Delaware deployments, wave parameters were computed from wave spectral data collected in hourly in ten-minute bursts. Data from all of the buoy deployments described above were used to identify seasonal variations in wave parameters binned in four, three-month intervals as follows: Winter (January–March); Spring (April–June); Summer (July–September); Fall (October–December).

Chapter 3

RESULTS AND INTERPRETATION

3.1 Distribution of Shoreline Types

The shoreline classification study yielded results that were consistent with our general understanding of the system: the lower estuary and bay are characterized by mostly tidal wetland and barrier beach coasts, respectively (Figure 7). The transition from a predominantly barrier beach coast to wetlands occurs at Port Mahon on the Delaware side and near Fortescue on the New Jersey side. In the lower estuary, there were a larger number of barrier beach and transitional shore segments on the Delaware side, and these segments were generally more continuous than on the New Jersey side. There was no obvious relationship between shoreline morphology and shoreline class in the lower estuary; however, transitional shores tended to occur within embayments such as coves and river mouths. This suggests that marine transgression, and the evolution of wetland coasts to barrier beaches, is highly localized and does not occur in a continuous wave-like fashion from the bay to lower estuary.

About 59% of the lower estuary and bay study are composed of tidal wetland shorelines. Barrier beaches comprise 27% of the coast followed by transitional (4%) and hardened shores (10%). About 71% of the hardened shores overall fell on the New Jersey side of the lower estuary, particularly in the northern portion. Rates of shoreline change specific to these shoreline classes is discussed in Section 3.3.

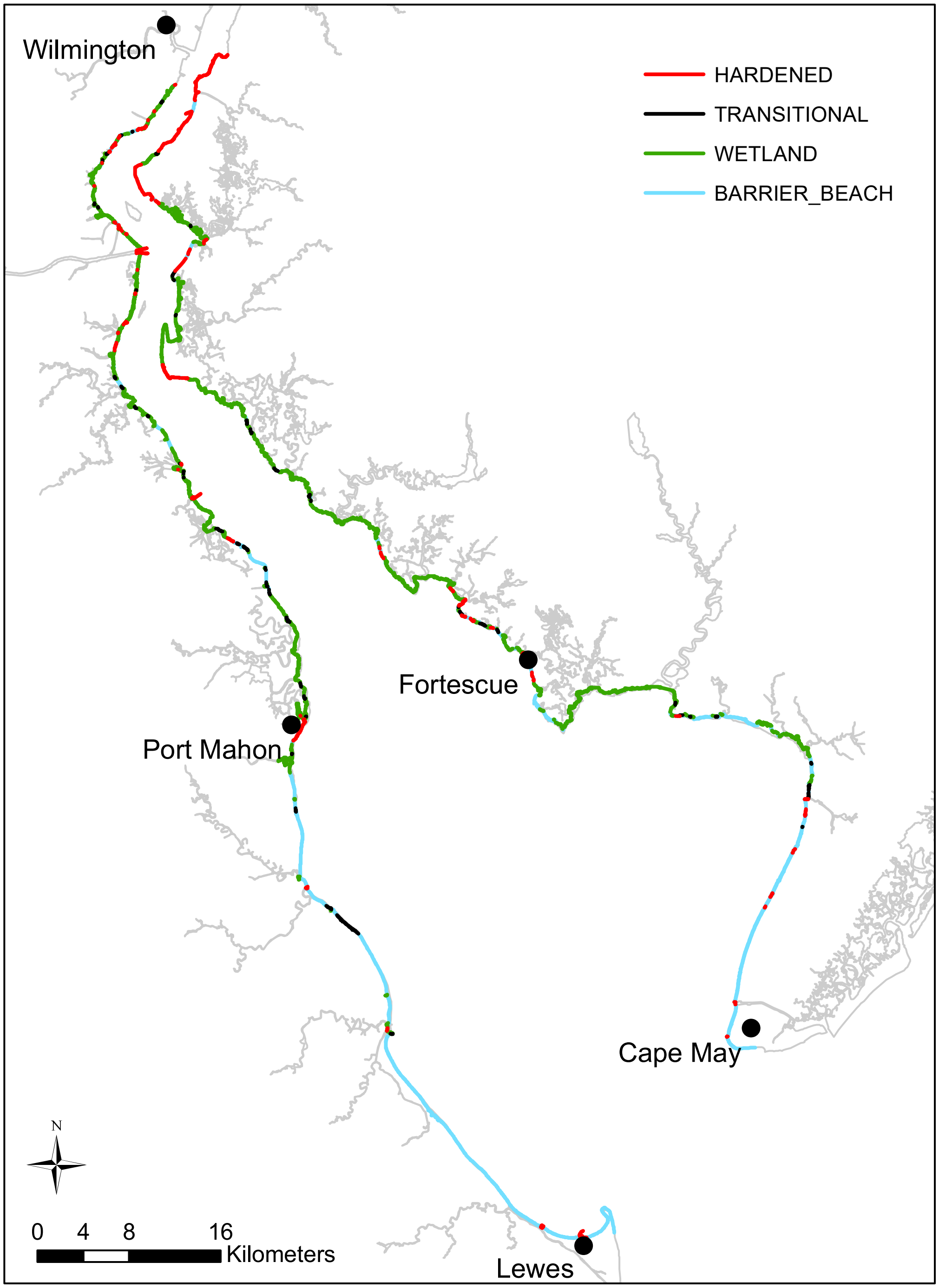


Figure 7. Distribution of shoreline classes in the lower Delaware estuary and bay.

3.1.1 Long-term Rates of Change

Based on LRR regression analysis, from 1879 and 2012 the long-term rate of shoreline change for the greater lower estuary and bay was -0.68 ± 0.13 m/yr. The long-term rate based on EPR analysis was comparable at -0.79 ± 0.06 m/yr. Overall, 77% of the shore has retreated, 24% has extended, and less than 0.5% has remained unchanged since 1880. Hence, the shoreline of Delaware Estuary as a whole has been in a state of net retreat. Mean LRR rates for both the Delaware and New Jersey sides of the lower estuary−bay were equivalent at -0.68 ± 0.13 m/yr, but when Artificial Island and the Killcohook confined disposal facility are removed from consideration (land reclamation is these areas has caused the shoreline to extend into the estuary as shown in Figure 8), the mean rate of retreat was over two times higher on the New Jersey side (-1.5 ± 0.13 m/yr). In addition to a more rapid rate of retreat, the New Jersey shore exhibited a wider range of variability in shoreline change, from -11.6 ± 0.13 m/yr to 16.8 ± 0.13 m/yr. Results of the shoreline change calculations by LRR and EPR analysis are summarized in Tables 3−5.

The spatial pattern of long-term shoreline change based on LRR analysis is shown in Figure 8. On the Delaware side the highest retreat rates were centered at Bombay Hook National Wildlife Refuge, just north of Port Mahon, whereas the highest rate of extension occurs south of Port Penn, directly across-estuary from Artificial Island. On the New Jersey side the highest rates of retreat occurred on the western side of Egg Island south of Fortescue (Figure 8). Retreat rates were highest at coastal promontories on the New Jersey side; however; rates were locally high within some embayed segments (Figure 8).

Table 3. Rates of long- and short-term shoreline change averaged over the entire estuary-bay study area based on LRR and EPR analysis.

Date Range	Mean distance (m)	Mean rate (m/yr)	Change rate uncertainty (m/yr)
Lower estuary-bay LRR			
1879-2012	-96.7	-0.68	0.13
1879-2012[a]	-149.3	-1.1	0.13
Lower estuary-bay EPR			
1879-1948	-6.3	-0.11	0.19
1879-1948[a]	-58.9	-0.95	0.19
1948-1991	-40.5	-0.89	0.34
1991-2007	-38.0	-2.5	0.64
2007-2012	-10.7	-2.1	0.47
1948-2012	-83.2	-1.3	0.17
1879-2012	-96.7	-0.75	0.06

[a] With the Artificial Island and Killcohook shoreline segments removed

Table 4. Rates of shoreline change averaged over the lengths of the Delaware and New Jersey coasts of lower estuary–bay from LRR and EPR analysis.

Period	Mean distance (m)	Mean rate (m/yr)	Minimum rate (m/yr)	Maximum rate (m/yr)	Change rate uncertainty (m/yr)
Delaware LRR					
1879-2012	-101.1	-0.68	-8.2	5.1	0.13
Delaware EPR					
1879-1948	-36.4	-0.58	-19.4	18.3	0.19
1948-1991	-10.1	-0.22	-28.5	17.5	0.34
1991-2007	-40.2	-2.6	-29.3	26.6	0.64
2007-2012	-11.8	-2.4	-40.1	8.3	0.47
1948-2012	-57.6	-0.90	-40.1	13.0	0.17
1879-2012	-101.1	-0.79	-8.2	7.7	0.06
New Jersey LRR					
1879-2012	-92.6	-0.68	-11.6	16.8	0.13
1879-2012[a]	-197.9	-1.5	-11.6	4.7	0.13
New Jersey EPR					
1879-1948	21.7	0.33	-36.9	36.5	0.19
1879-1948[a]	-81.5	-1.3	-36.9	9.4	0.19
1948-1991	-68.4	-1.4	-44.2	50.5	0.34
1991-2007	-35.9	-2.3	-30.9	34.7	0.64
2007-2012	-9.7	-1.9	-38.2	13.8	0.47
1948-2012	-109.4	-1.7	-25.4	31.7	0.17
1879-2012	-92.6	-0.72	11.6	15.2	0.06

[a] With the Artificial Island and Killcohook shoreline segments removed

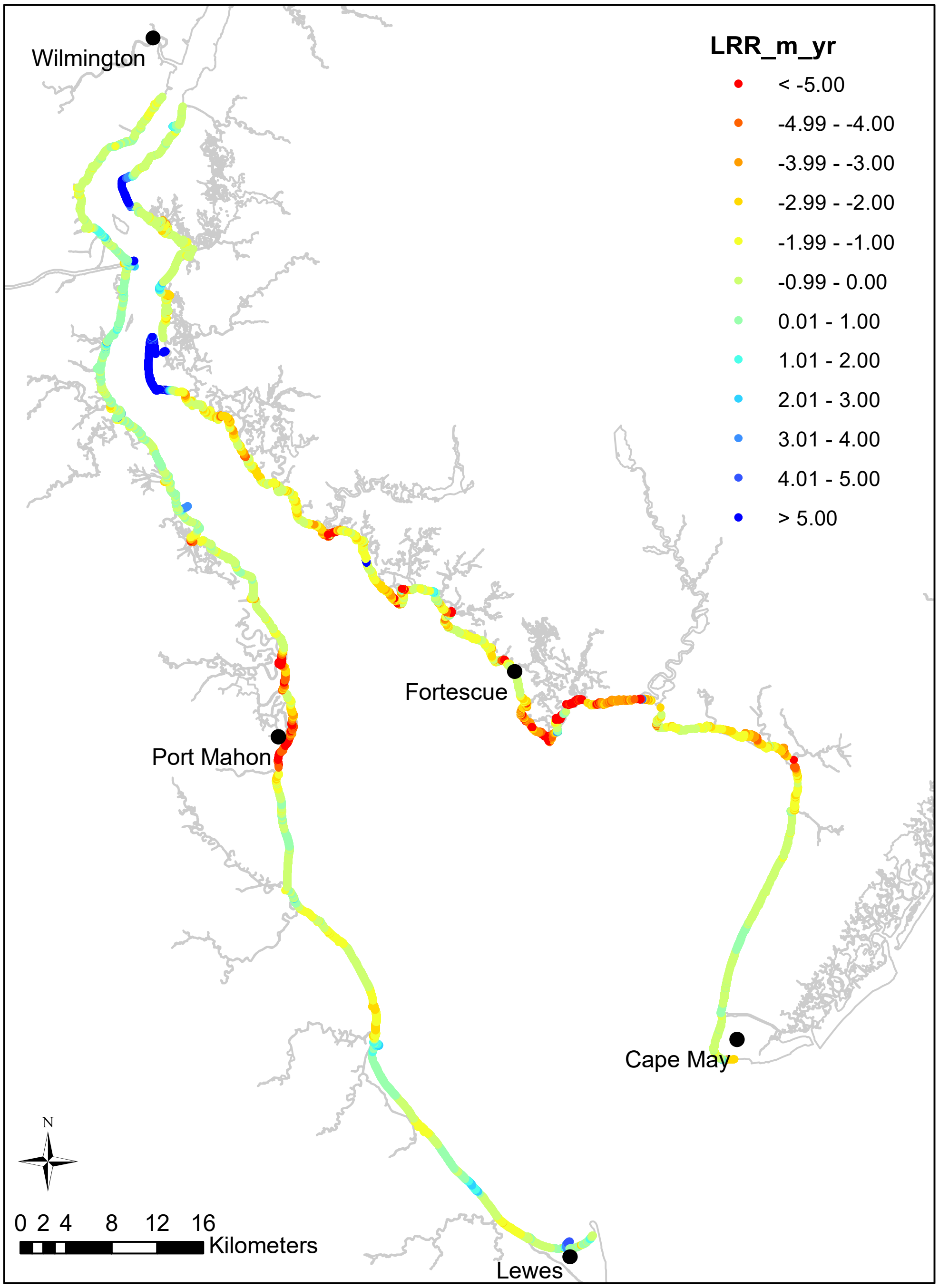


Figure 8. Map of the spatial distribution of long-term rates of change in m/yr determined by linear regression of the five shoreline datasets (1879–2012).

3.1.2 Short-term Rates of Change

The pattern of the most recent shoreline change (2007–2012) based on EPR analysis is shown in Figure 9 with the supporting data listed in Tables 3–5. The short-term pattern of change was broadly comparable to the long-term (1879–2012) pattern, and similarly indicated that the lower estuary–bay shore as a whole is retreating, albeit at a higher rate of -2.13 ± 0.47 m/yr. Between 2007 and 2012, 79% of the coast retreated, 21% extended, and <0.25% experienced no change. Mean rates of retreat for the Delaware and New Jersey shores were -2.35 ± 0.47 m/yr and -1.91 ± 0.47 m/yr, respectively (Table 4). Interestingly, the short-term rate of retreat was somewhat higher on the Delaware side of the estuary, just the reverse of the long-term trend. This higher short-term rate is conveyed by the relatively large number of retreat hotspots along the length of the lower estuary–bay coast (Figure 9).

Similar to the long-term results, some of the highest rates of short-term shoreline retreat on the Delaware side fell just north of Port Mahon in the Bombay Hook area, a well-known hotspot of erosion on the Delaware coast. However, not expected were hotspots of retreat in the southern portion of the bay at Prime Hook and between the Mispillion and St. Jones rivers (Figure 9). On the New Jersey side, the highest rates of short-term retreat fell in the same general area of the highest long-term retreat rates, Little Egg Island. The area with the highest rate of shoreline extension on the New Jersey side fell at a barrier beach on the northern part of Cape May National Wildlife Refuge.

Averaged over 1879–2012, it is apparent that coasts of the lower estuary and bay have been retreating at comparable rates (Table 5). This was an unexpected result given that the standing conceptual model holds that the lower estuary region is in a constructive phase of evolution versus destructive for the bay (Fletcher et al., 1990).

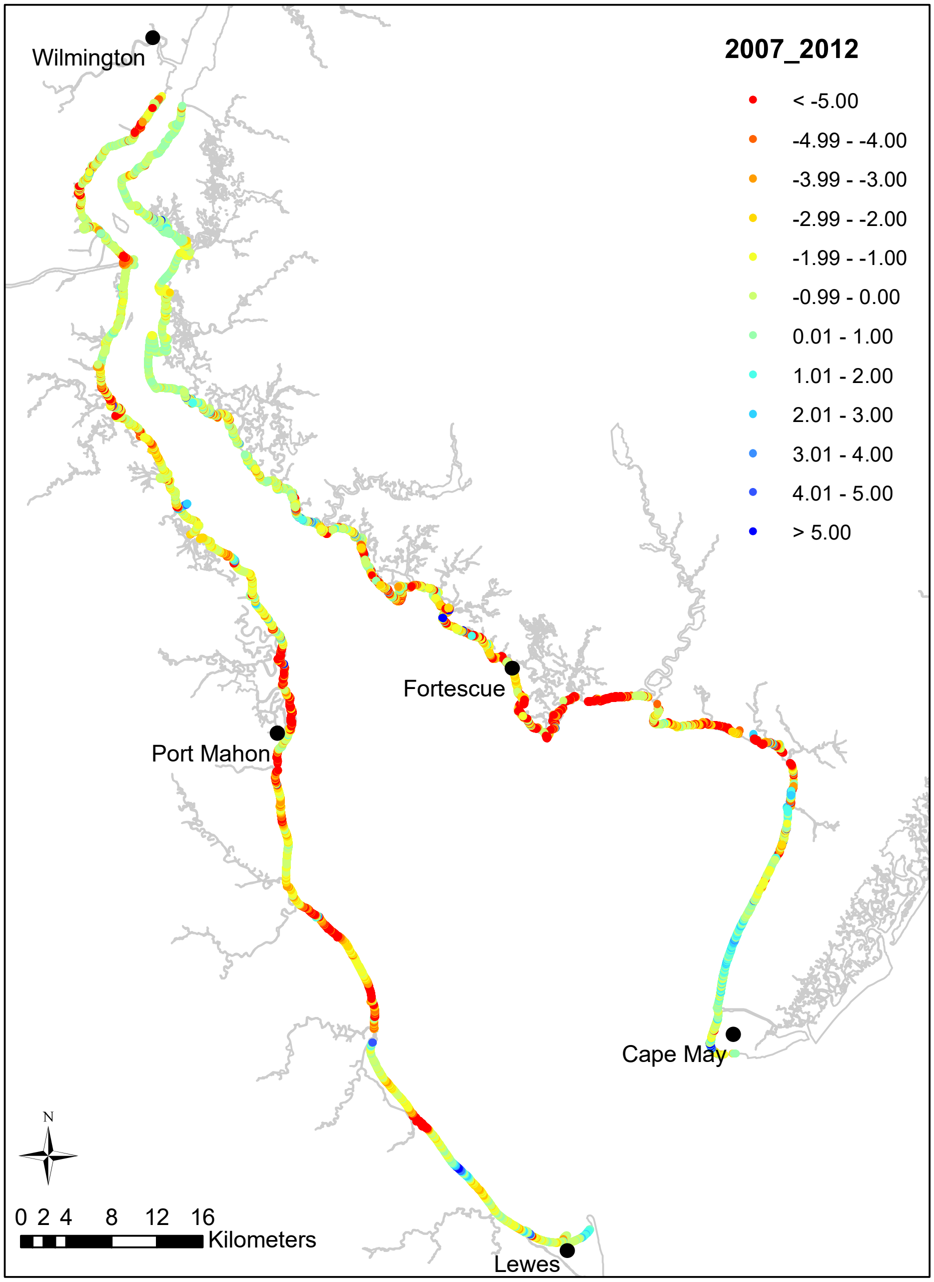


Figure 9. Mean of the spatial distribution of short-term rates of change in m/yr determined by end-point rate analysis of the 2007 and 2012 shoreline.

Table 5. Comparison of shoreline change rates averaged over different timespans and regions of the study area.

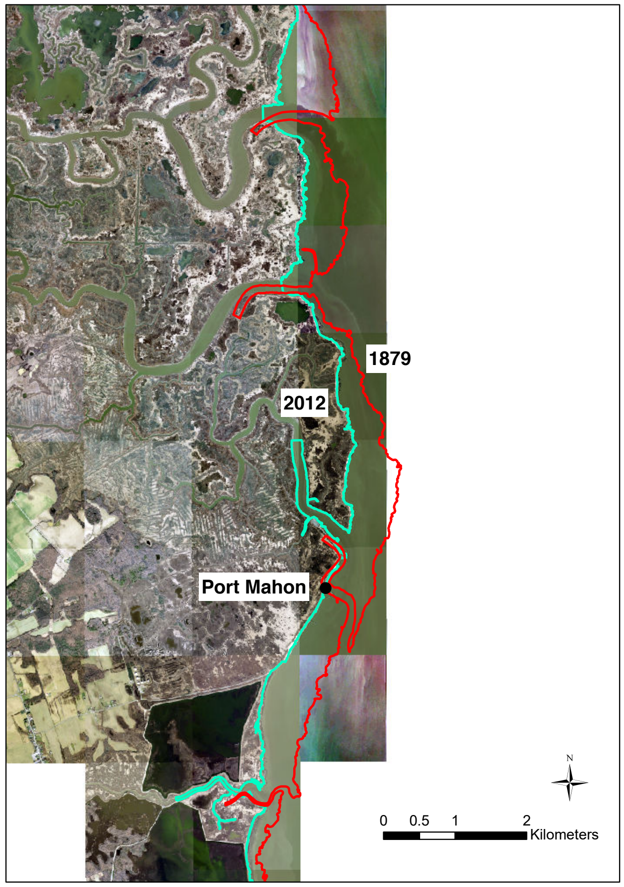
Area	Subarea	1879–2012 LRR rate (± 0.13 m/yr)	2007–2012 EPR rate (± 0.47 m/yr)
Lower estuary			
	Delaware	-0.64	-2.4
	New Jersey[a]	-1.3	-1.1
	both coasts[a]	-1.1	-1.7
Bay			
	Delaware	-0.73	-2.3
	New Jersey	-1.7	-2.9
	both coasts	-1.3	-2.7

[a] With the Artificial Island and Killcohook shoreline segments removed

In contrast, averaged over 2007–2012 the bay shoreline as whole retreated more rapidly (-2.7± 0.47 m/yr) than the lower estuary shoreline (-1.7± 0.47 m/yr). The difference between the long-term and short-term averages was largely due to shoreline retreat on the Delaware side of the lower estuary, and retreat on both the Delaware and New Jersey sides of the bay (Table 5). The main points of this result are that (1) the lower estuary coast is in a destructive stage, and (2) rates of shoreline retreat through the lower estuary–bay may have increased in recent times.

It is worth taking a closer look at hotspots of retreat in the study area, given that local rates are considerably higher than the regional mean values. On the Delaware side, the most prominent hotspot of retreat is the area of Port Mahon in Bombay Hook National Wildlife Refuge. Since 1879, retreat by erosion has led to up to a kilometer of shoreline erosion at Bombay Hook (Figure 10). On the New Jersey side, the area of Egg Island is an erosional hotspot where up almost two kilometers of shoreline retreat has taken place. Rates of retreat (erosion) in the vicinity of these hotspots are very site-specific and do not necessary represent conditions in the entire system.

In addition to hotspots of retreat, in the Delaware Bay there are localized spots of extension (Figure 9). Some of this extension could be due to beach nourishment projects in addition to natural deposition of sand. On the Delaware side, Pickering Beach, Kitts Hummock, Mispillion River, Port Mahon, Bowers Beach, South Bowers, Slaughter Beach, Primehook Beach, and Broadkill Beach are areas of past and ongoing beach nourishment projects (DNREC, 2010) that can explain shoreline extension during the 2007–2012 shoreline change period.



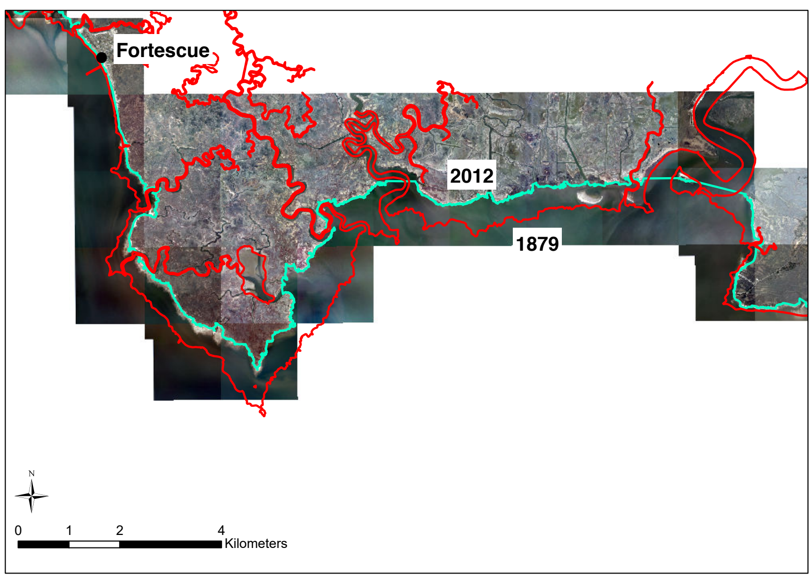


Figure 10. Shoreline segments from Bombay Hook (top) and Little Egg Island (bottom) showing the extent of shoreline retreat between 1879 (red line) and 2012 (green line).

3.2 Spatial Variability in Shoreline Change Rates

Results of this study make clear that there is considerable temporal and spatial variability in shoreline change in the Delaware Estuary. In an effort to quantity this variability, the standard deviation of the four EPR+ change rates (1879–1948, 1948–1991, 1991–2007, 2007–2012) determined for each of the 10-m spaced transects was computed. Shoreline change standard deviation is a measure of “shore mobility”, in other words, short-term variations in retreat or extension driven by transient erosional or accretionary processes (Dolan et al., 1979). Shore segments with low standard deviations are locations where rates of retreat or extension are more-or-less steady state. By contrast, segments high standard deviations characterize locations with large variations in retreat and extension rates about the long-term trend.

Plots of shoreline change standard deviation for the lower estuary and bay are shown in Figure 11 along with corresponding plots of change based on LRR analysis. Averaged over the Delaware and New Jersey coasts, standard deviations were 2.5 m/yr and 3.7 m/yr respectively. Hence, shoreline change rates for a given transect along the New Jersey coast are generally more temporally variable. On the Delaware coast the standard deviation in the bay (2.6 m/yr) was somewhat higher than in the lower estuary (2.5 m/yr). The reverse pattern was observed for the New Jersey coast with standard deviations of 3.9 and 3.4 for the lower estuary and bay, respectively. Removal of Artificial Island and Killcohook resulted in a standard deviation of 3.08 for the lower estuary, which is lower than the standard deviation in the bay and is consistent with results for the Delaware side of the estuary. These results suggest that that the Delaware and New Jersey sides of the lower estuary–bay experience comparable levels of temporal variability in shoreline position.

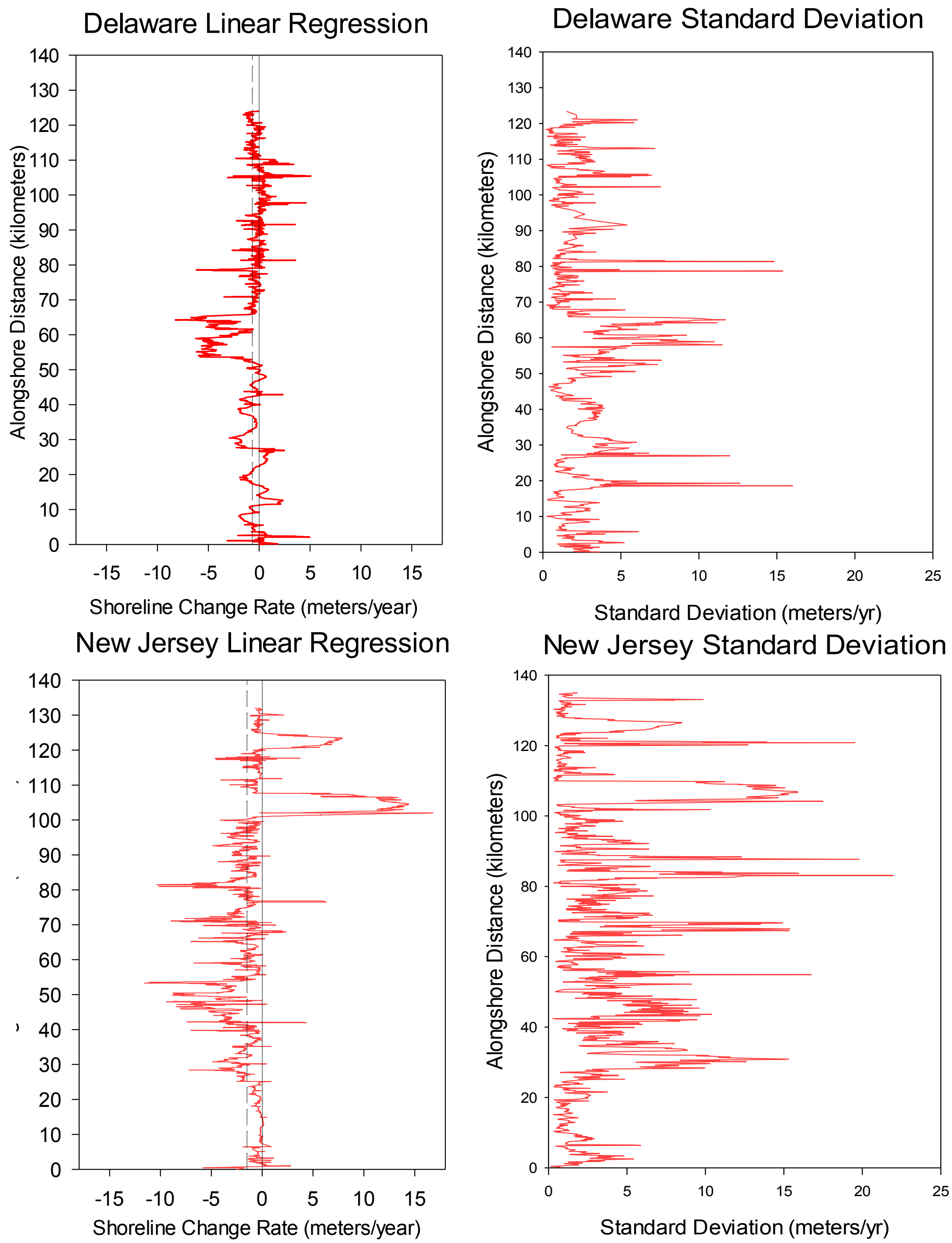


Figure 11. Plots of 1879–2012 shoreline change by LRR (left) and shoreline change standard deviation (right) for the Delaware and New Jersey coasts of the estuary. See text for interpretation.

Insight into the nature of shoreline change in the estuary can be gained by averaging the data according shoreline class. These results are shown in Figure 12 and listed in Table 6. Rates of long-term change averaged over the total area of barrier beach shoreline in the lower estuary and bay are considerably lower than rates for the transitional and wetland shore classes, a result that holds for both the Delaware and New Jersey sides. Hence, despite the fact long-term retreat rates in the bay are somewhat higher than in the lower estuary (data in Table 5), the barrier beach shores appear to be less mobile, or are more capable of recovering from disturbance by accretion, than transitional and wetland shores. This condition is most likely related to differences in sedimentary processes specific to sandy beach versus muddy wetland and transitional shores. For example, whereas sandy beaches can recover after storm-produced erosion events by subsequent accumulation of sand, for tidal wetland coasts to recover after edge erosion requires a particular balance between mud sedimentation on the fronting tidal flat and plant growth.

The high rates of retreat for the transitional shores (Figure 12) relative to wetland are more difficult to explain. It may be the case that wave-produced erosion on some wetland shores is particularly intense such that sand scoured from the underlying strata can become concentrated within intertidal zone. This could explain the coexistence of sands and muds that characterized the “transitional” nature of these shores. Lastly, because the hardened shores represented in Figure 12 do change in the same ways as the other types of shorelines, the rates of change are not directly comparable.

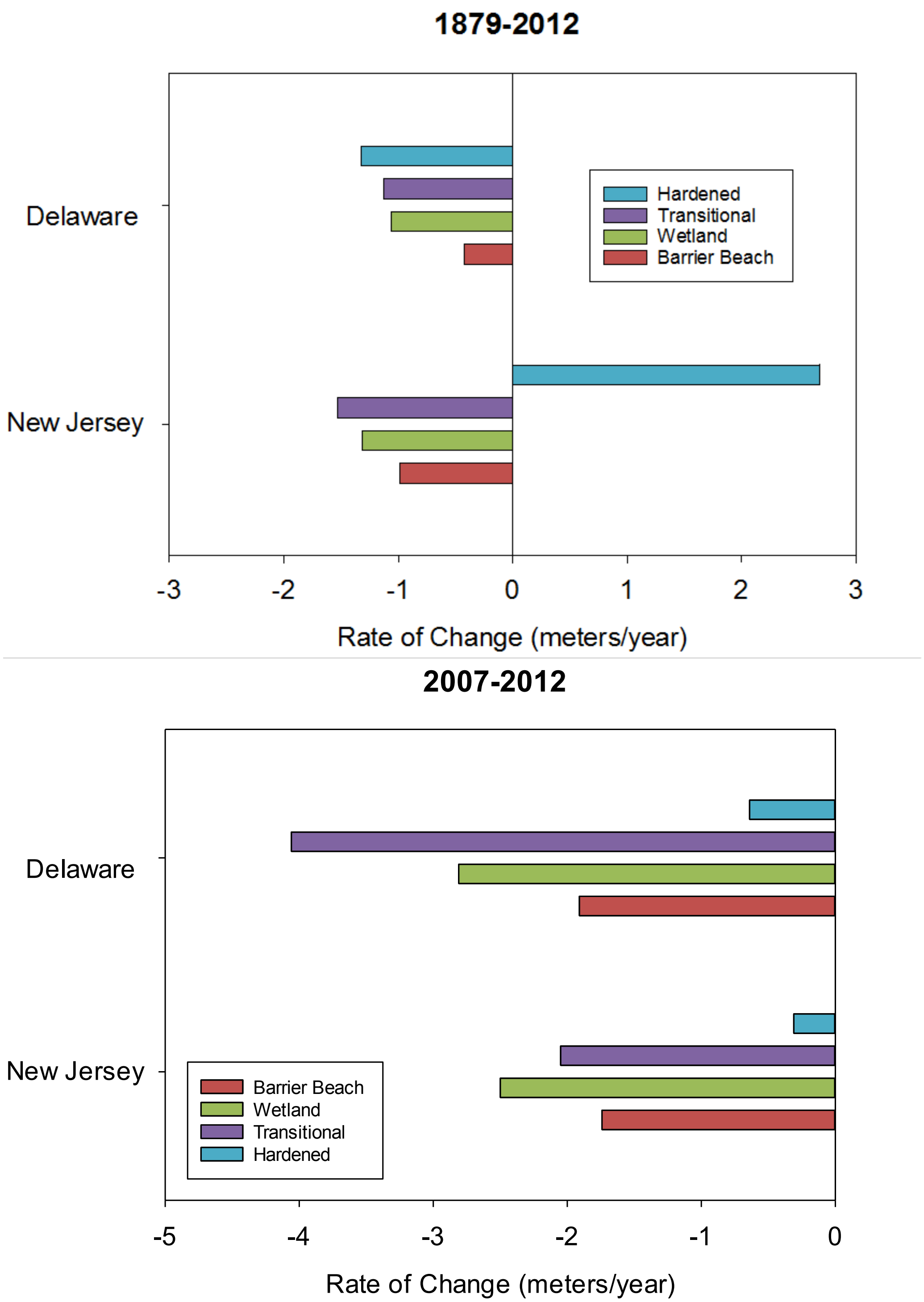


Figure 12. Long-term and short-term shoreline change averaged over the extent of the four shoreline classes in the lower estuary and bay.

Table 6. Long-term rates of shoreline change binned according to shoreline class. Rate uncertainties are ± 0.13 m/yr for LRR and ± 0.47 m/yr for EPR.

		Barrier Beach	Wetland	Transitional	Hardened
Delaware 1879-2012 LRR	min	-6.1	-8.2	-6.02	-6.7
	max	4.5	5.01	3.6	7.7
	mean	-0.43	-1.07	-1.1	-1.3
New Jersey 1879-2012 LRR	min	-11.6	-9.3	-3.9	-6.8
	max	7.1	15.2	0.62	17.7
	mean	-0.99	-1.3	-1.5	2.9
Both coasts LRR	mean	-0.67	-1.2	-1.3	1.5
Delaware 2007-2012 EPR	min	-40.1	-30.5	-20.7	-15.4
	max	7.1	8.3	3.5	6.5
	mean	-1.9	-2.8	-4.1	-0.64
New Jersey 2007-2012 EPR	min	-25.4	-36.8	-18.0	-13.1
	max	11.0	13.8	5.4	6.5
	mean	-1.7	-2.5	-2.1	-0.30
Both coasts EPR	mean	-1.8	-2.6	-3.2	-0.40

3.3 Wave Characteristics in the Lower Estuary and Bay

Analysis of wave buoy data indicated that wave properties at the four stations in the lower estuary–bay were not substantially different. Mean significant wave heights ranged from 0.24 to 0.46 m and varied seasonally (Table 7). Significant wave height decreased from NBDC44054 in the bay to Buoy 1 in the lower estuary by about 30% (see Figure 1 for buoy locations). Associated with this decrease is a threefold decrease in mean wave power, from 0.40 kW/m at NBDC44054 to 0.11 kW/m at Buoy 1. The results suggest that wave power available for shoreface erosion generally decreases up-estuary with shoreline convergence. Wave roses show that the largest waves at Buoy1, Buoy 2, and NDBC44055 propagate from the south-southeast to southeast (Figure 12). By contrast, the largest waves at NDBC44054 travel from the east-southeast. This difference is most likely related to the position of NDBC44054, which is exposed to oceanic fetch and swell waves (Jenkins, 2015). Wave heights at Buoy 1 station are distinct among the stations in that they are relatively small and have a less dominant direction of propagation, perhaps due to localized fetch-limitation.

Seasonal analysis of significant wave height indicates a major north-northwesterly component of large waves in winter (see wave roses in Appendix). Although only winter data were available for two of the four buoys, this signal is apparent in data for the fall season though it is less pronounced. Based on the ratio of water depth (h) to wavelength (L), which was computed from measured wave period (L=1.56T^2), the *average* waves measured at the buoy stations are transitional between deep-water (h/L>0.5) and shallow-water types (h/L<0.05), suggesting that they interact with the seafloor. However, both deep-water and shallow-water waves were present at the buoy locations at during times during the observational period.

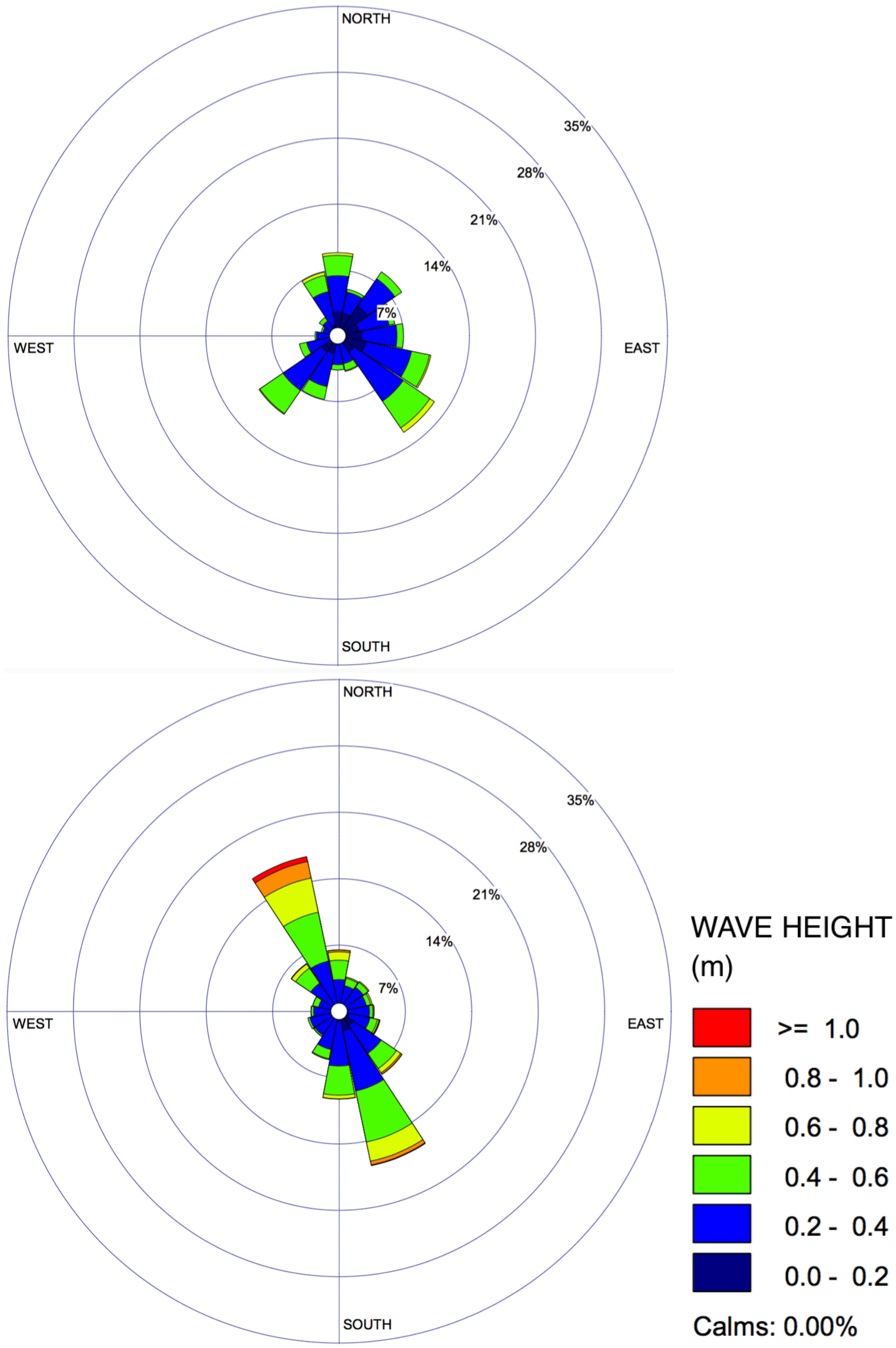


Figure 13. Wave roses of significant wave height with direction of propagation for Buoy 1 (top) and Buoy 2 (bottom) in the lower estuary. See Table 7 for summary of wave data.

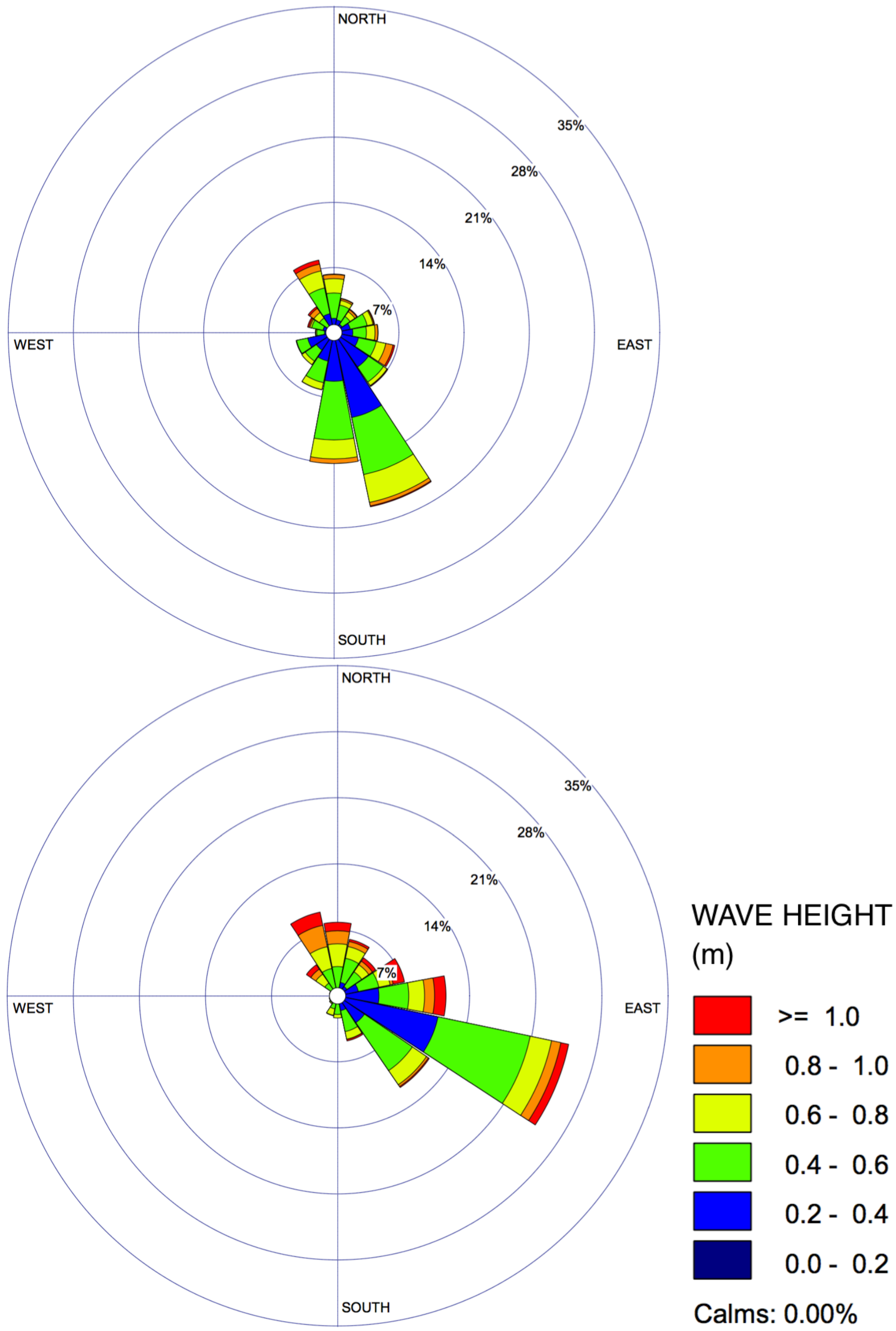


Figure 14. Wave roses of significant wave height with direction of propagation for NDBC44055 (top) and NDBC44054 (bottom) in the bay. See Table 7 for summary of wave data.

Table 7. Wave parameters for the buoy stations discussed in the text.

Parameter	NDBC 44054	NDBC 44055	Buoy 1	Buoy 2
Observations (*n*)	4978	3897	3151	7852
H_s (m)				
mean	0.46	0.37	0.24	0.33
max	2.0	1.4	0.68	2.3
T_s (sec)				
mean	3.9	3.4	4.09	4.3
max	28.5	18.0	9.9	30.6
min	1.5	2.0	1.3	1.5
Wave direction (deg.)				
mean	146	180	162	199
Wavelength (m)				
mean	23.2	18.03	24.8	28.4
max	1267.1	505.4	152.9	1460.7
min	3.5	6.2	2.6	3.5
h/L				
mean	0.32	0.42	0.30	0.26
max	0.01	0.01	0.05	0.01
min	2.14	1.20	2.84	2.14
Wave Power (kW/m)				
mean	0.40	0.23	0.11	0.22
max	55.7	17.2	81.1	2.2

Chapter 4

DISCUSSION

4.1 Variation in Shoreline Retreat Rates

4.1.1 Spatial Variation

Results of this research confirmed what was suspected from the outset that rates of shoreline change in the Delaware Estuary are significantly different than rates reported for the adjacent Atlantic Coast. As noted above, rates of long-term change (1879–2012) in the estuary as a whole were -1.1 ± 0.13 m/yr (net retreat) with similar rates for the lower estuary and bay subregions. Short-term rates of change (2007–2012) showed a similar pattern with a whole-estuary mean of -2.1 ± 0.47 m/yr For the Atlantic coast of Delaware, Hapke et al. (2013) reported long-term and short-term rates of -0.5 m/yr (1845–2000) and -0.8 m/yr (1980–2000), respectively, based on the same LLR and EPR methods used in the present study. Hence, the Atlantic coast of Delaware has been in a state of net retreat but at rates that are somewhat lower than long-term rates determined for the Delaware Estuary in this study. For the Atlantic coast of southern New Jersey, Hapke et al. (2013) reported long- and short-term rates of +0.8 m/yr and +0.2 m/yr, demonstrating that the coast is extensional. This is most likely related to the long history of beach nourishment on this coast and presence of structures designed to trap sand (e.g., Hapke et al., 2013).

Although the short-term rate of shoreline retreat in Delaware Estuary at -2.13 ± 0.47 m/yr is two times faster than the long-term rate of retreat at -1.1 ± 0.13 m/yr, the overall pattern of retreat is similar for both averaging periods, suggesting that the underlying dynamical and geological factors involved are similar. For both periods the highest rates of retreat on the Delaware side of the estuary occurred in the area of Bombay Hook National Wildlife Refuge (north of Port Mahon). On the New Jersey side, the highest rates of erosion took place near Egg Island Fish and Wildlife Management Area (south of Fortescue). Interestingly, both of these erosion hotspots are characterized by wetland-type shores with a high density of tidal channel networks, underlain by Holocene and older fluvial deposits. Both the Cape May Formation on the New Jersey coast and the Beaverdam Formation on the Delaware coast consist of medium to coarse sands and gravels, which are highly porous and thus susceptible to erosion when exposed to current-produced shear stress.

Independent of hydrodynamic and geological factors, higher rates of long-term shoreline retreat on the New Jersey side of the lower estuary–bay could be a consequence of higher rates of land subsidence. Parts of the New Jersey coast may be sinking faster, and thus relative sea level rising faster, than the mean rate of subsidence for the estuary as a whole. Land-elevation surveys by Holdahl and Morrison (1974) indicate that the middle of Delaware Estuary, including the Bombay Hook and Little Egg area, is subsiding at ~3.2 mm/yr, somewhat more rapidly that areas to the north and south (~2.8 mm/yr). Some authors have attributed the higher subsidence rates to groundwater withdrawals (Davis, 1987), but other factors such as sediment compaction and glacio-isostatic adjustment may be at work (Nikitina et al., 2015). Additional research is needed to determine the influence of subsidence processes on patterns and rates of shoreline change in Delaware Estuary.

There was a significant difference in retreat rates of wetland and transitional shoreline classes compared to barrier beaches. These results are consistent with studies in other estuaries showing that wetland-type shorelines can exhibit higher rates of erosion than sandy shores (Cowart et al., 2011; Currin et al., 2015). Similarly, there was a large amount of variability, both spatially and temporally, for wetland coasts on both sides of Delaware Estuary, consistent with observations in other estuaries (Cowart et al. 2010, 2011; Currin et al., 2015 McLoughlin et al., 2015). Wetland and transitional coasts in the lower estuary and bay are retreating by eroding at similar rates, mostly like due to the same wetland type substrate. As marine transgression progresses, the mud depocenter of the estuary migrates landward and gives way to localized sand accumulation over eroded marsh strata, thereby producing a transitional shoreline. The wetland-transitional shorelines retreat more rapidly than the barrier beach shoreline presumably because they are less able to recover after erosion events.

Long-term rates of shoreline retreat determined for this study are somewhat lower than those reported previously. For the Delaware coast of Delaware Bay, French (1990) determined long-term mean rate of -1.4 m/yr, two times the rate determined for the same region in this study (Table 4). However, consistent with the present work, French (1990) found that the wetland shores retreat (erode) more rapidly and more steadily than the sandy, barrier beach shores. Long-term rates of shoreline change calculated by Maurmeyer (1978) are three to four times the average rate of retreat in this study. Maurmeyer (1978) observed the highest rates of retreat on the Delaware estuarine coast (-4.4 to -6.9 m/yr) in the vicinity of Bombay Hook, attributing them to local subsidence. Here it is important to point out that differences in shoreline change rates reported for different studies of the same area can be related to factors including differences in temporal and spatial averaging, different shoreline datasets, and differences in methods of change analysis. For example, averaged over the Bombay Hook area alone, short- and long-term rates of retreat are about -5.0 m/yr and -3.09, respectively, more comparable to the rates reported by Maurmeyer (1978).

Rates calculated by Phillips (1986) in Cumberland County, New Jersey over a 60-year period were averaged at -3.2 m/yr. This is five times the average retreat rate and 1.5 times higher than the average 2007–2012 retreat rate. Athough the average retreat rates for the entire study area are significantly different, this area of the New Jersey coastline was found to have some of the highest rates of erosion. When the rates for the same area of coast that Phillip (1986) are extracted, both the long- and short-term rates of change are more comparable at 2.3 ± 0.13 m/yr and -3.9 ± 0.47 respectively.

Rates of shoreline retreat by processes of erosion in the Delaware Estuary fall within the same range of rates reported for other estuaries. McLoughlin et al. (2015) observed similar rates ranging from -1 to -1.6 m/yr of erosion at Hog Island over a 50-year period. In North Carolina estuaries, average rates of change ranged from -0.24 to -0.58 m/yr over 40–50 years (Cowart et al., 2010; Cowart et al., 2011; Currin et al., 2015). Rates in the Delaware Estuary are generally higher than these rates; however, differences in size of the study are, fetch, and wave power could explain some of these differences.

4.1.2 Temporal Variation

In addition to spatial variation in rates of shoreline change and variation among different shoreline types, there were significant differences in change rates averaged over different time spans. This was examined by plotting the EPR shoreline change rates (data in Table 5) as a function of time span between the shoreline endpoints. As shown in Figure 15, for both the Delaware and New Jersey sides of the lower estuary−bay the longer-term rates of change were lower than the shorter-term rates. This inverse relationship between change rate and time span has been observed in studies of other coastal and estuarine systems, and has be interpreted to reflect that fact that longer-term rates average-out episodes of shoreline retreat and extension related to storm processes (Crowell et al., 1993). By contrast, short-term change rates are more heavily influenced by the magnitude and frequency of storm events, which tend to cluster over time.

The finding that shoreline change rates correlate with averaging time span complicates addressing the question of whether rates of retreat have increased or decreased over time, because it requires rates to be averaged over a similar number of years. Nonetheless. the question of temporal variation in shoreline change rate is important to address given observations of accelerated sea-level rise along the Atlantic coast of the U.S. in recent decades (Boon and Mitchell, 2015). Although plots of EPR shoreline change rate (data in Table 4) for the four periods suggest that retreat rates increased after 1948−1991 (Figure 16a), because the second two rates of retreat (1991−2007, 2007−2012) average over much shorter spans of time than the first two (1879−1948; 1948−1991) this apparent increase is biased by the inverse relationship between rate of change and averaging time span (Figure 15). Rather, rates of shoreline change for 1879−1948 (69 years) and 1948−2012 (64 years) average over a similar number of years and therefore provide a better measure of temporal variability, albeit at a relatively low temporal resolution. As shown in Figure 16b, the mean rate of change (retreat) increased from 1879−1948 (Delaware: -0.58 ± 0.19 m/yr; New

Jersey -1.3 ± 0.19 m/yr) to 1948–2012 (Delaware: -0.9 ± 0.17 m/yr; New Jersey -1.7 ±0.17 m/yr) on both sides of the lower estuary–bay. Paired t-testing of the transect EPR rates for 1879–1948 and 1948–2012 indicates that the overall mean rates of change are significantly different ($P<0.0001$), thus the null hypothesis that there has been no change in retreat rate over time can be rejected. In other words, in all probability there was an increase in retreat rate for the greater lower estuary–bay sometime after 1948. Identifying the possible causes of this change in relation to relative sea-level rise, storm magnitude and frequency, and estuarine processes and geomorphology is beyond the scope of the present study but warrants further investigation.

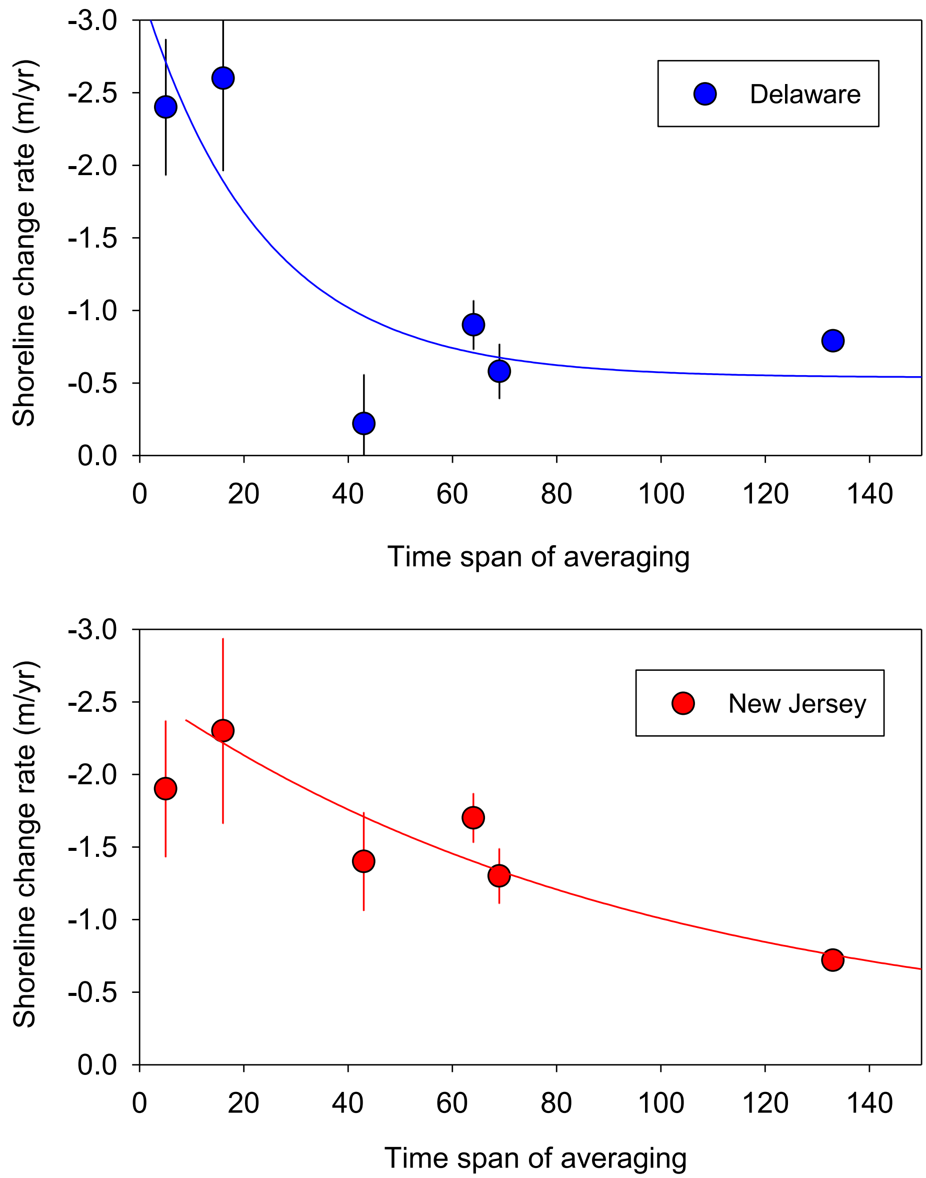


Figure 15. EPR shoreline change rates for the Delaware and New Jersey coasts of the lower estuary–bay plotted as a function of time span between shoreline endpoint dates. In both case shoreline change (retreat) time span of averaging are inversely related. In both cases the trendline is exponential.

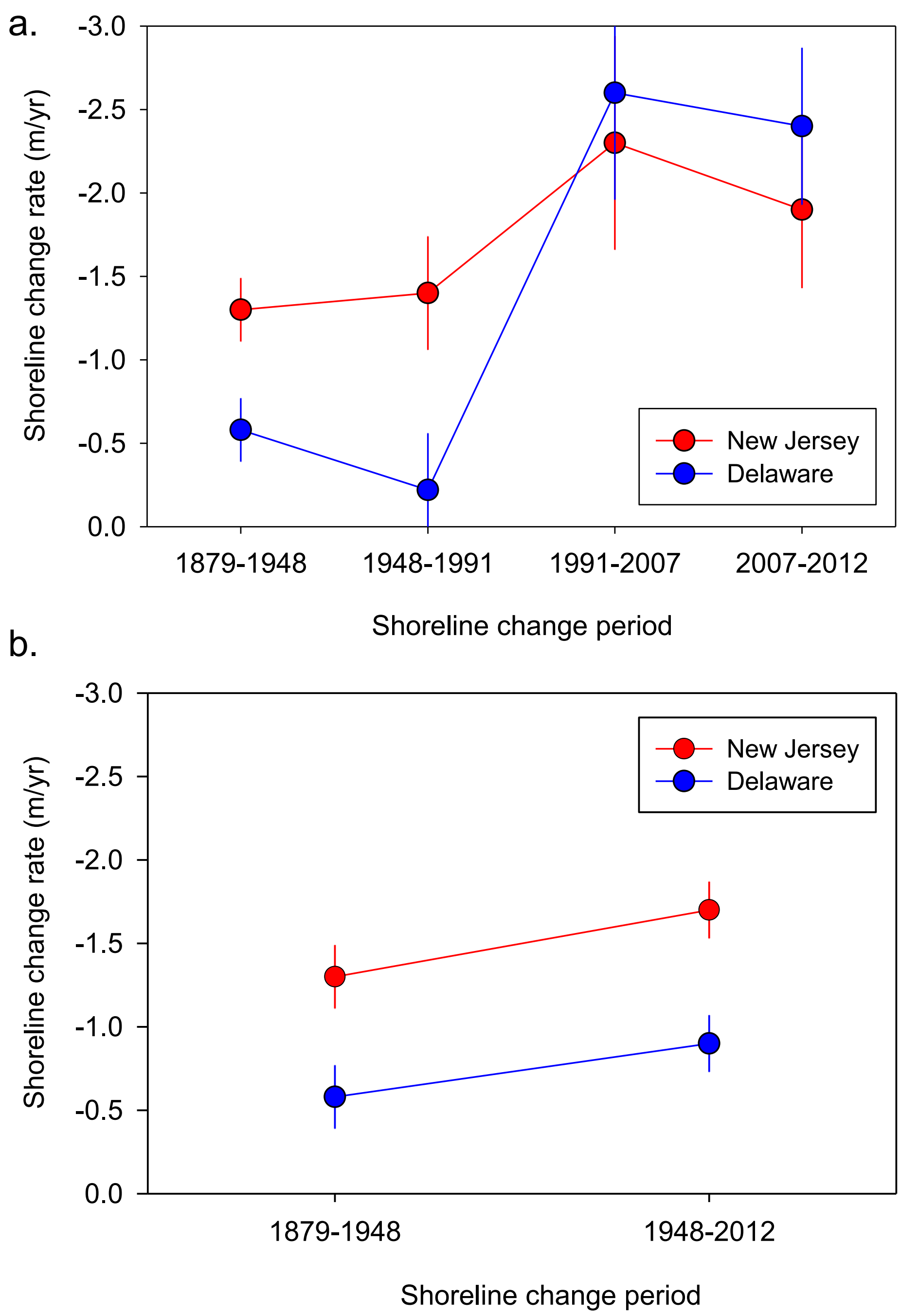


Figure 16. (a) Mean rates of shoreline change (retreat) for each of the EPR averaging intervals listed in Table 4. (b) Rates of change in the lower graph averaged over similar time spans of 69 (1879-1948) and 64 (1948-2012) years.

4.2 Shore Erosion as a Sediment Source

Results of the volumetric change analysis are summarized in Table 8. The total mass of sediment associated with shoreline retreat (erosion) during 1879–2012 was -2×10^{10} kg or -1.5×10^{8} kg/yr when the hardened shoreline segments of Artificial Island and Killcohook are removed from consideration. To put these numbers into perspective, about 1.3×10^{9} kg/yr of suspended sediment are delivered to tidal waters of Delaware Estuary from the Delaware, Schuylkill, and Brandywine-Christina rivers (Mansue and Commings, 1974). Hence, as a potential sediment source, shore erosion yields roughly 12% of the sediment load delivered by rivers, the chief source of new sediment entering the estuary. An estimated 1.2×10^{9} kg/yr of mud accumulates in tidal wetlands of the lower estuary and bay, comparable to the rate of sediment produced by shore erosion in the same region. Although the transport pathways and fate of this eroded sediment are difficult predict—some can be transported to the subtidal estuary whereas some to the tidal wetlands—results of this research make clear that shore erosion is a quantitatively import source of sediment in the estuarine system. Similarly, Wells et al. (2003) determined that shore erosion is an important source of sediment in the Chesapeake Bay region.

Table 8. Results of the shore volumetric change analysis for the Delaware Estuary.

Date Range	Average net movement (m)	Average end point rate (m/yr)	Net volume (m^3)	Net mass (kg)	Net mass (kg/yr)	% Uncertainty
1879-1948[a]	-58.9	-0.95	-2.1×10^{7}	-7.9×10^{8}	-1.1×10^{8}	172
1948-1991	-40.5	-0.89	-3.6×10^{5}	-1.2×10^{8}	-2.9×10^{6}	38
1991-2007	-38	-2.46	-9.5×10^{5}	-3.2×10^{8}	-2.0×10^{7}	25
2007-2012	-10.7	-2.13	-7.9×10^{5}	$- 1.4 \times 10^{9}$	-2.9×10^{8}	15
1879-2012	-96.7	-0.75	-3.6×10^{7}	-1.1×10^{10}	-9.0×10^{6}	8
1879-2012[r,a]	-149.3	-1.09	-5.6×10^{7}	-2.0×10^{10}	-1.5×10^{8}	12

[a] With the Artificial Island and Killcohook shoreline segments removed

[r] Regression based analysis

4.3 Shoreline Change and Sea Level Rise

Local relative sea-level rise, inclusive of eustatic sea level and vertical motion of the land surface, has an overarching influence on the shoreline change and time and space. As such there has been much work among scientists and engineers to predict rates of change based on simple models such as the Brunn Rule (reviewed by Rosati et al., 2013). As a general rule of thumb, for typical shore slopes the Brunn Rule predicts a ratio of shoreline change rate (*r*) to rate of relative sea-level rise of (*s*) of 50 to 100. For the U.S. Mid-Atlantic oceanic coast, values of *r*/*s* vary from about 50 to 120 with an average value of 78 (Zhang et al., 2004). This empirical measure of shoreline change per unit sea-level rise provides a useful index for comparing shoreline change among different coastal segments, and also for predicting near-term changes in shoreline position based on rates of relative sea-level rise.

Using measured rates of relative sea-level rise for Delaware Estuary (see Background section) and the long-term shoreline retreat rates, average values of *r*/*s* range from 167 (Delaware coast) 345 (New Jersey) of the lower estuary. For the bay, *r*/*s* ranges from 190 to 444 for the coasts of Delaware and New Jersey, respectively. Interestingly, for both the wetland-dominated lower estuary and barrier-beach dominated bay, these ratios are 3–4 times higher than those of the adjacent oceanic coasts of Delaware and southern New Jersey (Zhang et al., 2004). Because the rate of relative rate of sea-level rise does not vary much between Delaware Estuary and the oceanic coast, differences in *r*/*s* are mostly a reflection of differences in coastal topography, hydrodynamics, substrate erodibility, and sediment supply and transport.

4.4 Influence of Waves on Shoreline Change

To gain insight into relationships between wave properties and shoreline change in Delaware Estuary, results of wave power modeling by Chen et al. (2016) for Delaware Estuary were compared to patterns and rates of shoreline change observed in this study. Using the Simulating WAves Nearshore (SWAN) model and a high-resolution bathymetric grid for the lower estuary–bay, Chen et al. (2016) computed wave power (Equation 1) at the shoreline for the 12 ordinal wind directions. The model was run using local wind climatological data from 2014 measured at the NOAA Ship John Shoal Station, and the output was presented geographically with wave power (in kW/m) plotted along the shoreline. Plots for the 12 model runs are presented individually in the Appendix A.

In general, areas of high wave power correspond to areas of rapid shoreline retreat (Figure 17). This is consistent with the notion that wave power and shoreline retreat in estuaries follows a linear relationship (Schwimmer, 2001; McLoughlin et al., 2015; Leonardi et al., 2016). There are some areas where there are high rates of erosion and low wave power and vice versa. For example, on the New Jersey side of the Delaware Bay extending from the mouth of Dennis Creek westward to Egg Island, high rates of shoreline retreat are present when modeled values of wave power are relatively low. One possible explanation is that swell from the Atlantic Ocean contributes to wave height and power locally. Because the SWAN modeling did not include remotely forced waves (oceanic swell), only locally driven wind waves, modeled wave power by Chen et al. (2016) could be somewhat low along swell-impacted shores. Swell propagation modeling by Jenkins et al. (in review) indicates that that swell entering Delaware Bay refracts eastward and toward the shore of the embayment between Cape May and Little Egg Island. Hence, it is possible that oceanic swell contribute wave power relevant to shoreface erosion and shoreline retreat on the New Jersey side of the lower bay.

The model results of wave power from SWAN are comparable to the results of wave power calculated from unweighted wind statistics from 2014. Mean wave power at each of the buoy locations range from 0.11 to 0.40 kW/m and fall within range of wave power modeled by Chen et al. (2016) (Table 7). In summary, wave power modeling by Chen et al. (2016) can explain the patterns and rates of shore retreat where retreat rates are highest, but additional work will be required to establish a connection between wave characteristics and shoreline change for Delaware Estuary as a whole.

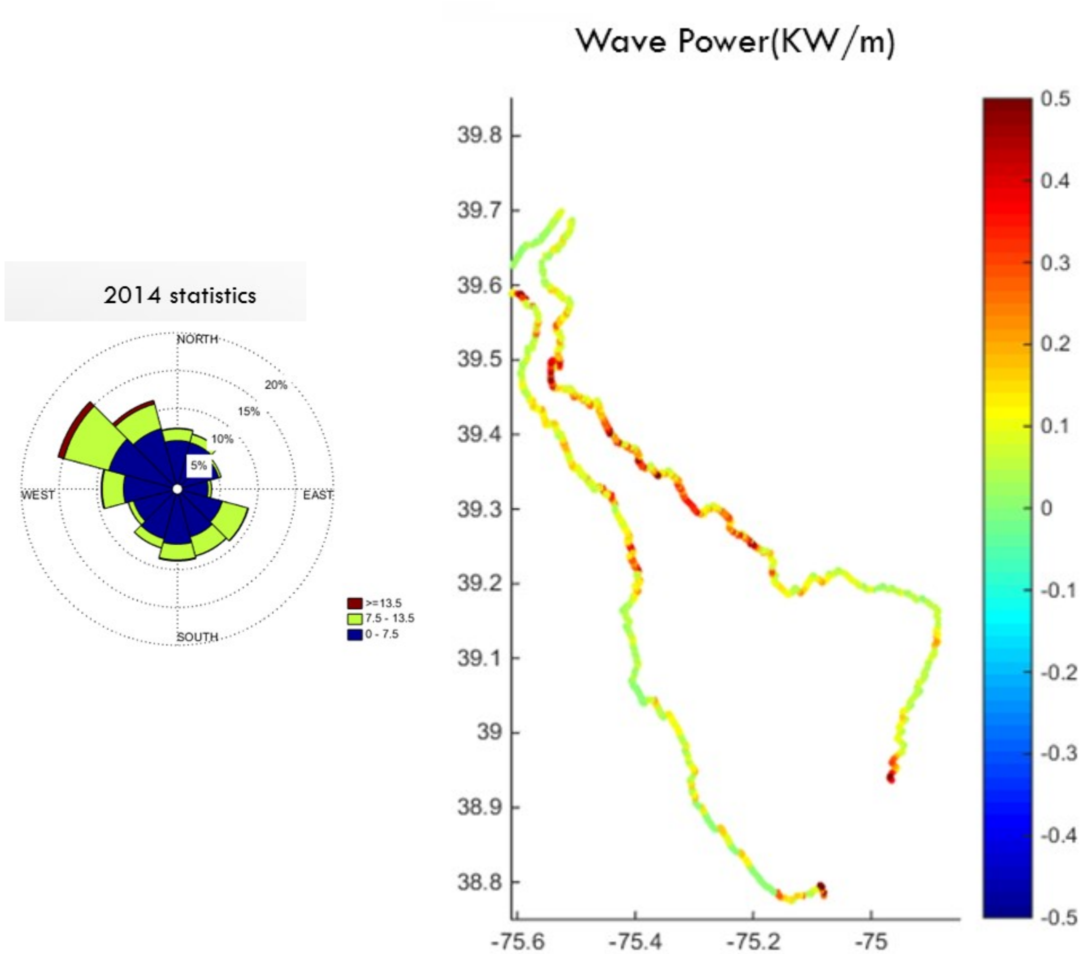


Figure 17. (Left) Wind speed and direction data for 2014 used to drive the SWAN model by Chen et al. (2016). (Right) Composite wave power map for the 12 model runs. Note that locations of high wave power generally fall in area of high shoreline retreat shown in Figures 8 and 9

Chapter 5

CONCLUSIONS

The aim of this thesis research was to quantify patterns and rates of historical shoreline change in Delaware Estuary, in an effort to fill gaps in our understanding of change in response to natural processes and anthropogenic factors. Specific objectives included characterizing shoreline types throughout the lower estuary and bay, computing rates of shoreline change using synoptic shoreline datasets, and relating measured wave properties to observed rates of shoreline change. The major findings of this study are summarized below:

1. The Delaware Estuary shoreline as a whole has been retreating at an average long-term rate of -1.1 ±0.13 m/yr (1879–2012). The New Jersey shore of the lower estuary–bay is retreating (-1.5 ± 0.13 m/yr) about two times more rapidly than the Delaware shore (-0.68 ±0.13 m/yr). These rates of retreat are nearly two times higher than rates reported for the Atlantic coast of Delaware (Hapke et al., 2013).
2. Short-term rates of shoreline retreat determined by end-point analysis (2007–2012) were two times higher than average long-term rates determined by linear regression (1879–2012) and were more variable spatially and temporally. Comparison end-point rates of shoreline change for the entire lower estuary–bay suggests that there was a statistically significant increase in retreat rate from 1879–1948 (69 years) to 1948–2012 (64 years). Identifying the possible causes of this change in relation to relative sea-level rise, storm

magnitude and frequency, and estuarine processes and geomorphology is warrants further investigation.

3. Widespread shoreline retreat by shoreface erosion in the estuary calls into question the notion that the estuary is in a "constructive phase" of evolution (e.g., Fletcher et al., 1992). Accordingly, the standing conceptual model for the transgressive evolution of the estuary should be revisited and revised to include results of the present study.

4. Wetland and transitional coasts of the lower estuary−bay have been retreating faster (-1.2 and -1.3 m/yr) than barrier beach coasts (-0.67 m/yr). This difference may be related to sedimentary processes that allow the profile of sandy beaches to recover from wave erosion more rapidly than vegetated coasts.

5. A direct, qualitative relationship between modeled (Chen et al., 2016) and measured wave power and rates of shoreline retreat was observed in this study. This finding suggests that patterns and rates of shoreline retreat in the estuary and be explained partly in terms of dynamical processes that influence shoreface erosion. Further research is needed to detail the nature of these processes, as well as sedimentological and ecological controls on substrate erodibility.

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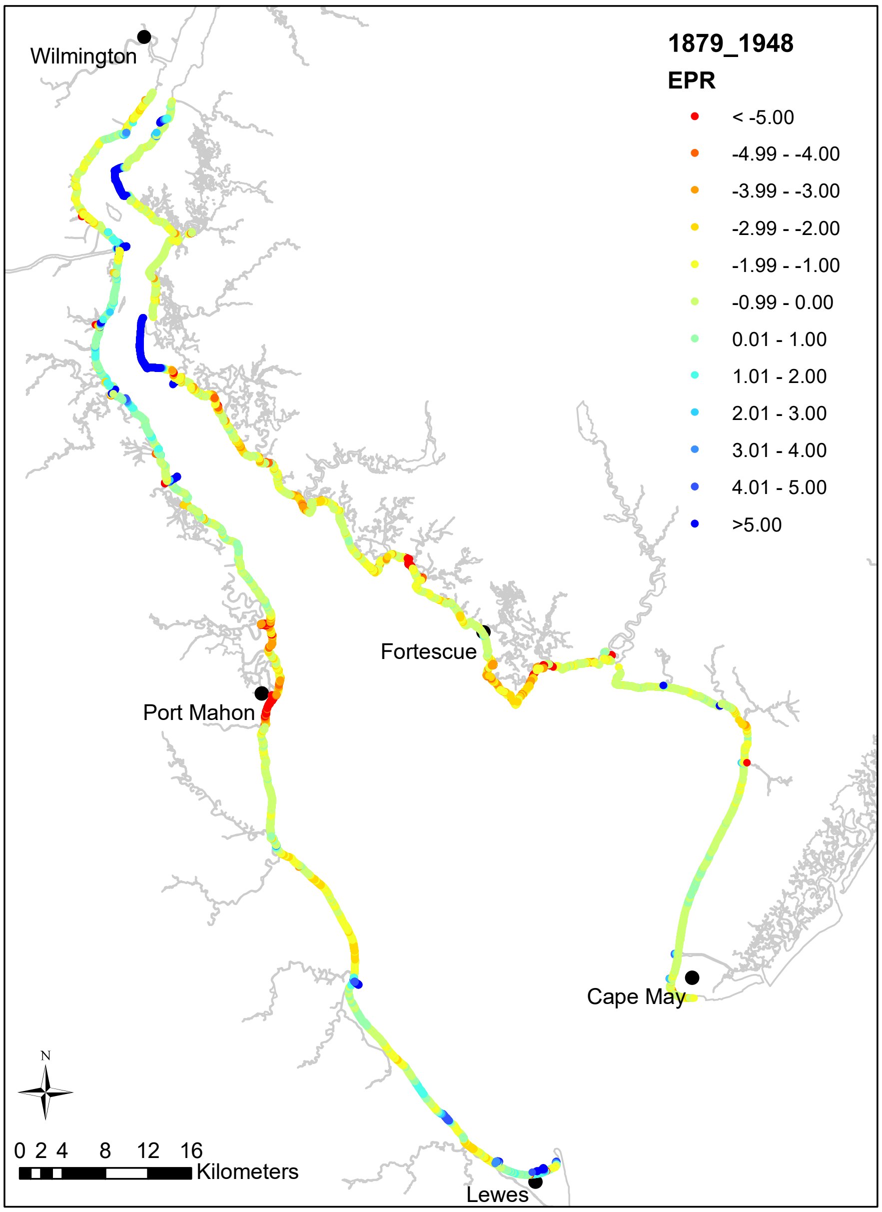
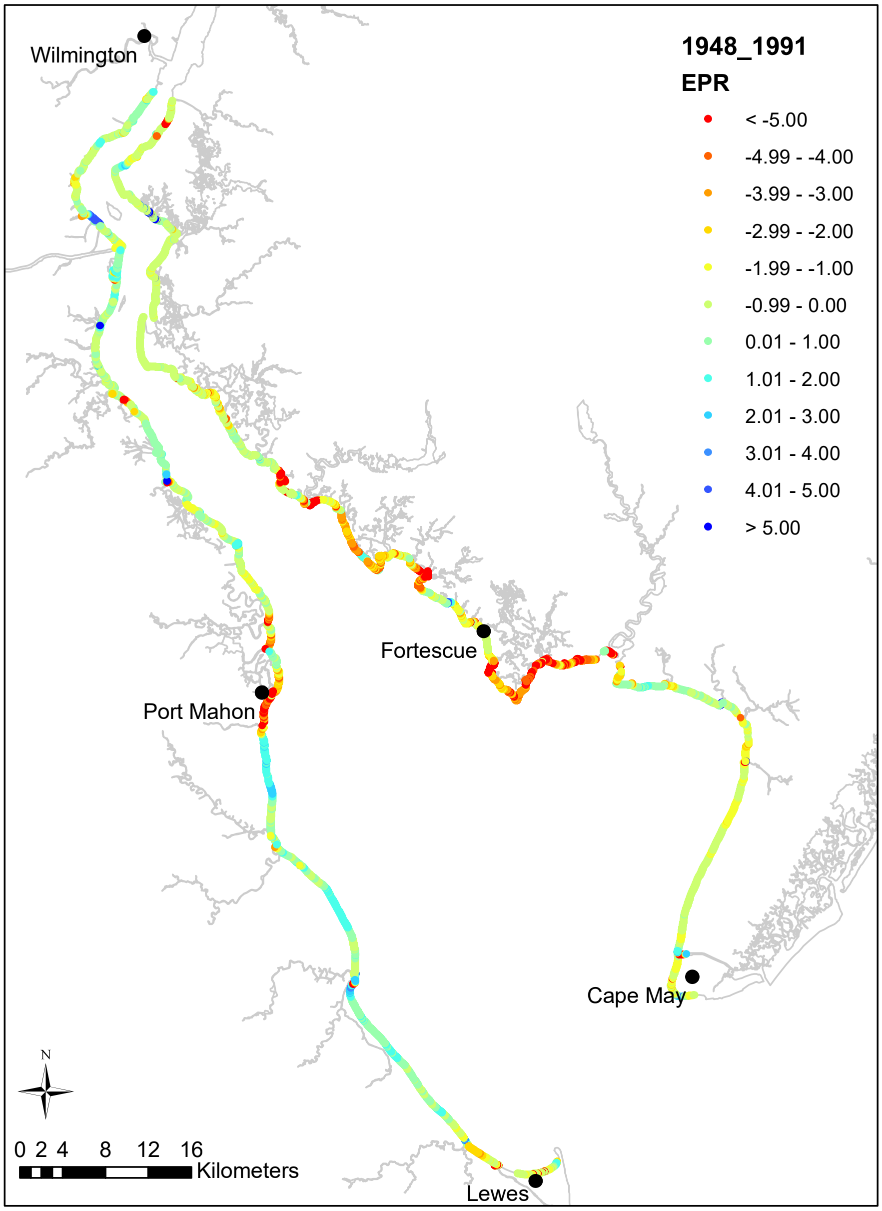
Appendix A

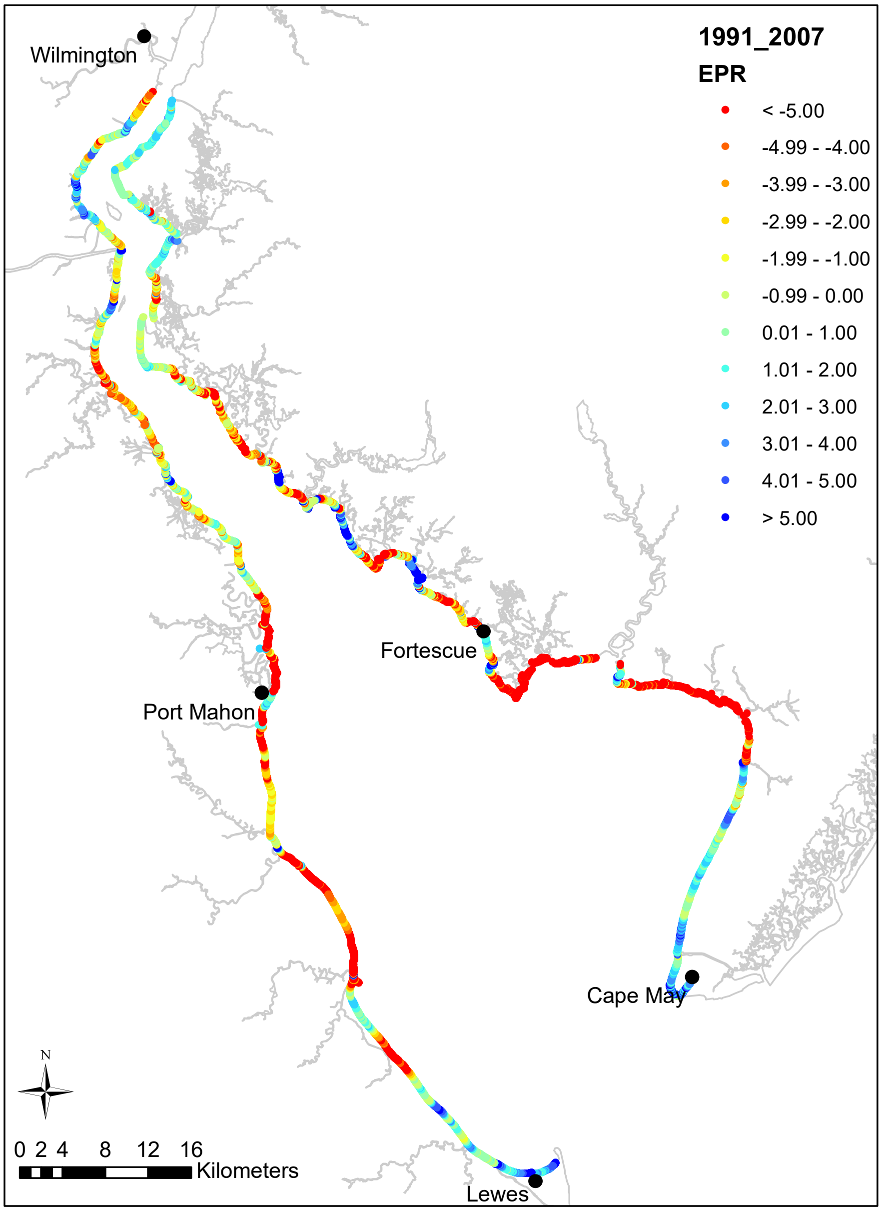
SHORELINE CHANGE AND WAVE DATA



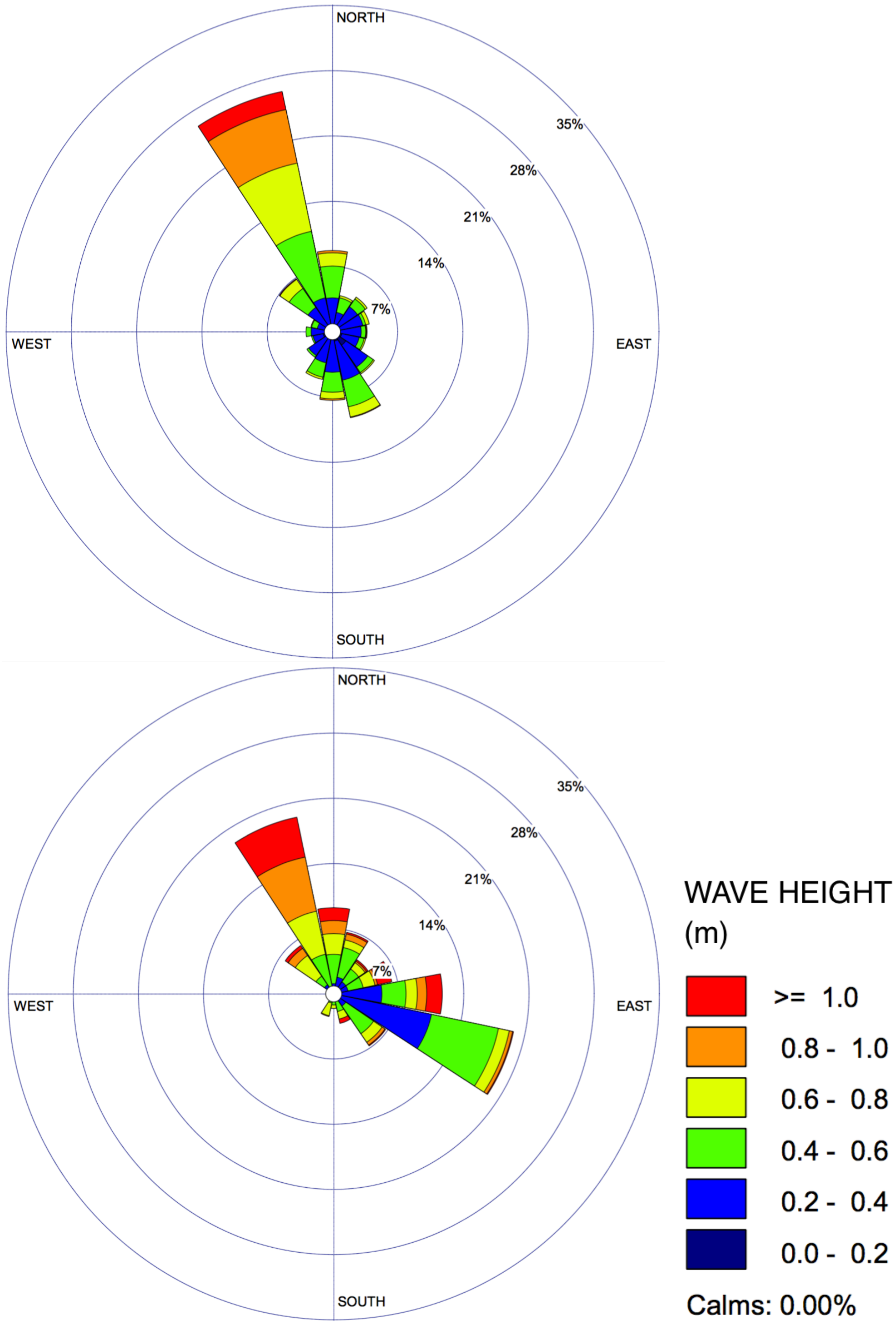
A.1. EPR Shoreline Change between 1879 and 1948.



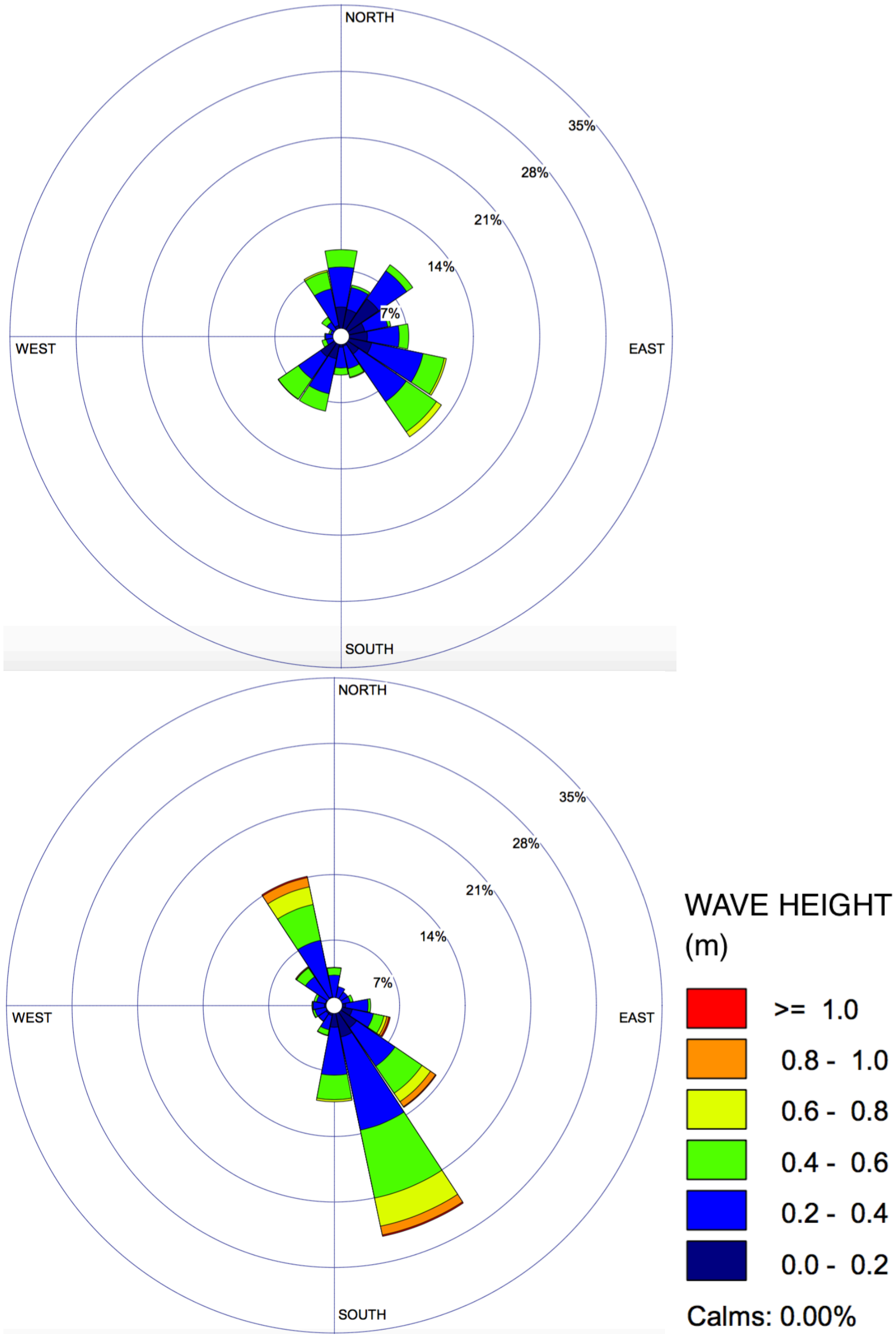
A.2. EPR shoreline change between 1948 and 1992.



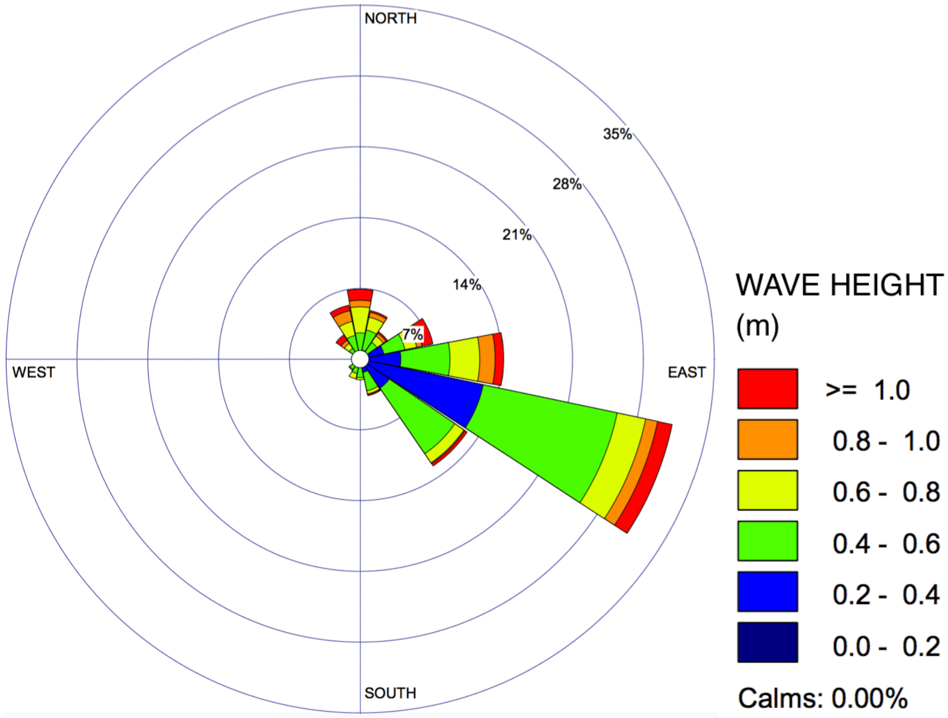
A.3. EPR shoreline change between 1992 and 2007.



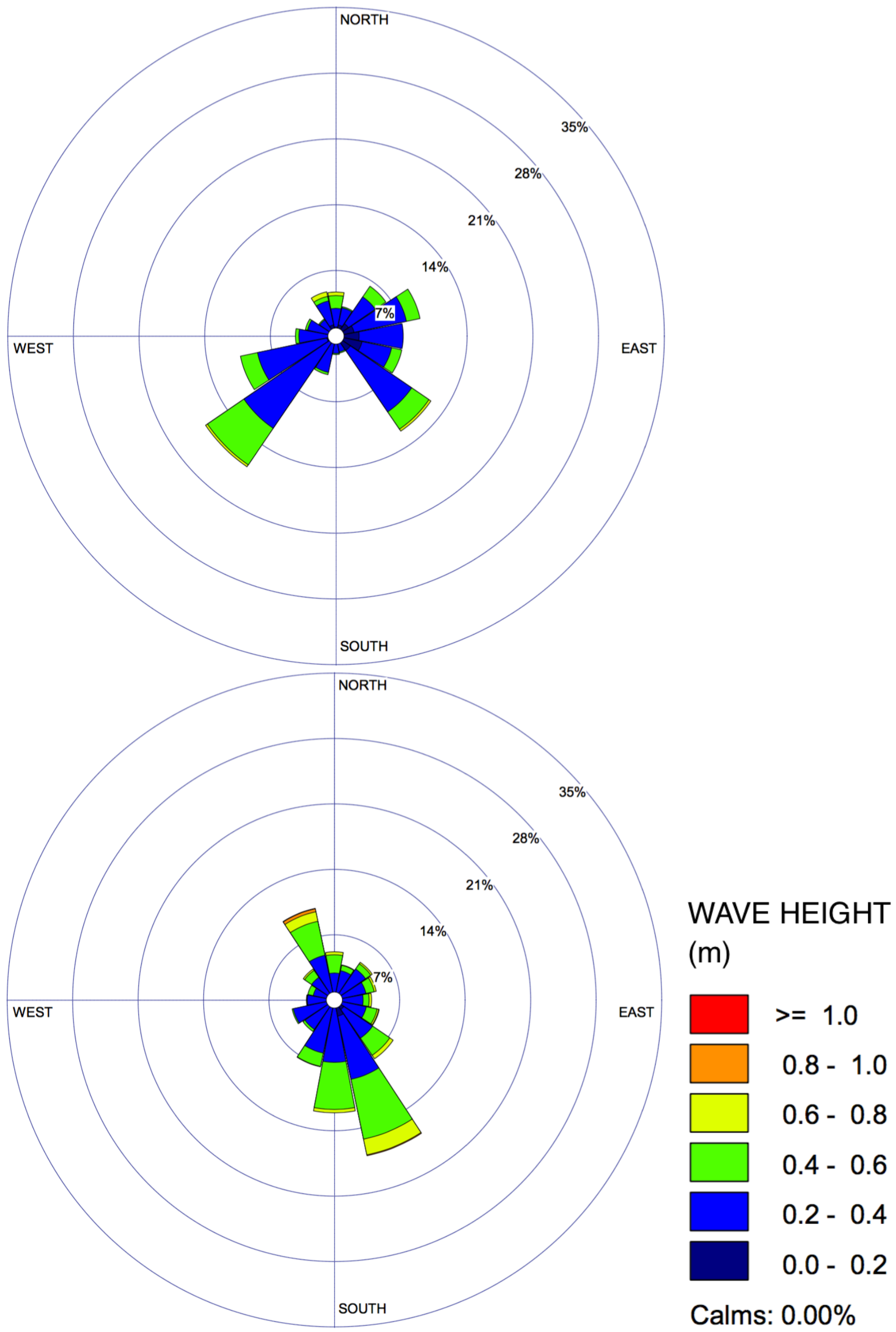
A.4. Winter wind roses for Buoy 2 (top) and NDBC 44055 (bottom).



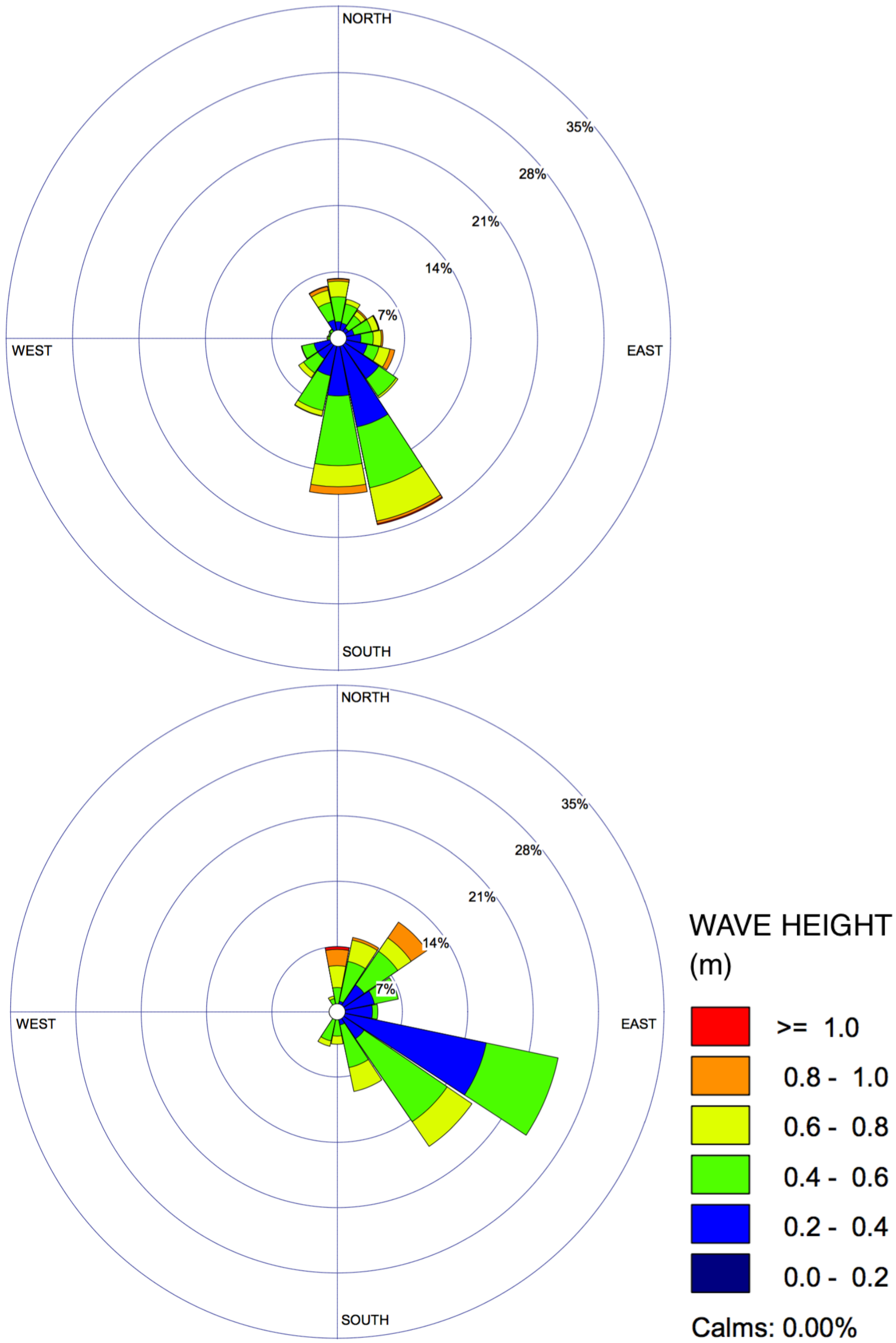
A.5. Wave roses of spring data from Buoy 1 (top) and Buoy 2, (bottom).



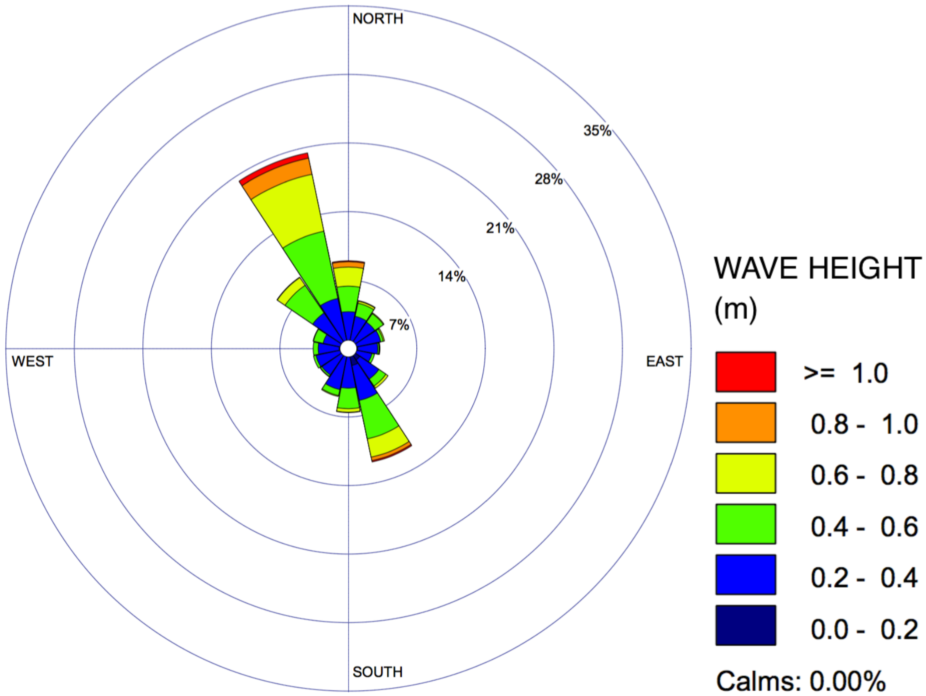
A.6. Wave rose of spring data from NDBC4054.



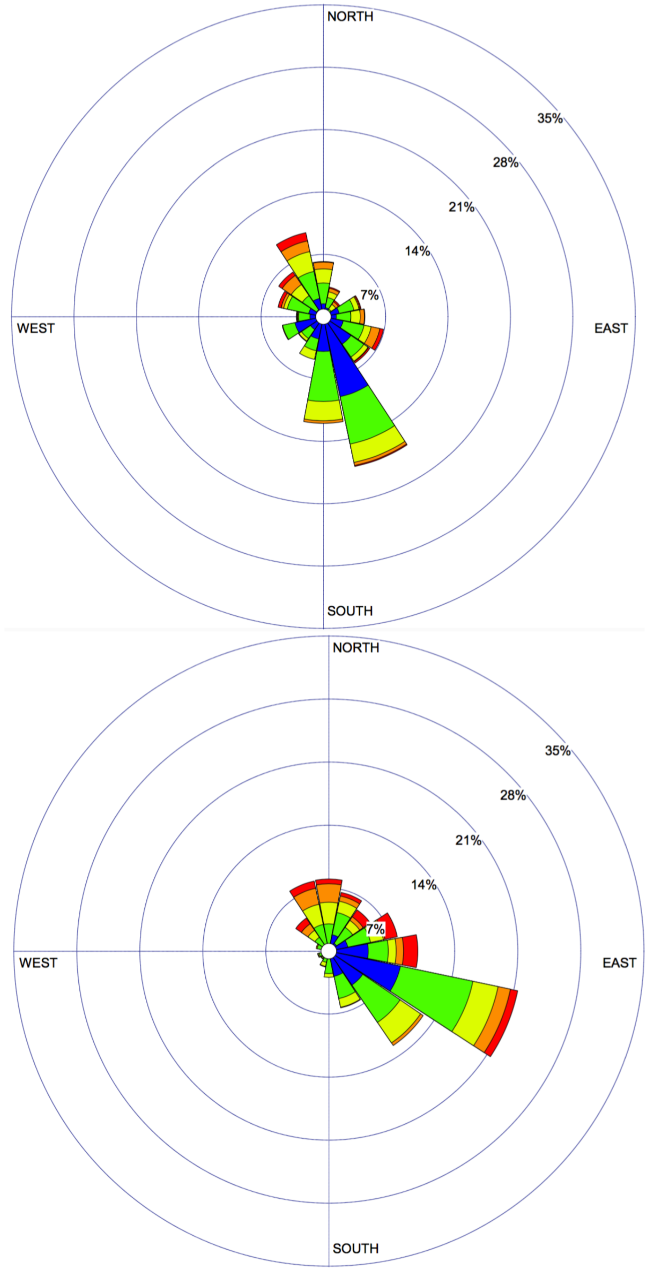
A.7. Summer season wave roses for Buoy 1 (top) and Buoy 2 (bottom).



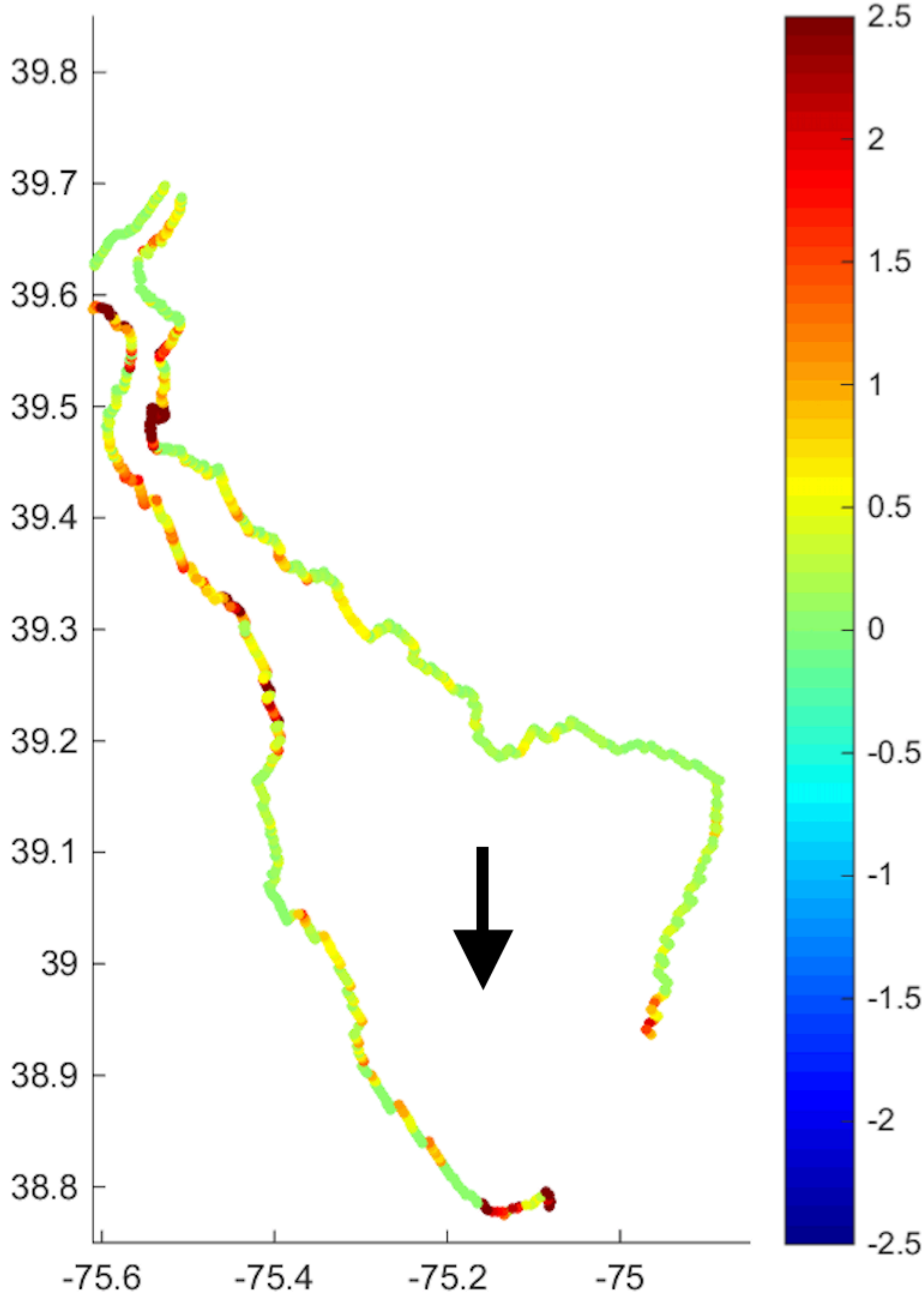
A.8. Summer wave roses for NDBC44055 (top) and NDBC44054 (bottom).



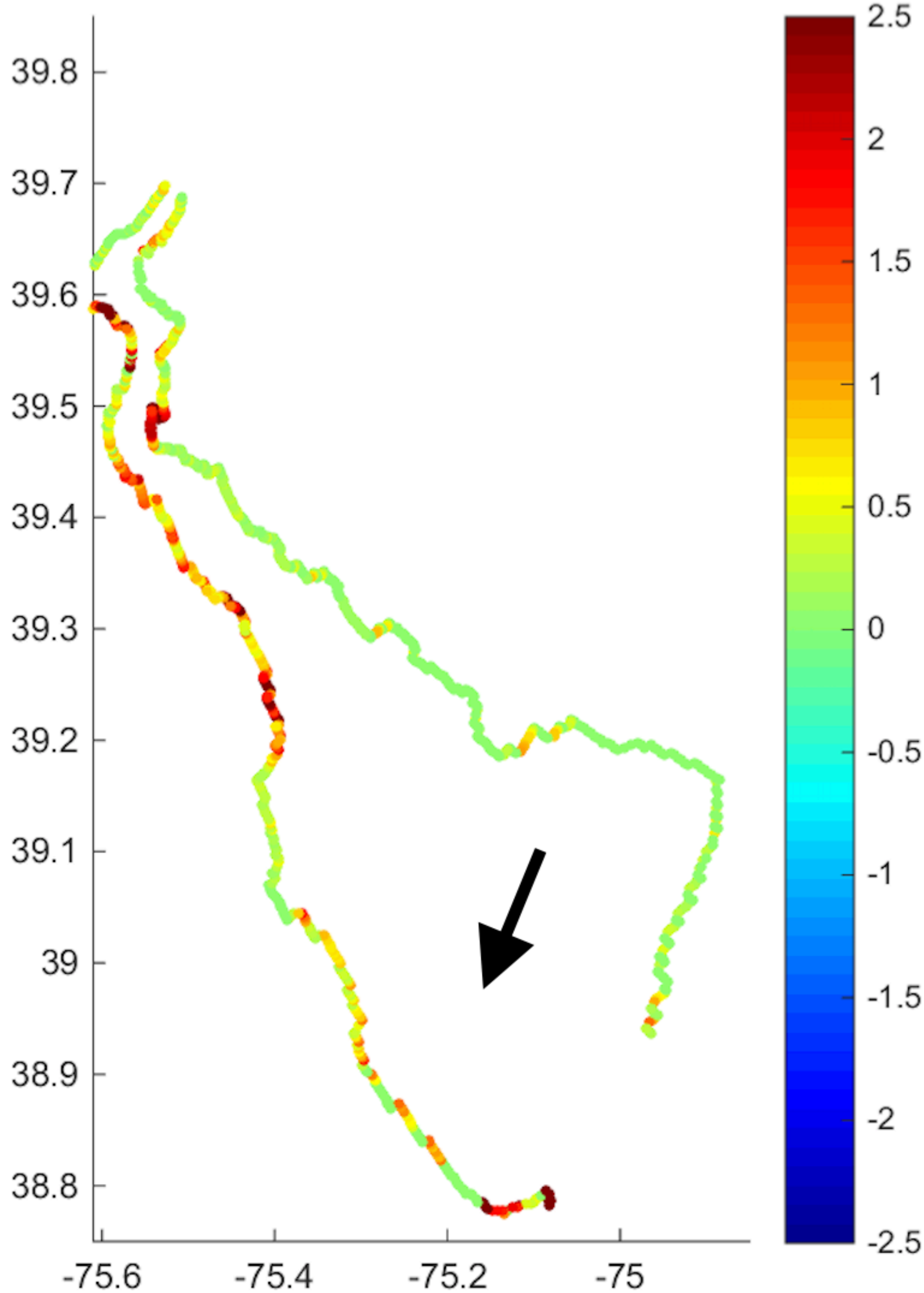
A.9. Fall wave height measurements of Buoy 2.



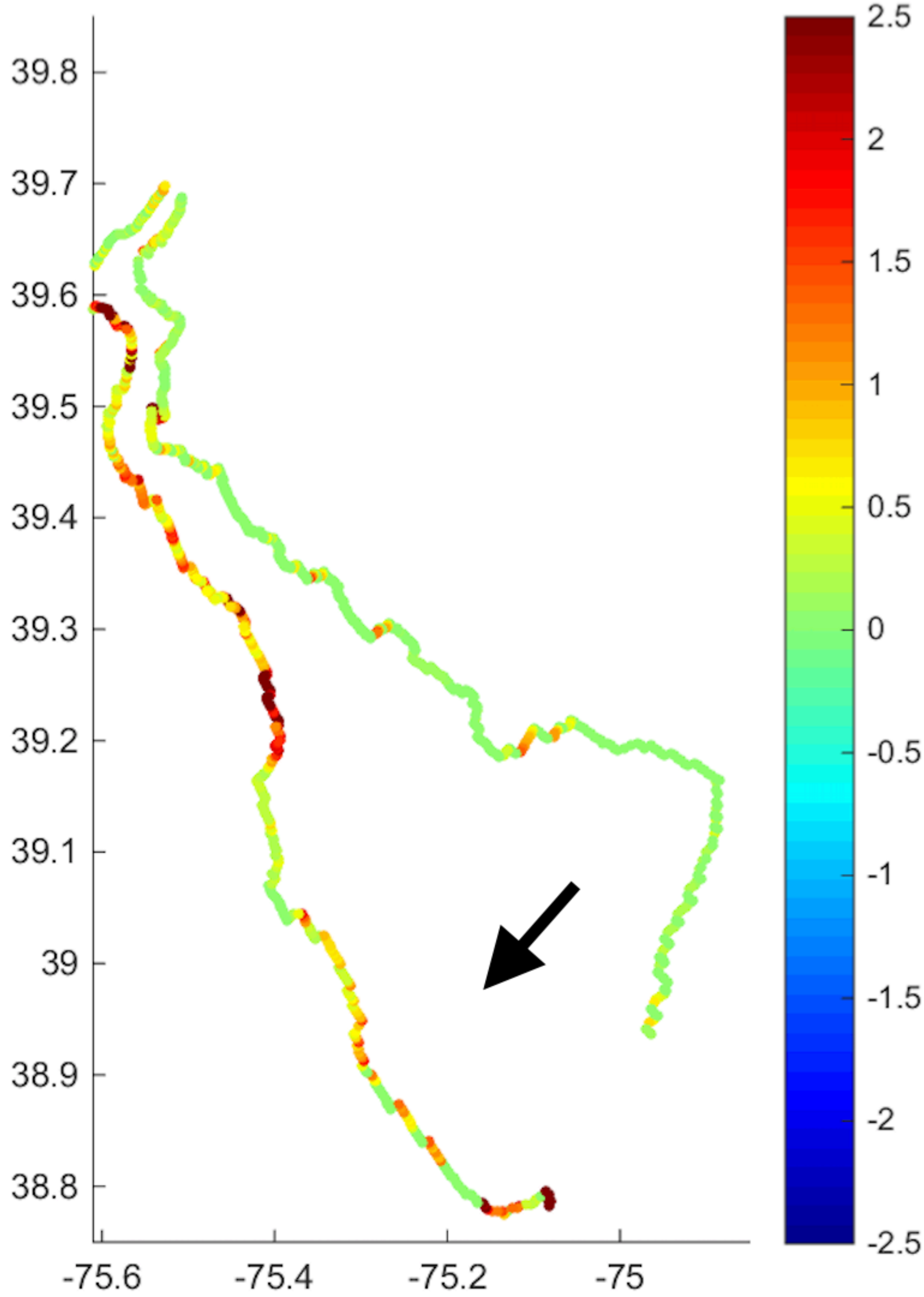
A.10. Fall wave data for NDBC44055 (top), and NDBC44054 (bottom).



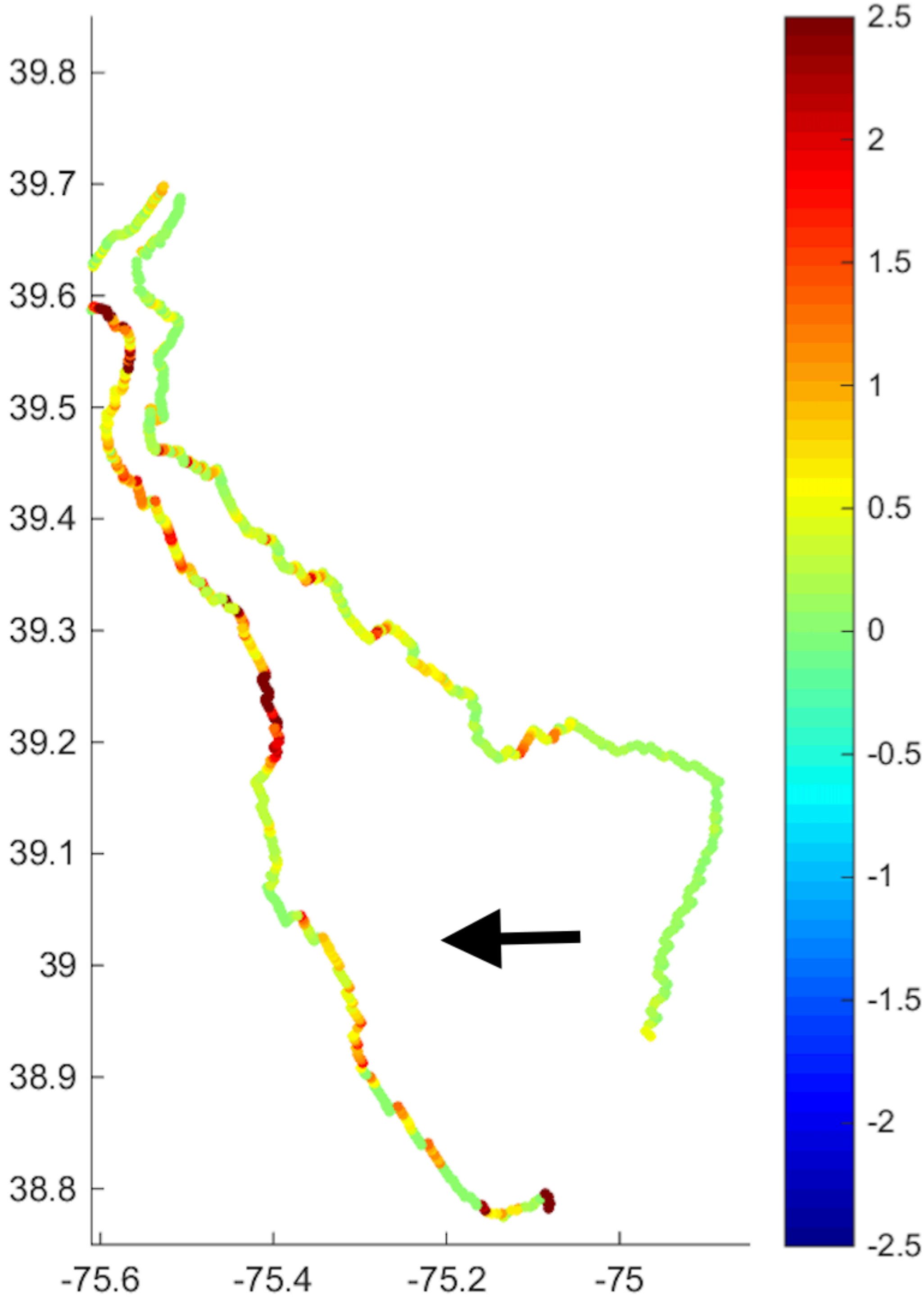
A.11. Wave power with winds from 0 degrees by Chen et al. (2016).



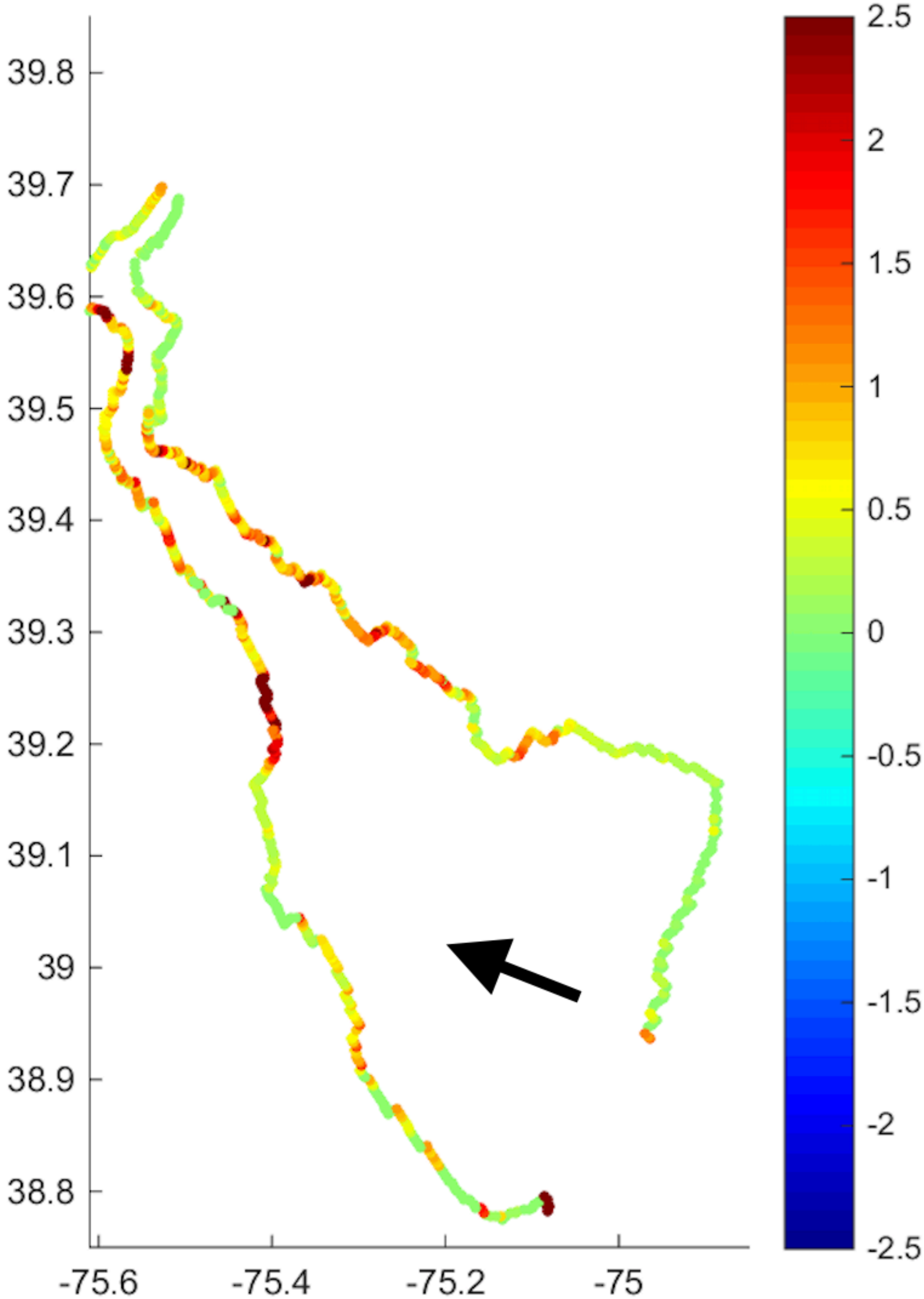
A.12. Wave power with winds from 30 degrees by Chen et al. (2016).



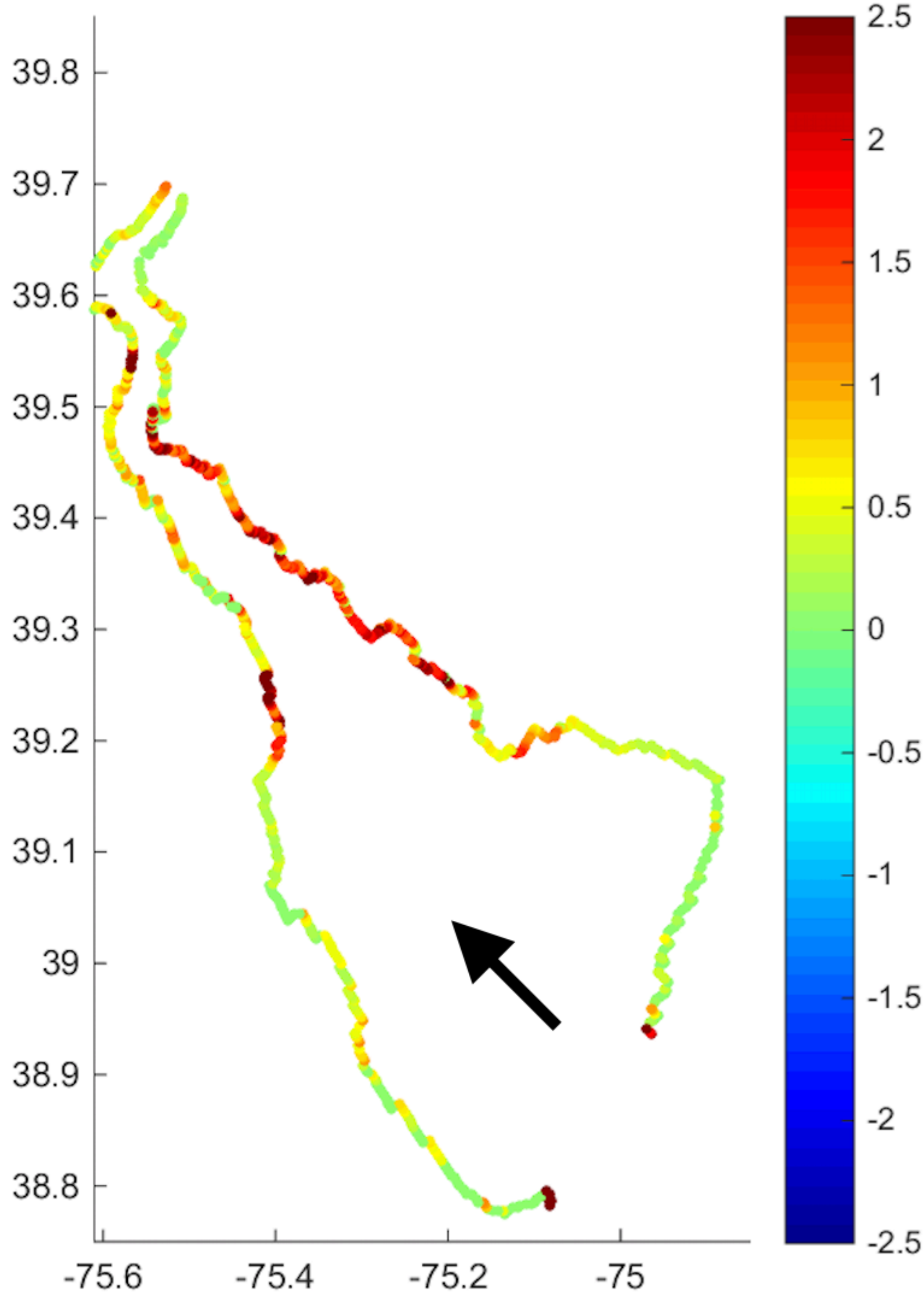
A.13. Wave power with winds from 60 degrees by Chen et al. (2016).



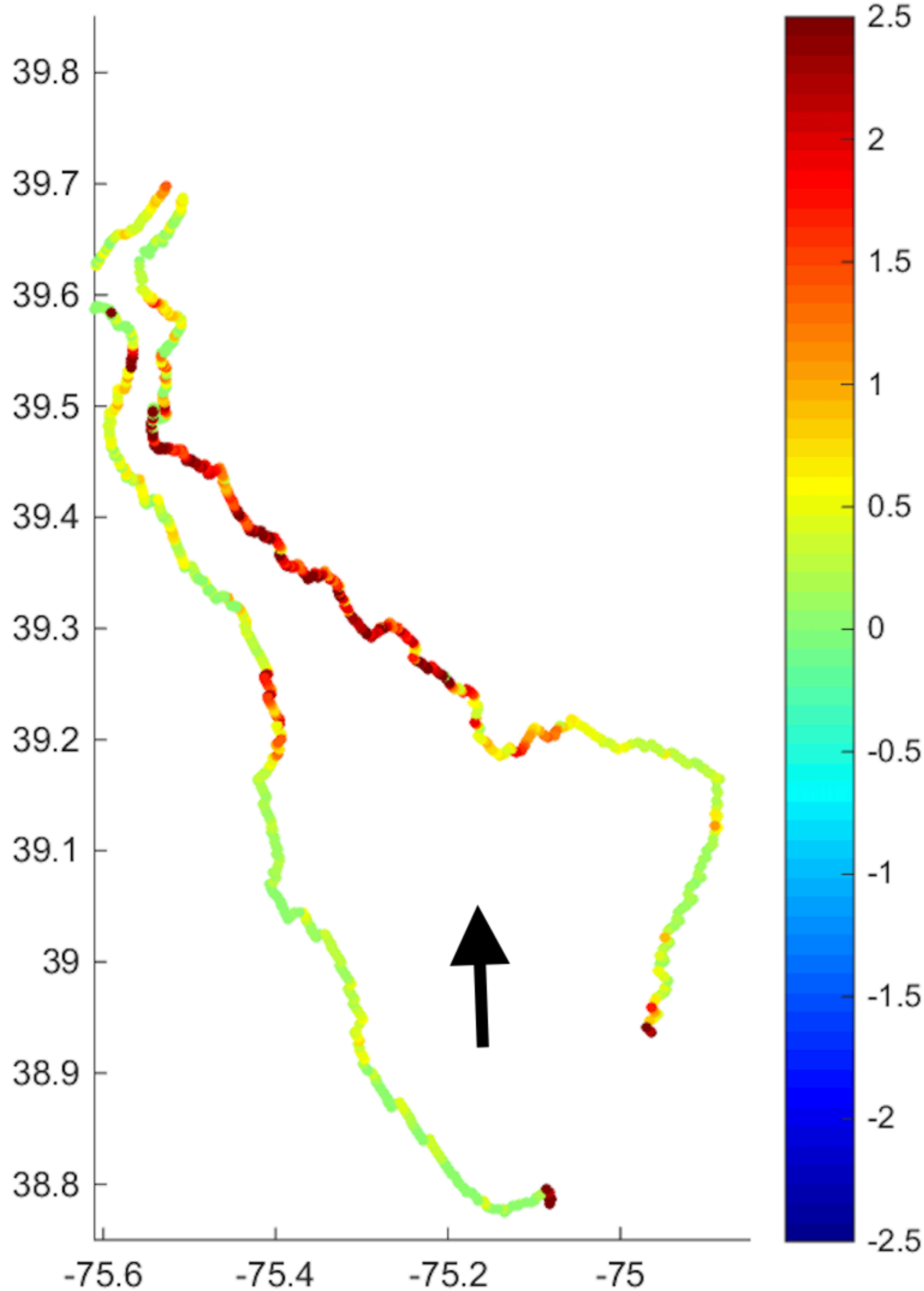
A.14. Wave power with winds from 90 degrees by Chen et al. (2016).



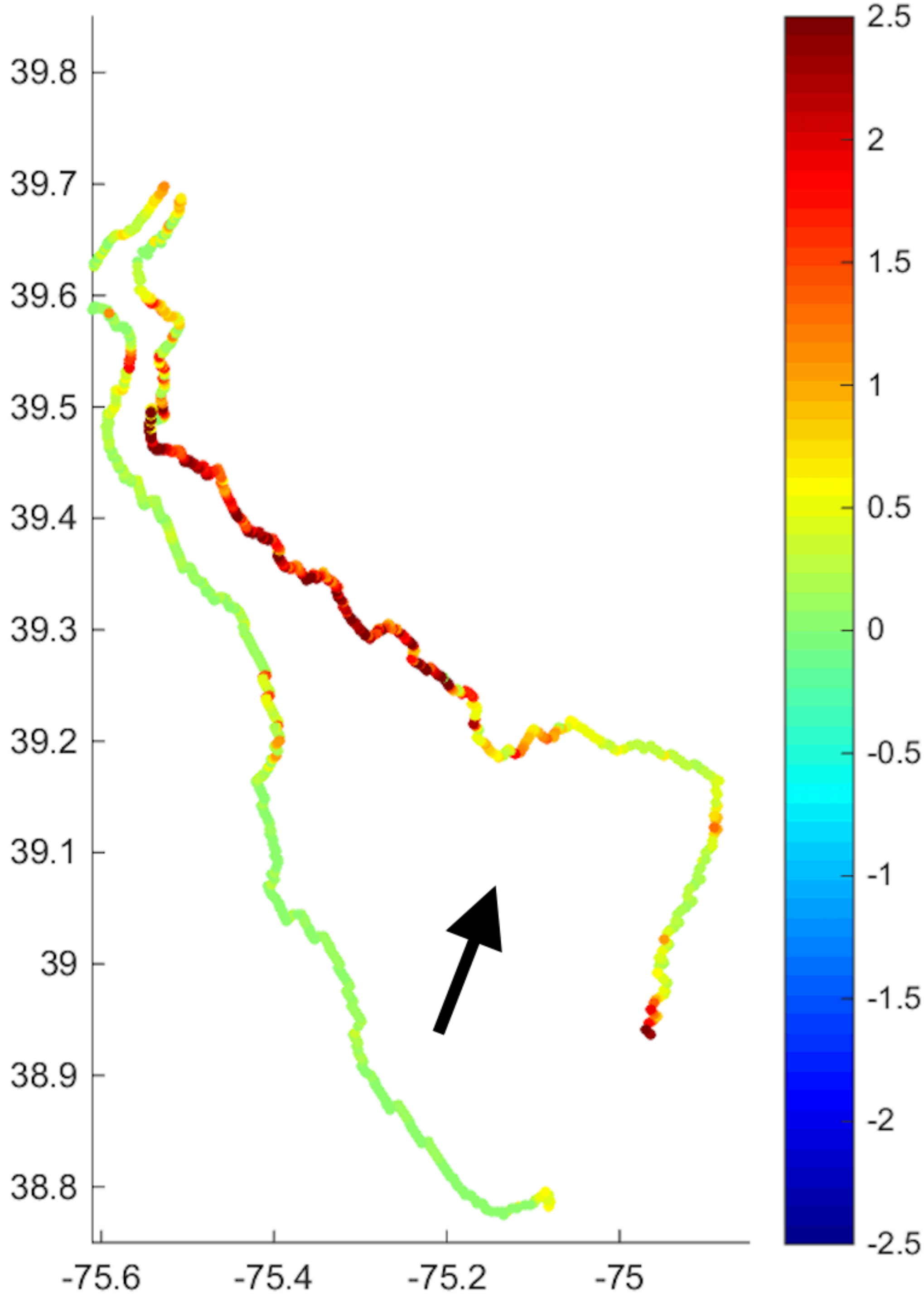
A.15. Wave power with winds from 120 degrees by Chen et al. (2016).



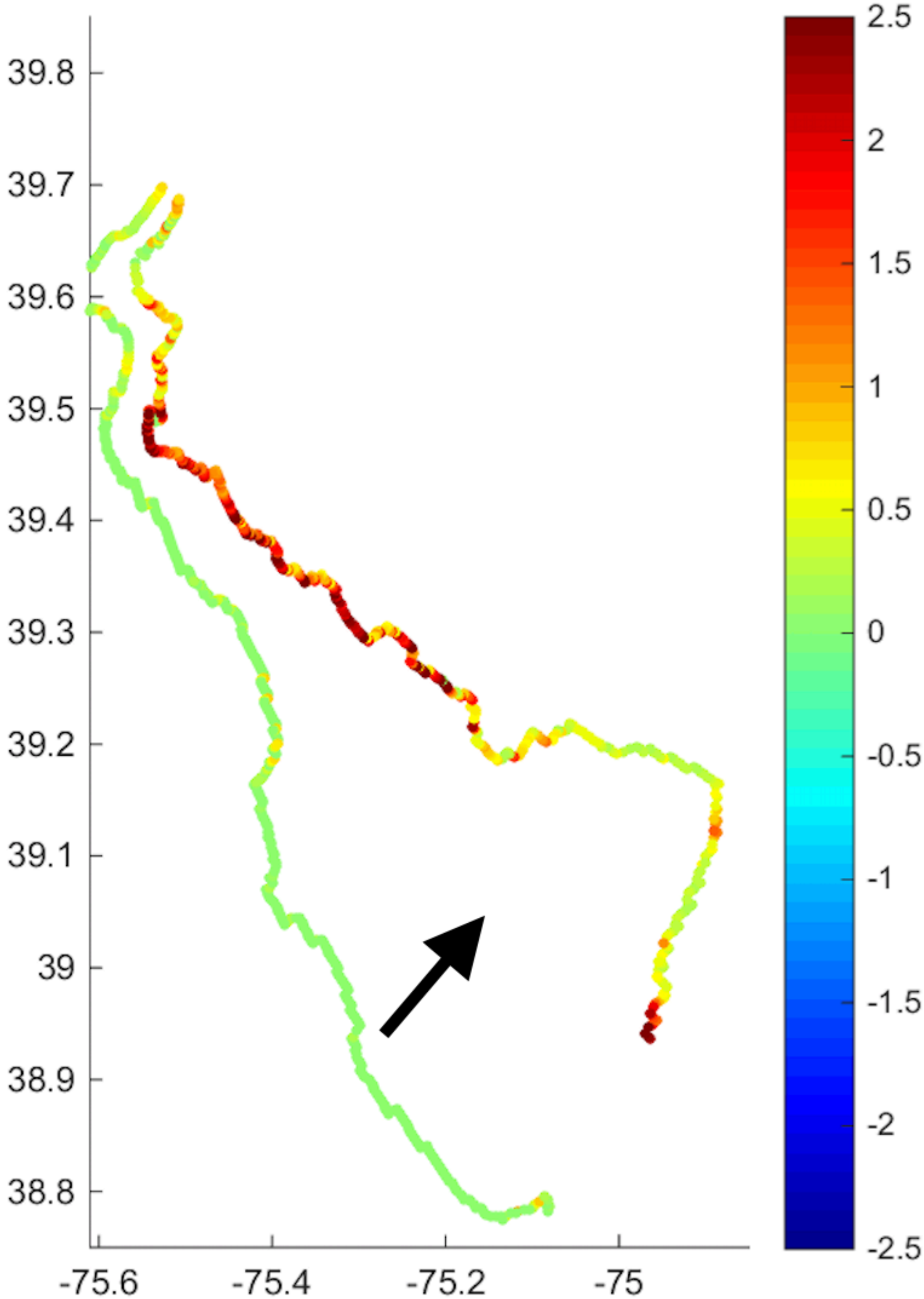
A.16. Wave power with winds from 150 degrees by Chen et al. (2016).



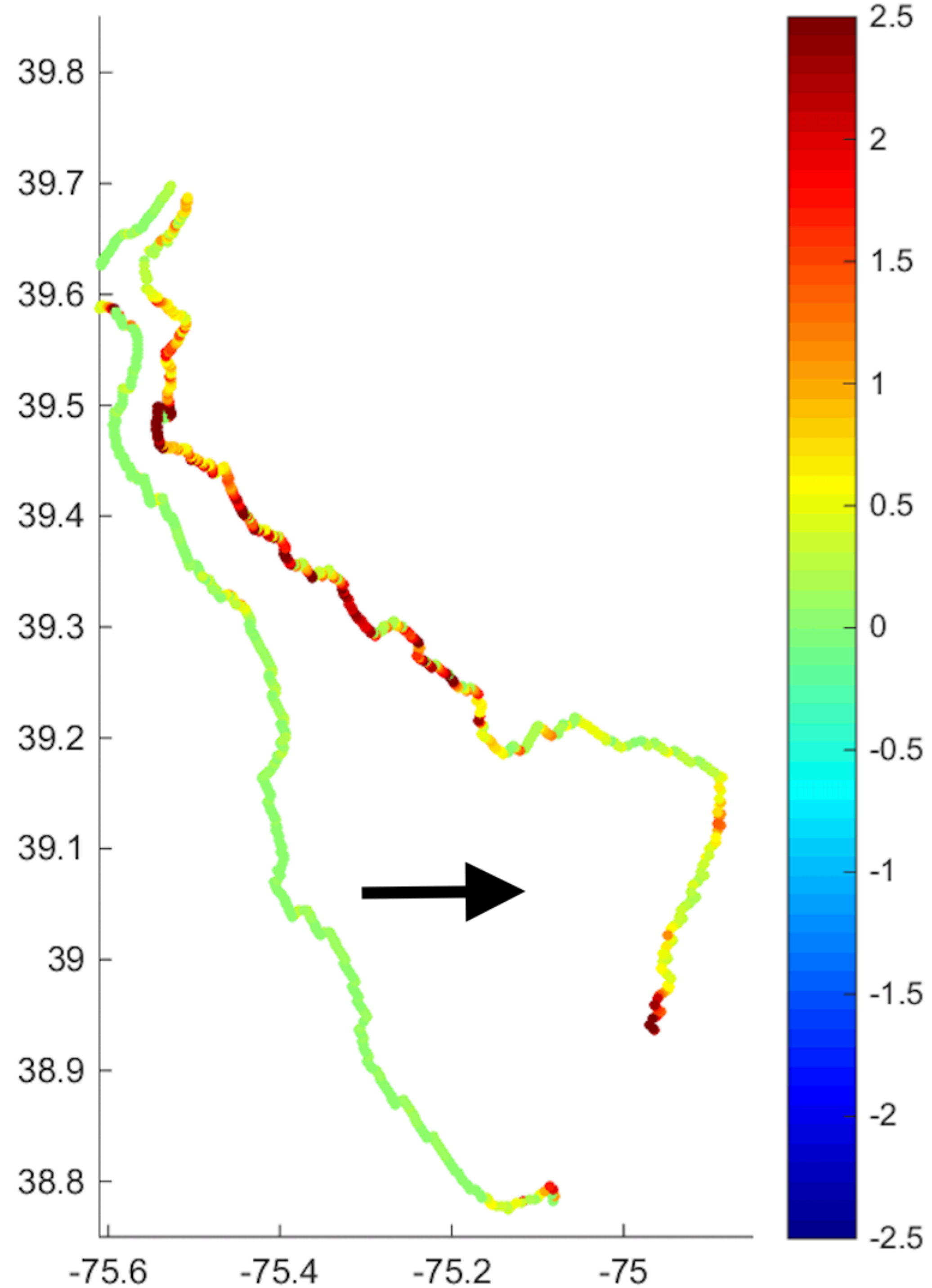
A.17. Wave power with winds from 180 degrees by Chen et al. (2016).



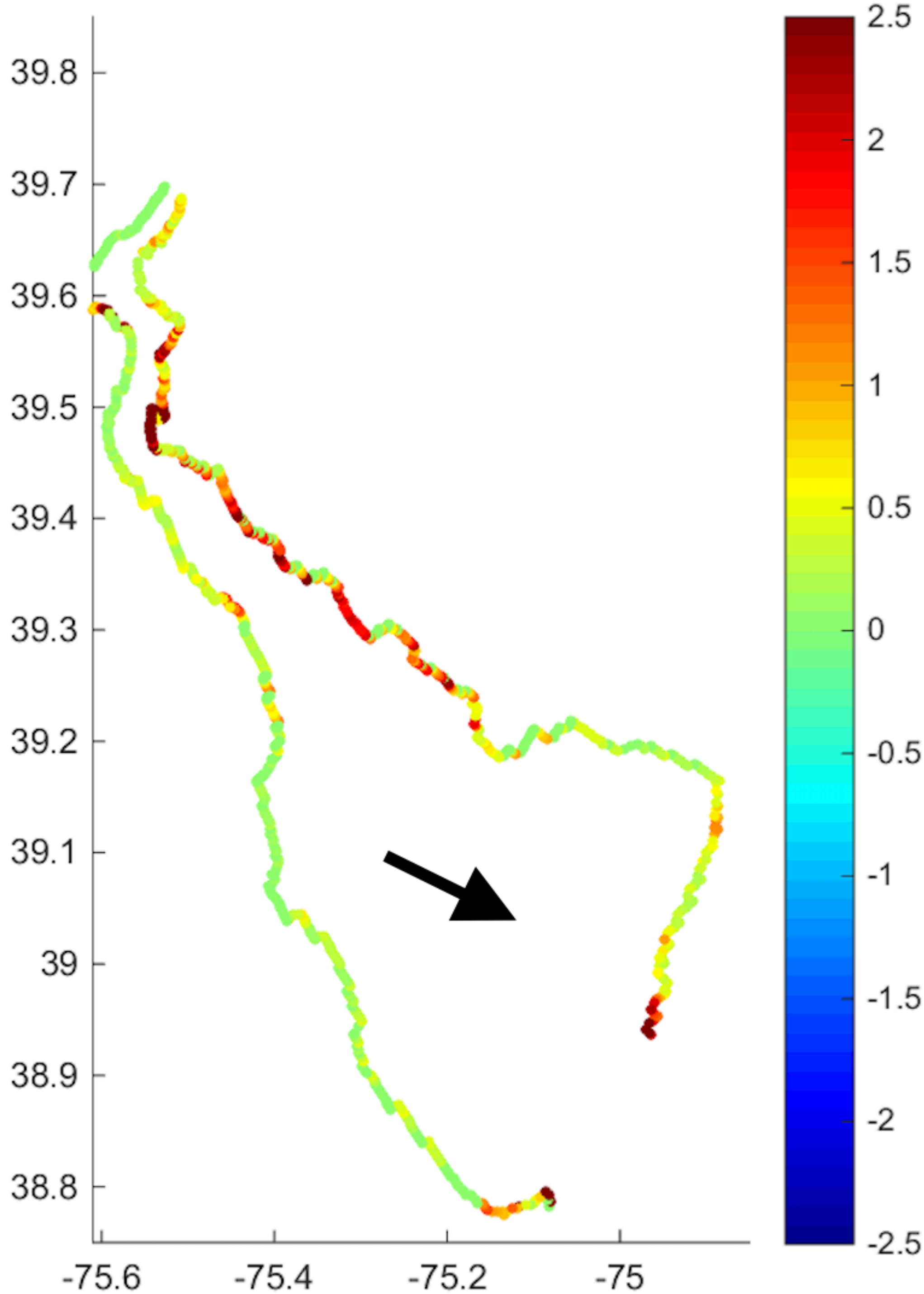
A.18. Wave power with winds from 210 degrees by Chen et al. (2016).



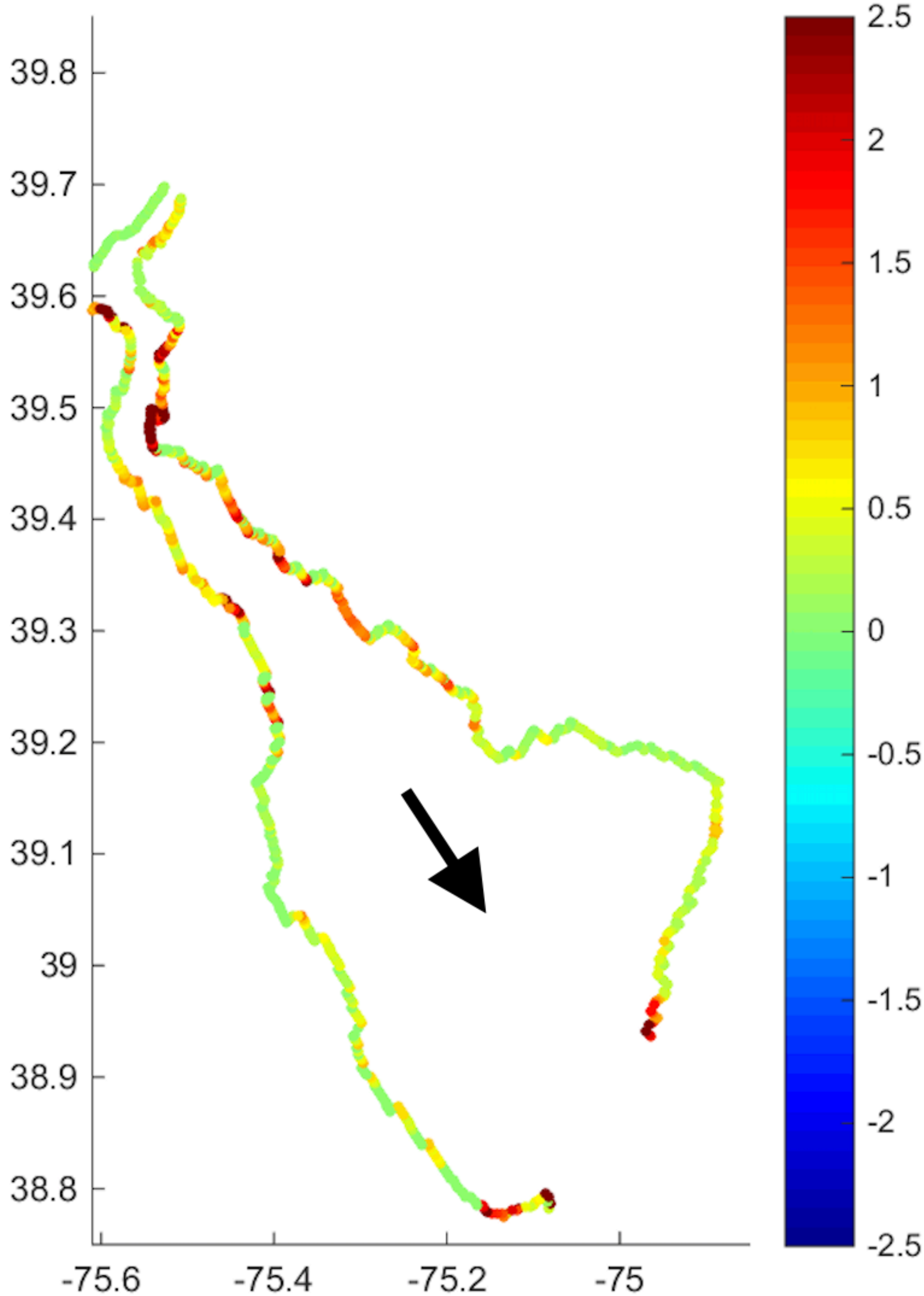
A.19. Wave power with winds from 240 degrees by Chen et al. (2016).



A.20. Wave power with winds from 270 degrees by Chen et al. (2016).



A.21. Wave power with winds from 300 degrees by Chen et al. (2016).



A.22. Wave power with winds from 330 degrees by Chen et al. (2016).

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