Modeling large-scale shoreline change caused by complex bathymetry in low-angle wave climates

Patrick W. Limber^{1,†}, Peter N. Adams¹, and A. Brad Murray²

¹Department of Geological Sciences, University of Florida, 241 Williamson Hall, Gainesville, FL, USA 32611

²Division of Earth and Ocean Sciences, Nicholas School of the Environment, Duke University, Durham, NC, USA 27708

[†]*Corresponding Author, now at U.S. Geological Survey, Pacific Coastal and Marine Science Center, Santa Cruz, CA, USA 95060, plimber@usgs.gov*

KEYWORDS: Sediment transport; coastal geomorphology; shoreline change; wave transformation

Abstract. Coastlines where waves consistently approach at highly oblique angles experience anti-diffusional behavior, causing perturbations to grow seaward and form sand waves, capes, and spits. Coasts where waves approach at low offshore angles experience the opposite: perturbations diffuse and the coastline remains (or becomes) smooth. In this paper, by coupling a 2-D large-scale coastline evolution model to a spectral wave model, we show that anti-diffusional behavior is also possible in low-angle wave climates if the nearshore wave field is altered by complex bathymetry. In model simulations, low-angle waves refract over local shoals, creating a convergence in alongshore sediment flux behind the shoals that coalesces into small

‘minor capes’. Depending on wave height and period, shoreline features take 80-400 years to reach an equilibrium cross-shore relief of 1 – 1.5 km over an alongshore distance of ~20 km. The modeled equilibrium time scale is consistent with analytically-determined characteristic shoreline diffusion time scales, and the modeled cross-shore relief and aspect ratio are similar to observed ‘minor capes’ along the U.S. Atlantic Coast where offshore bathymetric anomalies have previously been linked to shoreline change patterns. Better understanding of the links among nearshore bathymetry, wave transformation, and alongshore sediment transport is critical to understanding shoreline change patterns at the local level and how they fit into broader, regional-scale behavior.

Introduction

On sedimentary coasts, large-scale (>1 km), long-term (>10¹ years) shoreline change is caused in large part by gradients in wave-driven alongshore sediment transport (Lazarus et al., 2011; van den Berg et al., 2012; Ashton et al., 2001). Sediment transport gradients are determined by the coastline’s local orientation relative to incoming wave angles: where waves regularly approach the coast at high offshore wave angles (~>42 degrees relative to the local shoreline angle), an instability arises that causes perturbations to grow seaward and form sand waves, capes, and spits (Ashton et al., 2001). Likewise, on coasts where waves approach at low deep-water wave angles, perturbations tend to diffuse, and the coastline remains (mostly) smooth. Considerable attention has been given to the high-angle instability as an elegant and widely applicable explanation of plan-view coastline shapes, and multiple models have recently been developed to explore it (Ashton et al., 2001; Falques, 2003; Falques and Calvete, 2005; Ashton and Murray, 2006a, 2006b; van den Berg et al., 2012; Kaergaard and Fredsoe, 2013a;

Kaergaard and Fredsoe, 2013b; Kaergaard and Fredsoe, 2013c; Hurst et al., 2015). But, perplexing and subtle shoreline patterns also arise in low-angle wave climates (Figure 1) and have received much less consideration (Lazarus and Murray, 2011). Examples include low-relief seaward-convex ‘minor capes’ (Riggs et al., 1995) in North Carolina, False Cape near Cape Canaveral, Florida, USA (Kline, 2013), and a local shoreline bulge on Fire Island, New York (Schwab et al., 2000, 2013; Figure 1).

In addition to the high wave angle instability, coastline features can also be explained by the antecedent geologic framework that underlies the modern, active beach profile (Riggs et al., 1995; Schwab et al., 2000; Bender and Dean, 2004; Honeycutt and Krantz, 2003; McNinch, 2004; Browder and McNinch, 2006; Valvo et al., 2006; Mallinson et al., 2010; Denny et al., 2013; Lazarus and Murray, 2011; Thieler et al., 2014). Antecedent geology can affect shoreline change rates in several ways, including through the differential erosion of alongshore-varying lithology as the shoreface retreats landward, and the erosion of relict sediments that can supply sand to the active profile (Honeycutt and Krantz, 2003; Browder and McNinch, 2006; Lazarus and Murray, 2011). For example, Riggs et al. (1995) found that shoreline erosion rates in North Carolina, USA were highest along shoreline segments underlain by relict drainage channels that provided little sediment to the active shoreface, and that subtle, kilometers-long shoreline undulations were correlated with structural highs in the geologic framework. Using a numerical model, Valvo et al. (2006) simulated the plan-view evolution of a sandy shoreline underlain by different lithologies and found, in support of Riggs et al. (1995) and others, that small shoreline undulations (<100 m in total cross-shore relief or amplitude) can be maintained in steady-state by variations in antecedent lithologic strength and composition. Sand-poor, softer lithologies erode

rapidly and form recessed bays, whereas sand-rich, harder lithologies erode slowly and form small promontories.

Antecedent geology can also help create complex nearshore shoals or submarine headlands. Such bathymetric anomalies can alter the alongshore distribution of wave energy and induce gradients in alongshore sediment transport, even in the absence of high-angle waves. At smaller alongshore scales, McNinch (2004), Schupp et al., (2006), and Browder and McNinch (2006) observed spatial correlations between erosional hot spots and subaqueous shore-oblique sand bars, hypothesizing that local wave field modification over the complex bathymetry is responsible for local erosion. Similarly Riggs et al. (1995) qualitatively linked wave energy redistribution over nearshore shoals to shoreline change patterns in North Carolina. Bender and Dean (2004) developed an analytical model that examined how wave shoaling and refraction over nearshore shoals and dredge borrow pits influenced shoreline change over short time scales (days). They found that bathymetric anomalies altered incoming wave directions and set up gradients in alongshore sediment transport such that the shoreline tended to recede behind borrow pits and accrete behind and near shoals. Aforementioned numerical modeling by Valvo et al. (2006) over longer time scales ($>10^1$ years) treated wave transformation in a basic way and could not account for complex bathymetry or wave energy convergence and divergence. All examples in Figure 1 have prominent nearshore shoals that have been qualitatively linked to nearshore wave transformation and large-scale shoreline change (Riggs et al., 1995; Kline, 2013; Schwab et al., 2000; Schwab et al., 2013; Smith et al., 2008), but only one study by Idier et al. (2011) has quantitatively considered the two-way interactions between wave transformation over complex bathymetry and shoreline change. That study (Idier et al., 2011) modeled the coupling between wave transformation and shoreline change and suggested a plausible mechanism by

which low-angle waves locally refracted by nearshore shoals can cause shoreline undulations (megacusps) over time scales of days to weeks and alongshore scales of $10^2 - 10^3$ m.

In this paper, we explore how complex nearshore bathymetry can affect large-scale (>1 km) shoreline change in low-angle wave climates over $10^1 - 10^2$ year time scales. We use a modified form of the Coastline Evolution Model (CEM; Ashton et al., 2001; Ashton and Murray, 2006a, 2006b) that simulates shoreline change caused by gradients in alongshore sediment transport. For simplicity and efficiency, the original CEM shoals and refracts waves in the most basic way using linear wave theory and Snell's Law. The model assumes that bathymetric contours are shore-parallel and extend seaward only to a given shoreface depth (e.g. 10-20 m). This simplification inherently excludes alongshore energy convergence and divergence and wave transformation over the continental shelf and, as such, it cannot account for complex bathymetry. To better honor wave transformation processes, we dynamically coupled the CEM to a 3rd generation spectral wave model, Simulating WAVes Nearshore (SWAN; Booij et al., 1999; Ris et al., 1999). SWAN can transform waves over complex bathymetry and provide the nearshore pattern of wave forcing (i.e. wave height, direction, period) that is necessary for CEM to compute longshore sediment transport gradients. The nearshore conditions computed by SWAN have been used to evaluate "instantaneous" patterns of erosion and accretion (Adams et al., 2011), but have not yet been coupled to a large-scale model, such as CEM, which allows the shoreline morphology to evolve. Unlike Idier et al. (2011), however, the shelf bathymetry in this study remains stationary. Using the coupled models, we demonstrate a mechanism by which gradients in alongshore sediment transport can form steady-state, large-scale shoreline features such as 'minor capes' (Riggs et al., 1995) with cross-shore amplitudes (i.e. bay to cape tip) of up to 1.5 km. An improvement in our understanding of the links among nearshore bathymetry, wave

transformation, and alongshore sediment transport is fundamental to understanding of shoreline change patterns at the local, municipality level.

Model background and methods

The intention of CEM is not to model actual coastlines in detail or to accurately represent site-specific nearshore morphodynamics. CEM is an exploratory model (Murray, 2007) and instead provides a behavioral perspective of the dynamics between nearshore wave transformation and shoreline change. We are asking basic questions about the effects of bathymetric shoals on coastline evolution (e.g. do shoals build coastline bumps, and if so, what are the length and time scales involved?), rather than about the details of a particular shoreline. As a result, the modeled shorelines should correspond to actual shorelines qualitatively as opposed to being an exact match.

The CEM simulates large-scale shoreline behavior caused by gradients in alongshore sediment flux and conservation of sediment according to the following equation (Ashton and Murray, 2006a, 2006b):

$$\frac{d\eta}{dt} = -\frac{1}{D} \frac{\partial Q_s}{\partial x} \quad (1)$$

where η is shoreline position (e.g., mean high water or another elevation datum on the beach face); t is time; D is the shoreface depth, or depth over which accretion or erosion occurs; Q_s is volumetric alongshore sediment flux; and x is in the alongshore direction. The variable Q_s is expressed by the semi-empirical CERC equation (Komar, 1971; Ashton and Murray, 2006a, 2006b):

$$Q_s = KH_b^{5/2} \cos(\phi_b - \theta) \sin(\phi_b - \theta) \quad (2)$$

where K is an empirical constant that relates wave forcing to volumetric sediment flux (Ashton and Murray, 2006a); H_b is breaking wave height (here we use significant wave height); ϕ_b is

breaking wave angle, with zero being shore-normal and 90° shore-parallel; and θ is the local shoreline orientation. Sediment is assumed to be spread evenly over the depth of the active profile. As in most previous applications of CEM (e.g. Ashton and Murray 2006a, 2006b), Equation 1 is discretized over a cellular domain with periodic boundary conditions (PBCs) so that sediment is conserved. PBCs are often used where modeled coastline features repeat alongshore, or are unbounded; for example, to simulate a series of sand waves moving continuously along an open coast. Here, we are modeling a single shoreline feature (minor cape) induced by stationary offshore shoals, so an alternate choice would be to use open boundary conditions (e.g., Barkwith et al., 2014a). Unlike open boundaries, PBCs do not require decisions or assumptions to be made about sediment fluxes to and from the model boundaries. Our model is exploratory, so we use PBCs for simplicity. Open boundary conditions, however, may be more appropriate for simulating a specific stretch of coastline in higher quantitative detail, especially if the boundaries are coincident with known sediment sources or sinks (Barkwith et al., 2014a).

The wave climate, or the time-averaged distribution of approach directions of the wave field relative to the orientation of coastline weighted by wave height and period, determines how the coastline will evolve. Two key dimensionless parameters are used to define the statistical distribution of wave climate: ‘highness’ (U) and ‘asymmetry’ (A). Highness quantifies the proportion of alongshore transport that occurs under high-angle wave conditions (defined previously by Ashton and Murray (2006a) as >42 degrees at the base of the shoreface) in the overall wave climate. For example, a highness value of 0 means that deep-water high angle waves are not witnessed for the coastal location, and a highness value of 1 means that a wave climate is wholly high-angle. Asymmetry is the proportional influence of waves approaching the shoreline from the left and right directions. An asymmetry value of 0.5, for example, indicates a

161 wave climate with an equal influence on sediment transport from both directions. The wave
162 climate parameters U and A used here are different than those used in previous studies involving
163 CEM (e.g., Ashton and Murray, 2006a, 2006b). In those studies, U and A are defined at the base
164 of the shoreface – or the offshore extent of shore-parallel depth contours – in water depths less
165 than about 25 m. Here, U and A are defined at the offshore boundary of the outer computational
166 grid in 100 m water depth. So, we are transforming waves across the shelf from much greater
167 depths and depth contours are not necessarily shore-parallel. Because of that, U and A values,
168 and thus the long-term wave climate, are not directly comparable between our study and
169 previous studies that employed CEM. Given that we are simulating coastlines in low-angle wave
170 climates, highness values in deep water will be $0 \leq U < 0.5$. We test asymmetry values between
171 zero and one.

172 By replacing the CEM's existing basic wave transformation routine with a more robust
173 spectral wave transformation model, a broader array of coastline behaviors – especially those
174 influenced by nearshore shoals or irregular bathymetry – is open to model exploration.
175 Henceforward, the version of CEM that is coupled to SWAN will be called **CEMSWAN**, while
176 the original version with simple wave transformation will be called **CEM**. SWAN is a 3rd
177 generation spectral wave model developed by the Delft University of Technology (Booij et al.,
178 1999; Ris et al., 1999). It includes processes such as wave shoaling and refraction, wave
179 generation by wind, wave-wave interactions, and energy dissipation due to whitecapping, bottom
180 friction, and depth-induced breaking. For our purposes, SWAN is run using nested bathymetry
181 grids. Stationary wave boundary conditions are assigned to an outer, coarsely gridded
182 bathymetry domain that transforms waves and in turn provides boundary conditions at the
183 margins of an inner, finely gridded domain (Figure 2). The outer grid repeats itself alongshore to

stay consistent with the PBCs applied to CEM. It is made large enough to allow waves to fully transform before passing boundary conditions to the smaller nest grid. The finer-gridded nest bathymetry is coincident with the CEM domain, where wave-breaking conditions drive shoreline evolution. The simulations are based on the barrier island coast surrounding Rodanthe, North Carolina (Figure 1). Bathymetry grids are Coastal Relief Models downloaded from the National Geophysical Data Center (NGDC), resampled to 320 m and 960 m for the nested and coarse grids, respectively, from an original cell size of ~90 m. Thus smaller-scale bathymetric features, such as shore-oblique bars (e.g., Schupp et al., 2006) that can potentially affect shoreline change, cannot be resolved. Grids were also smoothed using a low-pass filter to reduce SWAN computation time. The coastline was manually straightened to provide synthetic initial model conditions from which the shoreline can grow and evolve (Figures 2 and 5). After resampling, smoothing, and altering the coastline shape, the bathymetry grids represent the broad-scale characteristics of the Rodanthe area but are not meant to be exact representations, which is in line with the exploratory spirit of CEM.

Wave boundary conditions, including significant wave height, period, and direction, are applied along the offshore, shore-parallel domain boundary in ~112 m water depth. The 112-m depth is the minimum depth for which wave conditions used in the model (see below) can be considered ‘deep-water’. Wave conditions are then transferred to the higher-resolution nest grid along all (cross-shore and alongshore) offshore boundaries (Figure 2). SWAN is run in stationary, 2-D mode with a 10-degree directional spread using a JONSWAP spectrum (Hasselmann et al., 1973); quadruplet wave-wave interactions are disabled, as is wave energy dissipation via breaking because we use a depth-induced breaking threshold instead (see below). To save time, SWAN is run every other model time step (1 day). The shoreline morphology does

not change enough during a single time step to make a significant difference in how the waves transform over the nearshore. For a single model simulation of 300 years, SWAN would therefore run 54,750 times. As a result, CEMSWAN model runs can take up to four weeks using 4-8 processor cores with 2000 megabytes of memory each.

Wave breaking conditions are found in SWAN by applying a depth-induced breaking threshold to the SWAN-calculated significant wave height grids. The threshold is expressed as $\gamma = H/d$, where H is significant nearshore wave height (determined in part by wave period and angle relative to the coastline) and d is water depth. Typical observed and experimental values of γ range from 0.5 – 1.5 (Komar, 1998). However, using SWAN, a lower value of γ is often necessary depending on the grid cell size, wave height, and shoreface slope. For example, taking $\gamma = 0.78$ (a commonly used value; Komar, 1998) and a breaking wave height of 1 m, the estimated wave breaking depth is 1.3 m. If we further assume a linear shoreface slope of ~ 0.01 , the wave breaks approximately 130 m offshore. In order to resolve the $\gamma = 0.78$ threshold, the *maximum* grid cell size should be 130 m. A maximum γ value, γ_{crit} , below which the wave-breaking threshold can be resolved for a given wave height and cell size is shown in Figure 4. List and Ashton (2007) used the same method to extract wave-breaking conditions from SWAN model output and also found that the value of γ must be less than what is typically used to identify the breakpoint. They used $\gamma = 0.3$, and found that varying γ between 0.2 and 0.4 had little effect on the results. In our experiments, model stability deteriorated significantly when $\gamma > 0.5$, and varying γ within the range $0.2 \leq \gamma < 0.5$ did not affect model behavior. Ideally, one would prefer the grid cell size to be as small as possible. But, iterating SWAN tens of thousands of times to simulate a century of shoreline change with CEMSWAN is computationally expensive. There is an optimal balance between reducing computation time (by increasing cell size) and

adequately resolving the wave-breaking threshold and bathymetric features (by decreasing cell size). Here we strike that balance with a cell size of 320 m and $\gamma = 0.2$.

As the shoreline changes position through time, the shoreface bathymetry must also change. Using a uniform slope of 0.01, shoreface bathymetry is interpolated linearly offshore from each shoreline cell during each time step (Figure 2). If the shoreline accretes, the shoreface will extend perpendicular seaward offshore until it intersects the nested bathymetry grid. Seaward of the intersection, the bathymetry is not altered. If the shoreline erodes, the shoreface leaves behind a flat ravinement surface at its offshore base of depth D . A consequence of this simple method is that the shoreface depth may not be constant in time or space. The alternative is to force the shoreface to a prescribed depth regardless of shelf bathymetry. But that can leave bathymetry artifacts at the base of the shoreface that induce discontinuities in nearshore wave height and direction. To maximize bathymetric continuity and accurate wave transformation, we let the shelf depth determine the intersection between it and the shoreface. This approach works sufficiently well for simple, low-curvature coastlines, like the ones presented here. However, for high-curvature coastlines with steep capes and recurved spits, shoreface evolution should be improved (e.g. Kaergaard and Fredsoe, 2013c; Hurst et al., 2015).

The models are coupled using the University of Florida High Performance Computing Cluster. Because CEM and SWAN ‘speak’ different languages (C and Fortran, respectively) a UNIX shell script sequentially executes the models and mediates the exchange of information between them. The order of operations is shown in Figure 3. At each time step (1 day), a deep-water wave angle relative to the coastline orientation and approach direction (i.e., from the left or right) is selected randomly, with replacement (meaning a given value can be picked multiple times), from a probability distribution function defined by U and A calculated from nearby wave

buoys. SWAN uses the wave angle and direction, as well as a specified wave height and period, as its boundary conditions. The waves are then transformed over the nested bathymetry grids every other time step. CEMSWAN imports the computed significant wave height and direction grids exported by SWAN, and finds the depth-induced wave-breaking threshold offshore of each shoreline cell. Wave breaking conditions are extracted from the cell containing the wave-breaking threshold. Next, CEMSWAN uses the wave breaking conditions to calculate alongshore sediment flux (and associated gradients) and updates the shoreline position. Shoreface bathymetry is interpolated linearly offshore until it reaches the maximum shoreface depth (D) or the shelf. Finally, the updated shoreline position and updated bathymetry are exported back to SWAN for the next computational iteration of wave transformation. The model is typically run for 80-425 model years.

The general influence of bathymetric anomalies on shoreline behavior is outlined in the Results section using a suite of initial, generic model simulations. The Rodanthe study site was chosen because 1) substantial shoals and a submarine headland are found offshore (Figure 2), 2) a ‘false’ or ‘minor’ cape exists there (Figure, 1; Riggs et al., 1995; Smith et al., 2008) that has been linked to wave transformation across the shoals, and 3) the town of Rodanthe is experiencing locally severe shoreline retreat of over 3 m/yr that may be related to the shoals and the shoreline configuration (Riggs et al., 1995). Eight model runs were completed to illustrate the effects of different wave climates (defined by highness, asymmetry, wave height, and wave period) and the presence of nearshore shoals on shoreline behavior.

Results

An initially straight shoreline position with featureless nearshore and shelf bathymetry that increases in depth monotonically was allowed to evolve by running CEMSWAN and serves

as a reference point for the model simulations with bathymetric anomalies (Figure 5A-5C). The wave climate is diffusional (all waves are low-angle; $U = 0$) with a wave height of 2 m and period of 12 s. We expect the shoreline position to remain straight given the diffusional wave climate (Ashton and Murray, 2006a). That is indeed the case (Figure 5A), and Figure 5B and 5C show the final time-averaged significant wave height and wave direction from several thousand SWAN outputs. There is no variation in the alongshore wave height and direction, and the mean wave direction is $\sim 0^\circ$ ($A = 0.5$; average wave rays maintain shore-perpendicular paths).

Six additional test model runs were completed with CEMSWAN to test model behavior with different wave heights, periods, and approach directions. For the additional runs, the initial shoreline was straight, like the reference simulation above; however, waves were transformed over nearshore shoals based on the nearshore bathymetry at Rodanthe, NC. Two characteristic wave height and period sets were used: 2 m and 12 s, and 1 m and 8 s. Wave asymmetry (A) values of 0, 0.5, and 1 were applied to each wave height and period set. For both wave sets, the shoreline accretes behind the shoal complex and erodes on either side of it, forming a ‘minor cape’. Results from an example with the larger wave set and wave asymmetry ($A=0.5$), the same wave conditions applied to the run in Figure 5A-C, are shown in Figure 5D-5F. Relative to the simulation with planar nearshore and shelf bathymetry (Figure 5A-C), the presence of irregular bathymetry alters the wave transformation process. As waves transform over the alongshore-variable depths, they refract and converge on the submarine headland. Coincident with the locally refracted wave direction, alongshore sediment transport is directed towards the shoreline behind the shoals where sediment converges and the coast accretes seaward. On either side of the shoal complex, there is a divergence of alongshore sediment flux and the shoreline retreats (Figures 5A and 6). The shoreline continues to accrete behind the shoals and erode along its

flanks until it reaches an equilibrium curvature at which gradients in alongshore sediment flux equal zero. Although Figure 5 shows results from only one of the model runs with shoals present, the shoreline evolution and wave shoaling patterns look similar for all runs.

The equilibrium cross-shore relief and the alongshore position of the minor cape exhibit dependence on wave climate asymmetry as well as wave height and period. Varying the wave climate asymmetry changes the direction and distribution of net alongshore sediment transport and the cape feature shifts alongshore: predictably, when all waves come from the left ($A = 1$), the equilibrium bump is shifted to the right, whereas when all waves come from the right ($A = 0$), the bump shifts to the left (Figure 7). Comparing results from the two different wave height/period input conditions, the cross-shore relief of the minor cape is larger for the 2 m, 12 s wave set relative to the 1 m, 8 s wave set (Figures 7 and 8). This is because the longer-period waves have deeper wave orbital motions and therefore refract more over the shoals, bending further from shore-normal. Correspondingly there is a larger gradient in alongshore sediment flux between the adjacent straight, shoal-less coast and the coast behind the shoals. Conversely, the shorter-period waves refract less across the shoals, and the gradient is less pronounced. As a result the equilibrium curvature is smaller for the milder wave set, and larger for the more energetic wave set. The minor cape associated with the smaller waves requires a longer time to reach equilibrium (~400 years) because Q_s is smaller (and Q_s gradients have smaller magnitude). Within each wave set, the $A = 0$ simulations (all waves approach from the right) produced the largest total shoreline relief, measured from the bay to the cape apex. The bathymetry is asymmetric in that it is shallower on the right side as compared to the left (Figure 5), which causes differences in the transformation of deep-water waves from right versus left approach directions. Consequently, gradients in alongshore sediment flux differ depending on wave

asymmetry, with slightly different equilibrium shoreline curvatures satisfying those gradients. In the case of $A = 0$, the shallower bathymetry on the right side of the submarine headland causes more wave refraction there and larger gradients in alongshore sediment flux across the rest of the domain, leading to larger curvature and planform relief of the shoreline.

CEMSWAN was run again using the same bathymetry but with a wave climate that is more representative of conditions at Rodanthe, NC. Using U.S. Army Corps of Engineers Wave Information Study (WIS) hindcasts from 1980-2010 (<http://wis.usace.army.mil/>) and the methods outlined in Ashton and Murray (2006b) for calculating a long-term wave climate, we found $U = 0.4$, $A = 0.6$, mean wave height is ~ 1.6 m, and mean period is 9 s. The results are similar to those presented above, as the shoreline evolves to a stable steady-state configuration with a planform relief of ~ 1300 m (Figures 7 and 8). The larger proportion of high-angle waves (within the overall low-angle wave climate) causes the shoreline to evolve more slowly, however. High-angle waves refract more across the shelf than low-angle waves and arrive to the coast with smaller heights and less energy to drive alongshore sediment transport. The shape of the model shoreline, with an aspect ratio of 0.06 (cross-shore relief of 1250 m divided by alongshore length of 22.4 km), is similar to the observed aspect ratio of 0.08 – 0.09 at Rodanthe (amplitude of approximately 2200 - 2500 m divided by alongshore length of 28 km).

Discussion

The difference in time to equilibrium for the different wave sets (Figure 8) can be considered in terms of characteristic landscape or shoreline diffusion (or anti-diffusion) equilibrium time scales (Haff, 2007; Lazarus et al., 2011). The time to equilibrium (t_0) is

$$t_0 \approx \frac{x^2}{\mu} \quad (3)$$

where x is the alongshore spatial scale, in this case ~22 km, and μ is shoreline diffusivity expressed as

$$\mu = \frac{K}{D} T^{1/5} H^{12/5} \quad (4)$$

where K is an empirical coefficient and D is the shoreface depth (Ashton and Murray, 2006a, 2006b). The ratio of the times to equilibrium, t^* , provides a compact comparison

$$t^* \approx \frac{t_{0,s}}{t_{0,l}} \approx \frac{T_s^{1/5} H_s^{12/5}}{T_l^{1/5} H_l^{12/5}} \quad (5)$$

where the subscripts l and s denote the large (2 m, 12 s) and small (1 m, 8 s) wave sets, respectively. The remaining terms (K , D , x) are the same for both wave sets and cancel out. Solving Eq. 5 gives $t^* \approx 0.2$, meaning that the time to equilibrium for the large wave set should be approximately 0.2 of the time for the small wave set. Results in Figure 8 show that this is indeed the case: time to equilibrium is ~80 years and ~400 years ($80/400 = 0.2$) for the large and small wave sets, respectively. Model results also show that the time to equilibrium using the time-averaged Rodanthe wave climate is ~170 years. To apply Eq. 2 to Rodanthe, we take $K = 0.15 \text{ m}^{3/5} \text{ s}^{-6/5}$, which is appropriate when using significant wave heights in the CERC equation (Ashton and Murray, 2006b), a mean shoreface depth of 10 m (Lazarus et al., 2011), and use the alongshore extent of the model domain ($x = 22,400 \text{ m}$). Using output from WIS hindcasts between 1980 and 2010, as above, the calculated time to equilibrium is ~152 years.

The equilibrium minor cape produced by the model is similar to that at Rodanthe, as are the observed long-term shoreline change rates between 1860 and 1999 (Miller et al., 2005). Rodanthe has experienced erosion on either side of the shoal, and accretion behind it, but the pattern is complicated by the existence of historic inlets on the north side of the minor cape (Figure 9). The most recent inlets, Loggerhead Inlet and New Inlet, were active during the 1860s and during the 1940s, respectively (Smith et al., 2008; Figure 9). Since the inlets have closed, the

shoreline retreat rates in this area have been among the highest in North Carolina, in some cases greater than 3 m/yr (N.C. Division of Coastal Management, 2011, <http://portal.ncdenr.org/web/cm/oceanfront-construction-setback>). The chronic erosion has been qualitatively attributed to wave energy focusing caused by nearshore bathymetry (e.g. Riggs et al., 1995). Our results suggest that the severe erosion at Rodanthe is a response of the coast readjusting to a stable equilibrium curvature (caused by wave energy focusing due to along nearshore bathymetry) after local, temporary perturbations (caused by nearby inlet breaching and subsequent closure). The time to equilibrium of 152 – 170 years is comparable to the time elapsed since the inlets closed – ~156 years – suggesting that the coastline is close to an equilibrium configuration. CEMSWAN cannot simulate inlet dynamics, however, so this explanation remains speculative and is worthy of further investigation.

Other examples of minor capes located landward of prominent nearshore bathymetric anomalies in low-angle wave climates are False Cape, Florida (near Cape Canaveral) and an area of Fire Island, New York (Figures 1 and 9). Both locations show coastline shapes and long-term shoreline change patterns (Morton and Miller, 2005; Hapke et al., 2010) qualitatively similar to those produced by CEMSWAN with shoreline accretion directly behind the shoals and erosion on either side (Figure 5D, Figure 9). The minor cape at Fire Island (Figure 1C) has an aspect ratio of approximately 0.06 ($A \approx 1.5$ km, $x \approx 23$ km), which is the same as the aspect ratio from our model results and slightly less than that of Rodanthe (0.08). Schwab et al. (2000, 2013) documented a submarine headland near the apex of the Fire Island minor cape, along with many oblique shoreface-connected sand ridges that influence nearshore wave transformation (Warner et al., 2014). The Fire Island shoals are also thought to be a cross-shore source of littoral sediment (Hapke et al., 2010; Schwab et al., 2013) that can also explain the observed shoreline

change signal. False Cape (Figure 1B) has an aspect ratio of approximately 0.1 ($A \approx 1.4$ km, $x \approx 14$ km) and is bounded offshore by a complex series of shallow ridges and sandy shoals that induce nearshore wave refraction (Kline, 2013). The NASA Kennedy Space Center is located along False Cape, where launch pads and critical launch infrastructure are threatened by rapid shoreline retreat north of cape apex (Figures 1 and 6, Kline, 2013). CEMSWAN simulations suggest a mechanism for how shoreline shapes like Rodanthe, Fire Island, and False Cape develop as a result of wave interactions with complex nearshore bathymetry. Locally observed behavior such as the rapid erosion at Rodanthe and the north side of False Cape is thus a cog within larger-scale regional shoreline change patterns. In the future, if long-term wave climate asymmetry shifts its directional distribution (Moore et al., 2013; Johnson et al., 2014), additional local and regional shoreline changes are likely if minor capes translate alongshore (Figure 7), as indicated by model simulations.

Idier et al. (2011) showed that the mechanism of nearshore shoal-induced refraction, discussed herein (Figure 6), could explain megacusp formation on alongshore scales of 10^2 - 10^3 m. They called the process Low-Angle Wave Instability (LAWI). The CEMSWAN model produces results similar to those predicted by LAWI, and supports the plausibility of LAWI, but operates on larger spatial scales (i.e. 10^4 m). In addition, unlike the model of Idier et al. (2011), the shoals in CEMSWAN do not interact with the shoreline and are static (unless they are overtaken by the shoreface as the minor cape grows seaward). If, however, the shoal complex continued to evolve seaward with the shoreline, the minor cape could continue to attract additional wave energy as it grew in cross-shore relief. Additional wave energy convergence would cause further cape growth, and in turn more wave energy convergence: a potential positive feedback (Idier et al., 2011).

A factor that is not explored here is the ‘prominence’ of the shoals, or how much they rise above the surrounding inner shelf relief and how much area they cover. We would expect cross-shore planform relief to decrease if the shoal area and/or vertical deviation above the shelf decreased, and vice versa. Because CEMSWAN is computationally expensive to run we did not resolve or consider smaller bathymetric features such as shore-oblique bars (Garnier et al., 2006), which can locally affect shoreline change (McNinch, 2004; Schupp et al., 2006; Browder and McNinch, 2006). But, the physical mechanism described here should also operate at smaller spatial scales, as shown by Idier et al. (2011). Using more detailed bathymetry and/or a smaller grid cell size at the Rodanthe site would likely alter the comparisons between the modeled and observed shoreline shapes – but we are making qualitative comparisons with the Rodanthe shoreline to see if the dynamics in the model seem relevant to the natural case. Redoing the model runs for Rodanthe or other sites with higher-resolution bathymetry (or a more complex model, see below) would provide additional insights to the dynamics presented here, but is beyond the scope of this initial work.

In fact, the spatial accuracy of the bathymetry may not be the limiting factor in the quantitative reliability of the model output. The model cell size of 320 m used here is quite large to accurately resolve wave-breaking characteristics given the restrictions on γ shown in Figure 4. For this introductory work on simple, low-curvature coastlines, we sacrifice accuracy to gain computational efficiency. But in some cases the opposite will be necessary, where accuracy (and precision) is absolutely necessary regardless of the decrease in efficiency. Modeling complex, high-curvature coastline features like spits and flying capes would require better resolution of the wave breakpoint. The instability arising from high-angle wave influence involves alongshore variations in breaking wave height related to local coastline orientation; where the coastline has a

higher angle relative to the offshore wave crests, more refraction will occur. That refraction reduces wave heights, delaying depth limited breaking, which fosters further refraction—and associated decreases in wave heights. The result is smaller waves breaking closer to shore for some coastline orientations than for others. Fully capturing this effect requires higher resolution in the cross-shore direction. The requirement for greater cross-shore resolution is exacerbated for waves with small offshore heights and short periods, which break close to shore in any case. One way to increase cross-shore resolution without significantly increasing computational time is to extract the wave characteristics from SWAN in a shore-normal direction from the local coastline at the depth of the shoreface, after the waves have transformed over the shelf, then use the ‘classic’ CEM shoaling routine. The classic routine uses linear wave theory and Snell’s law to bring the waves to the breakpoint, across the shore-parallel shoreface, at higher spatial resolution. This would allow the calculation of long-term wave climate (i.e. the parameters U and A) at the base of the shoreface, making the long-term wave climate transferable between this study and previous studies that have used CEM. Furthermore, with such an approach the U and A combinations that produce various coastline morphologies (Ashton and Murray, 2006a; 2006b) like flying spits and capes could be understood in the context of the entire shelf, and its inherent bathymetric complexity, rather than only the shoreface.

CEM is especially well suited to simulate large-scale shoreline behavior over long (centuries to millennia) time scales. But, a different shoreline model could be used instead, such as the well-known one-line model GENESIS (Hanson, 1989), or UNIBEST (e.g. Ruggiero et al., 2010; Szmytkiewicz et al., 2000). For wave transformation, the spectral wave model STWAVE (e.g. Goncalves et al., 2012) or a wave ray-based transformation model (e.g. Adams et al., 2002; Whitley, 2014) could be used in place of SWAN. Kaergaard and Fredsoe (2013c) completed a

detailed shoreline model that includes alongshore sediment transport, nearshore hydrodynamics, and a spectral wave model. The model is capable of handling complex bathymetry, high-curvature coastlines, and high-angle wave climates over millennial time scales. The authors also developed an effective method to evolve shoreface bathymetry and successfully modeled the evolution of a migrating spit in Denmark over 2,700 years. Kaergaard and Fredsoe's (2013c) model could be used to more rigorously compare the general predictions presented herein to actual coastlines, and to parse out what other hydrodynamic processes might be important to the shoreline's response to wave transformation over complex bathymetry. Although the quantitative details might differ, we would expect the basic behavior shown here by CEMSWAN to be similar to that produced by other models.

This study only scratches the surface of potential links among nearshore bathymetry, shoreline change, and coastal configurations that emerge in low-angle and high-angle wave climates. Future model additions and improvements could include dynamic shoals that are linked to the shoreline and exploration of shoreline features formed by high-angle waves and their relationship to shoals (e.g. McNinch and Wells, 1999; Park and Wells, 2005). Improving coupled model efficiency using, for example, the open-source Community Surface Dynamics Modeling System (CSDMS) framework (<https://csdms.colorado.edu/>), or using a more streamlined wave model, would improve the efficiency of running simulations at higher spatial resolutions and over longer time scales.

Conclusions

By coupling a large-scale shoreline change model to a spectral wave model, we have shown that wave transformation over complex nearshore bathymetry in low-angle wave climates can produce observable large-scale shoreline change (Figure 5): wave energy flux convergence

along nearshore shoals, and energy flux divergence along the adjacent coastline, promotes the growth of minor capes with equilibrium cross-shore amplitudes (planform relief) up to 1.5 km. The time scale of minor cape growth to equilibrium depends on long-term wave height and period (equation 5) and in our model runs it varied between 80 and 400 years for energetic (2 m, 12 s) and mild (1 m 8 s) wave conditions, respectively. The cross-shore relief of the shoreline undulations also vary with wave characteristics. Small-period waves cause smaller gradients in alongshore sediment flux, and thus a smaller shoreline curvature (and cross-shore relief) is required to reach equilibrium. At equilibrium, gradients in alongshore flux equal zero across the model shoreline. Varying wave approach direction shifts the apex of the cape alongshore: a dominant wave approach direction from the right (left) shifts the apex left (right). Modeled minor capes are qualitatively similar in aspect ratio to those observed along the East coast of the United States (Rodanthe, Fire Island, and False Cape; Figure 1) where nearshore bathymetric anomalies have previously been linked to shoreline change.

Acknowledgments

This work was supported by National Science Foundation Grant EAR-1053033. We wish to thank Dylan McNamara, Andrew Whitley, two anonymous reviewers, and the UF High Performance Computing Center for providing additional insight and technical support.

References

- Adams, P. N., Anderson, R. S., Revenaugh, J., 2002. Microseismic measurement of wave-energy delivery to a rocky coast. *Geology*, 30(10), 895-898.
- Adams, P. N., Inman, D. L., Lovering, J. L., 2011. Effects of climate change and wave direction on longshore sediment transport patterns in Southern California. *Climatic Change*, 109(S1), 211–228.
- Ashton, A., Murray, A. B., Arnoult, O., 2001. Formation of coastline features by large-scale instabilities induced by high-angle waves. *Nature*, 414(6861), 296-300.

510 Ashton, A. D., Murray, A. B., 2006a. High-angle wave instability and emergent shoreline
 511 shapes: 1. Modeling of sand waves, flying spits, and capes. *Journal of Geophysical Research:*
 512 *Earth Surface (2003–2012)*, 111(F4).
 513
 514 Ashton, A. D., Murray, A. B., 2006b. High-angle wave instability and emergent shoreline
 515 shapes: 2. Wave climate analysis and comparisons to nature. *Journal of Geophysical Research:*
 516 *Earth Surface (2003–2012)*, 111(F4).
 517
 518 Barkwith, A., Thomas, C. W., Limber, P. W., Ellis, M. A., Murray, A. B., 2014. Coastal
 519 vulnerability of a pinned, soft-cliff coastline. Part I, assessing the natural sensitivity to wave
 520 climate. *Earth Surface Dynamics*, 2, 295-308.
 521
 522 Bender, C. J., Dean, R. G., 2004. Potential shoreline changes induced by three-dimensional
 523 bathymetric anomalies with gradual transitions in depth. *Coastal engineering*, 51(11), 1143-
 524 1161.

 525 Booij, N., Ris, R. C., Holthuijsen, L. H., 1999. A third-generation wave model for coastal
 526 regions: 1. Model description and validation. *Journal of Geophysical Research: Oceans (1978–*
 527 *2012)*, 104(C4), 7649-7666.

 528 Browder, A. G., McNinch, J. E., 2006. Linking framework geology and nearshore morphology:
 529 correlation of paleo-channels with shore-oblique sandbars and gravel outcrops. *Marine*
 530 *Geology*, 231(1), 141-162.

 531 Denny, J. F., Schwab, W. C., Baldwin, W. E., Barnhardt, W. A., Gayes, P. T., Morton, R. A.,
 532 Voulgaris, G., 2013. Holocene sediment distribution on the inner continental shelf of
 533 northeastern South Carolina: Implications for the regional sediment budget and long-term
 534 shoreline response. *Continental Shelf Research*, 56, 56-70.

 535 Falqués, A., 2003. On the diffusivity in coastline dynamics. *Geophysical Research*
 536 *Letters*, 30(21).

 537 Falqués, A., Calvete, D., 2005. Large-scale dynamics of sandy coastlines: Diffusivity and
 538 instability. *Journal of Geophysical Research: Oceans (1978–2012)*, 110(C3).

 539 FitzGerald, D. M., 1996. Geomorphic variability and morphologic and sedimentologic controls
 540 on tidal inlets. *Journal of Coastal Research*, S.I. 23, 47-71.

 541 Garnier, R., Calvete, D., Falqués, A., Caballeria, M., 2006. Generation and nonlinear evolution
 542 of shore-oblique/transverse sand bars. *Journal of Fluid Mechanics*, 567, 327-360.

 543 Gonçalves, M., Rusu, E., Soares, C. G., 2012. Evaluation of the wave models SWAN and
 544 STWAVE in shallow water using nested schemes. *Maritime Engineering and Technology*, 481.
 545 Haff, P. K. (2007). The landscape Reynolds number and other dimensionless measures of Earth
 546 surface processes. *Geomorphology*, 91(3), 178-185.

547 Hanson, H., 1989. GENESIS: a generalized shoreline change numerical model. *Journal of*
548 *Coastal Research*, 1-27.
549

550 Hapke, C. J., Lentz, E. E., Gayes, P. T., McCoy, C. A., Hehre, R., Schwab, W. C., Williams, S.
551 J., 2010. A review of sediment budget imbalances along Fire Island, New York: can nearshore
552 geologic framework and patterns of shoreline change explain the deficit? *Journal of Coastal*
553 *Research*, 263, 510-522.

554 Hapke, C. J., Himmelstoss, E. A., Kratzmann, M. G., List, J. H., Thieler, E. R., 2011. *National*
555 *assessment of shoreline change; historical shoreline change along the New England and Mid-*
556 *Atlantic coasts*. US Geological Survey Open-file Report 2010-1118.
557

558 Hasselmann, K., Barnett, T. P., Bouws, E., Carlson, H., Cartwright, D. E., Enke, K., Walden, H.,
559 1973. *Measurements of wind-wave growth and swell decay during the Joint North Sea Wave*
560 *Project (JONSWAP)*. Deutsches Hydrographisches Institut.

561 Honeycutt, M. G., Krantz, D. E., 2003. Influence of the geologic framework on spatial variability
562 in long-term shoreline change, Cape Henlopen to Rehoboth Beach, Delaware. *Journal of Coastal*
563 *Research*, S.I. 38, 147-167.

564 Hurst, M. D., Barkwith, A., Ellis, M. A., Thomas, C. W., Murray, A. B., 2015. Exploring the
565 sensitivities of crenulate bay shorelines to wave climates using a new vector-based one-line
566 model. *Journal of Geophysical Research: Earth Surface*.
567

568 Idier, D., Falqués, A., Ruessink, B.G., Garnier, R., 2011. Shoreline instability under low-angle
569 wave incidence. *Journal of Geophysical Research*, 116, F04031, doi:[10.1029/2010JF001894](https://doi.org/10.1029/2010JF001894).

570 Johnson, J. M., Moore, L. J., Ells, K., Murray, A. B., Adams, P. N., MacKenzie, R. A., Jaeger, J.
571 M., 2015. Recent shifts in coastline change and shoreline stabilization linked to storm climate
572 change. *Earth Surface Processes and Landforms*, 40(5), 569-585.

573 Kaergaard, K., Fredsoe, J., 2013a. Numerical modeling of shoreline undulations part 1: Constant
574 wave climate. *Coastal Engineering*, 75, 64-76.

575 Kaergaard, K., Fredsoe, J., 2013b. Numerical modeling of shoreline undulations part 2: Varying
576 wave climate and comparison with observations. *Coastal Engineering*, 75, 77-90.

577 Kaergaard, K., Fredsoe, J., 2013c. A numerical shoreline model for shorelines with large
578 curvature. *Coastal Engineering*, 74, 19-32.

579 Kline, S. W., 2013. *Influence of Wave Energy Dissipation on the Geomorphic Behavior of Rocky*
580 *and Sandy Coasts* (Doctoral dissertation, University of Florida).

581 Komar, P. D., 1971. The mechanics of sand transport on beaches. *Journal of Geophysical*
582 *Research*, 76(3), 713-721.

583 Komar, P. D., 1998. Beach processes and sedimentation. Prentice Hall, New Jersey, 544 p.

584 Lazarus, E. D., Murray, A. B., 2011. An integrated hypothesis for regional patterns of shoreline
585 change along the Northern North Carolina Outer Banks, USA. *Marine Geology*, 281(1), 85-90.

586

587 Lazarus, E., Ashton, A., Murray, A. B., Tebbens, S., Burroughs, S., 2011. Cumulative versus
588 transient shoreline change: Dependencies on temporal and spatial scale. *Journal of Geophysical*
589 *Research: Earth Surface* (2003–2012), 116(F2).

590 List, J. H., Ashton, A. D., 2007. A circulation modeling approach for evaluating the conditions
591 for shoreline instabilities. In *Proceedings of Coastal Sediments* (Vol. 7, pp. 327-340).

592 Mallinson, D. J., Culver, S. J., Riggs, S. R., Thieler, E. R., Foster, D., Wehmler, J., Pierson, J.,
593 2010. Regional seismic stratigraphy and controls on the Quaternary evolution of the Cape
594 Hatteras region of the Atlantic passive margin, USA. *Marine Geology*, 268(1), 16-33.

595 McNinch, J. E., Wells, J. T., 1999. Sedimentary processes and depositional history of a cape-
596 associated shoal, Cape Lookout, North Carolina. *Marine Geology*, 158(1), 233-252.

597 McNinch, J. E., 2004. Geologic control in the nearshore: shore-oblique sandbars and shoreline
598 erosional hotspots, Mid-Atlantic Bight, USA. *Marine Geology*, 211(1), 121-141.

599 Moore, L. J., McNamara, D. E., Murray, A. B., Brenner, O., 2013. Observed changes in
600 hurricane-driven waves explain the dynamics of modern cusped shorelines. *Geophysical*
601 *Research Letters*, 40(22), 5867-5871.

602 Morton, R. A., Miller, T. L., 2005. National assessment of shoreline change: part 2, Historical
603 Shoreline Changes And Associated Coastal Land Loss Along The US Southeast Atlantic Coast.
604 *USGS Open File Report*, 2005-1401.

605

606 Murray, A.B., 2007. Reducing model complexity for explanation and prediction.
607 *Geomorphology*, 90(3), 178-191.

608

609 Park, J. Y., Wells, J. T., 2005. Longshore transport at Cape Lookout, North Carolina: shoal
610 evolution and the regional sediment budget. *Journal of coastal research*, 1-17.

611 Riggs, S. R., Cleary, W. J., Snyder, S. W., 1995. Influence of inherited geologic framework on
612 barrier shoreface morphology and dynamics. *Marine Geology*, 126(1), 213-234.

613 Ris, R. C., Holthuijsen, L. H., Booij, N., 1999. A third-generation wave model for coastal
614 regions: 2. Verification. *Journal of geophysical research*, 104(C4), 7667-7681.

615 Schupp, C.A., McNinch, J.E., List, J.H., 2006. Nearshore shore-oblique bars, gravel outcrops,
616 and their correlation to shoreline change. *Marine Geology*, 233(1-4), 63-79.

- Schwab, W. C., Thielor, E. R., Allen, J. R., Foster, D. S., Swift, B. A., Denny, J. F., 2000. Influence of inner-continental shelf geologic framework on the evolution and behavior of the barrier-island system between Fire Island Inlet and Shinnecock Inlet, Long Island, New York. *Journal of Coastal Research*, 408-422.
- Schwab, W. C., Baldwin, W. E., Hapke, C. J., Lentz, E. E., Gayes, P. T., Denny, J. F., Warner, J. C., 2013. Geologic evidence for onshore sediment transport from the inner continental shelf: Fire Island, New York. *Journal of Coastal Research*, 29(3), 526-544.
- Smith, C. G., Culver, S. J., Riggs, S. R., Ames, D., Corbett, D. R., Mallinson, D., 2008. Geospatial analysis of barrier island width of two segments of the Outer Banks, North Carolina, USA: anthropogenic curtailment of natural self-sustaining processes. *Journal of Coastal Research*, 70-83.
- Szmytkiewicz, M., Biegowski, J., Kaczmarek, L. M., Okroj, T., Ostrowski, R., Pruszek, Z., Skaja, M., 2000. Coastline changes nearby harbour structures: comparative analysis of one-line models versus field data. *Coastal Engineering*, 40(2), 119-139.
- Thielor, E. R., Foster, D. S., Himmelstoss, E. A., Mallinson, D. J., 2014. Geologic framework of the northern North Carolina, USA inner continental shelf and its influence on coastal evolution. *Marine Geology*, 348, 113-130.
- Valvo, L. M., Murray, A. B., Ashton, A., 2006. How does underlying geology affect coastline change? An initial modeling investigation. *Journal of Geophysical Research: Earth Surface* (2003–2012), 111(F2).
- van den Berg, N., Falqués, A., Ribas, F., 2012. Modeling large scale shoreline sand waves under oblique wave incidence. *Journal of Geophysical Research: Earth Surface* (2003–2012), 117(F3).
- Warner, J. C., List, J. H., Schwab, W. C., Voulgaris, G., Armstrong, B., Marshall, N., 2014. Inner-shelf circulation and sediment dynamics on a series of shoreface-connected ridges offshore of Fire Island, NY. *Ocean Dynamics*, 64(12), 1767-1781.
- Whitley, A.E., 2014. Improved numerical modeling of large-scale cusped coastline dynamics. Master's Thesis, University of North Carolina at Wilmington, 128 pp.
- Xu, T., 2015. Wave transformation and alongshore sediment transport due to obliquely oriented shoreface-connected ridges. Master's Thesis, Georgia Institute of Technology, 108 p.

Figure 1. Examples of ‘minor capes’. A: Rodanthe area, Outer Banks, North Carolina; B: False Cape, near Cape Canaveral, Florida; and C: Fire Island, New York. The dashed line is a reference line used to visually compare these examples with model results in subsequent figures. Photos from Google Earth.

Figure 2. SWAN and CEM computational grids, and a cross-shore schematic illustrating how the shoreface is updated through time.

Figure 3. Flow chart of CEM and SWAN coupling procedure.

Figure 4. Contour map of γ_{crit} as a function of bathymetry cell size and breaking wave height. γ_{crit} is the maximum wave breaking threshold (H_b/depth) that can be spatially resolved for a given wave height and grid resolution. For small wave heights and/or coarse grid resolutions, the wave breaking threshold must be smaller in order to shift the wave breaking position seaward into water depths greater than zero. Plotted breaking wave heights can result from any combination of deep-water wave height, period, or direction.

Figure 5. Final results from two model simulations. Both runs were forced with the same wave climate ($A = 0.5$, $U = 0$, $H_0 = 2$ m, $T = 12$ s) and initially straight shoreline position. However, in Model Run 1 (panels A-C) the nearshore bathymetry deepens monotonically with no alongshore variation, whereas for Model Run 2 (panels D-F) the bathymetry consists of a submarine headland and shoals (outlined by bathymetry contours). Panels B,C and E,F are the mean significant wave height and direction, respectively for the entire model run durations (~ 80 years); these are the forcing conditions that drive the shoreline evolution in Panels A and D. Run 1 yielded no shoreline change (Panel A) because wave height and direction are uniform alongshore (B,C). In Model Run 2, waves refract and converge on the shoals (panels E-F), causing gradients in alongshore sediment transport that evolve the shoreline to an equilibrium shape (panel D).

Figure 6. Schematic of how nearshore shoals affect transformation of a shore-normal wave field, which leads to gradients in alongshore sediment transport (Q_s) and areas of negative divergence (coastal accretion) behind the shoal.

Figure 7. Panels A and B show the equilibrium plan-view shoreline positions for the small and large wave sets with varying wave climate asymmetry, as well as Rodanthe. At the bottom, wave roses show the model wave approach angles relative to the shoreline orientation for each asymmetry value. Note that all waves for the test model runs are low-angle (< 42 degrees), reflecting a highness (U) value of 0.

Figure 8. Cross-shore minor cape relief through time during the entire model run duration for all six test model runs, plus Rodanthe.

Figure 9. Long-term (> 100 years) shoreline change patterns at Rodanthe, False Cape, and Fire Island. At all sites, areas immediately behind an observed nearshore shoal complex have accreted, while neighboring shorelines have undergone erosion. Bathymetry for Fire Island and False Cape

are shown to the right of the shoreline change plots. The shoreline change pattern at Rodanthe is complicated by historical inlets, shown by the 1860 shoreline position: as the inlets have closed, the shoreline has relaxed to an underlying equilibrium curvature created by the minor cape.

Figure 1

Figure 2

Figure 3

Pre-simulation

Set up bathymetry
grids, CEM domain,
SWAN control files,
wave climate

{ Run model, begin loop

Step 1

Select deep water
wave angle and
direction from
U, A distributions

Apply wave conditions
to offshore boundary

Step 2

Run SWAN
coarse grid

Apply wave
conditions to
nest boundaries

Step 3

Run SWAN
nest grid

Transfer wave height
and direction to CEM

Step 4

Run CEM,
update shoreline
position & nest
grid bathymetry

1 iteration = 1 timestep

Figure 4

Figure 5

Model Run 1 (A,B,C)

Model Run 2 (D,E,F)

Figure 6

Figure 7

Figure 8

Figure 9