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Wind Wave Growth and Dissipation in a Narrow, Fetch-Limited Estuary: Long Island Sound

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Abstract: The geometry of the Long Island Sound (LIS) renders the wave field fetch-limited and leads to marked differences between western and eastern areas. The mechanisms that contribute to the formation and dissipation of waves in the LIS are not well understood. We evaluated the ability of the wave module of a wave-coupled hydrodynamic model to simulate different wind–wave scenarios. We were unable to capture wave statistics correctly using existing meteorological model results for wind forcing due to the low resolution of the models and their inability to resolve the LIS coastline sufficiently. To solve this problem, we modified the wind fields using in situ wind observations from buoys. We optimized both the Komen and Jansen parameterizations for the LIS to better present the peak winds during storms. Waves in the LIS develop more quickly than simple theory predicts due to quadruplet nonlinear wave–wave interaction effects. Removing quadruplet nonlinear wave–wave interaction increases the time to full saturation by 50%. The spatial distribution of wave energy density input reveals the complex interaction between wind and waves in the LIS, with the area of greatest exposure receiving higher wave energy density. The interaction of nonlinear wave–wave interaction and whitecapping dissipation defines the shape of the directional spectrum along the LIS. Dissipation due to whitecapping and shoaling are the main parameters modulating a fully developed wave field.



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Keywords: spectral wave model; estuary; Long Island Sound; wave growth; wave dissipation

1. Introduction

Fetch limitation, strong tidal currents, and complex bathymetry are features common to estuaries and produce an extensive range of deep to shallow water dynamics. One of the crucial dynamics in estuaries is wind waves. Understanding the importance of wind waves in estuaries is essential for managing and protecting coastal ecosystems. In the Long Island Sound, wetlands and salt marshes such as Stratford Point and Chittenden Park coastline in Guilford are vulnerable to local waves and experienced severe erosion, as noted by Ilia [1]. Also, the awareness for protecting coastline using a living shoreline has increased in recent years, as noted by O'Donnell, J.E.D [2]. These green structures' functionality differs by the wave characteristics, e.g., wave frequencies, as noted by Jadhav et al. [3]. Developing a protection and resiliency plan for these valuable coastal ecosystems requires the understanding of detailed wave characteristics in the region.

In this study, we employed numerical modeling and observations to investigate the mechanisms involved in the generation and dissipation of waves in the Long Island Sound (LIS), a large estuary characterized by a limited fetch environment. Previous studies have examined the behavior of wave models in estuaries. For instance, Lin et al. [4] compared the structured Simulation of Waves Nearshore (SWAN) and the empirical model of the Great Lakes Environmental Research Laboratory (GLERL) with observations in Chesapeake Bay, focusing on the effect of sudden changes in wind direction. Both models demonstrated good performance in capturing such events.

There are few studies on waves in the LIS, though. Ilia et al. [5] assessed two unstructured wave models, SWAN and MIKE21SW, in the New Haven estuary, where wind and breakwaters are the primary parameters influencing waves in the estuary. These models exhibited similar behavior in most events. In one event with a sudden change in wind direction, both models failed to simulate waves inside New Haven Harbor correctly. Furthermore, Liu et al. [6] estimated extreme wave heights in the LIS using the unstructured model FVCOM-SWAVE.

In addition, the formation and/or dissipation of wind waves in estuaries were the interest of some studies in recent years. Jouon et al. [7] conducted a study in a southeast lagoon of New Caledonia, a coral reef lagoon. The comparison of the observations and results of WaveWATCHIII indicated that, while wave intensity shows good agreement, directional spreading exhibited discrepancies due to the use of the Extended Maximum Entropy Method (EMEM) in WaveWATCHIII. Alomar et al. [8] simulated wind waves in the Northwest Mediterranean Sea and concluded that the observed wave growth rate is faster than simulated wave growth rate. And finally, Chen et al. [9] estimated different dissipation rates in Delaware Bay using an approach similar to the one employed in the current study. They identified whitecapping as the primary source of dissipation in Delaware Bay with friction playing a minimal role in high-slope areas.

Despite extensive research on wave intensity and variation in complex geometries and estuaries like the LIS, the mechanisms contributing to the wave generation and dissipation remain poorly understood. The work presented here summarizes our evaluation of the modeled wave statistics for several types of storm conditions: easterlies, westerlies, easterly to westerly, and westerly to easterly. The coupled hydrodynamic wave model was first optimized and compared to observations to recreate observed and previously documented wave dynamics inside the estuary under fully hydrodynamic-wave-coupled runs. The rate of generation and dissipation during a synthetic easterly storm in western and central LIS were quantified and compared with observation and theoretical equations. Subsequently, the mechanisms governing generation and dissipation, as well as their spatial variation along the sound, are presented.

2. LIS Geometry and Climatology

2.1. Geometry

The LIS is the sixth-largest estuary in the United States. It is located on the east coast between Connecticut and Long Island, New York. It is highly urbanized and adjacent to the New York City metropolitan area. The LIS is connected to the Mid-Atlantic Bight and the Atlantic Ocean at the eastern end and through the East River tidal strait at the west. The largest fetch length in the LIS is approximately 150 km and is oriented along 75° east of north. The maximum width of the Sound is 33 km and the average depth is 19.2 m [10].

2.2. General Circulation and Winds in LIS

The LIS is a tidal estuary that is forced by sea surface height (SSH) displacements on the adjacent continental shelf. Tides inside the LIS are near one-quarter-wave resonance at the M_2 frequency [11], with the strongest tidal currents located in the eastern section of the estuary (i.e., at the mouth of the LIS). In this area, tidal velocities are oriented along the Sound at the surface with along-isobath directions near the bottom [10,11]. The LIS receives 75% of its freshwater input from the Connecticut River [12], which is located close to the mouth of the estuary (Figure 1). The location of the primary freshwater source near the estuary's mouth makes the LIS unusual relative to other estuaries. Wind forcing in LIS exhibits a marked periodicity with consistent seasonal variability, both in wind magnitude and direction [10,13]. LIS wind data from buoys WLIS, CLIS, and ELIS (see Figure 1) show wind stresses that are primarily to the northeast during summer and to the southeast during winter, with overall stronger forcing during the winter months [14,15]. Annual means of stress magnitudes are dominated by the winter months, and the CLIS and WLIS locations are significantly different with annual means of 0.026 N m^{-2} and 0.012 N m^{-2} ,

respectively [10]. The seasonality and strength of the wind forcing have direct implications for surface currents and wave generation, where the directionality of the forcing affects the fetch limitation of the system.

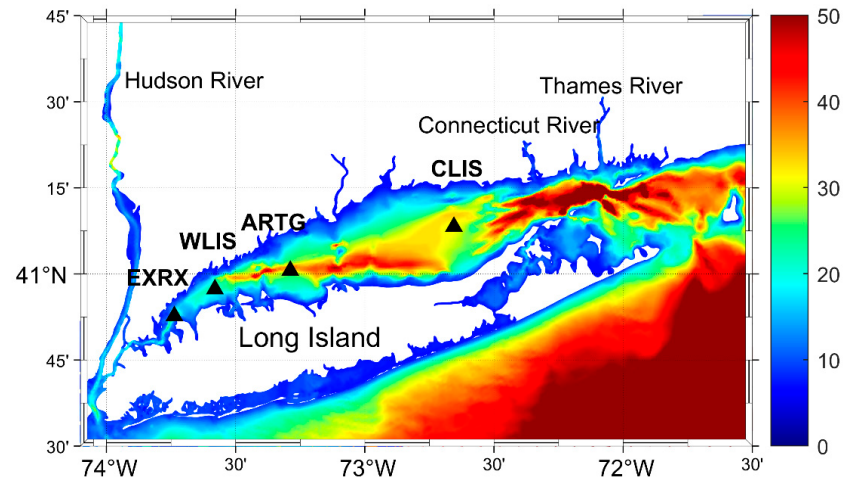


Figure 1. Model Domain Bathymetry for the Long Island Sound Estuary for the FVCOM-SWAVE model with a 250 m unstructured grid resolution (Δx , Δy). Locations of the buoys and instruments at EXRX, WLIS, ARTG, and CLIS are shown as black triangles.

The wind patterns in the LIS exhibit spatial variations. The central and eastern regions of the LIS experience stronger westerly winds, while the western region has relatively lower wind speeds and a higher occurrence of easterly winds. However, during storm events, the spatial variability of wind intensity decreases, as illustrated by the colored circles in Figure 2. This indicates a more uniform distribution of wind across the LIS during severe weather events.

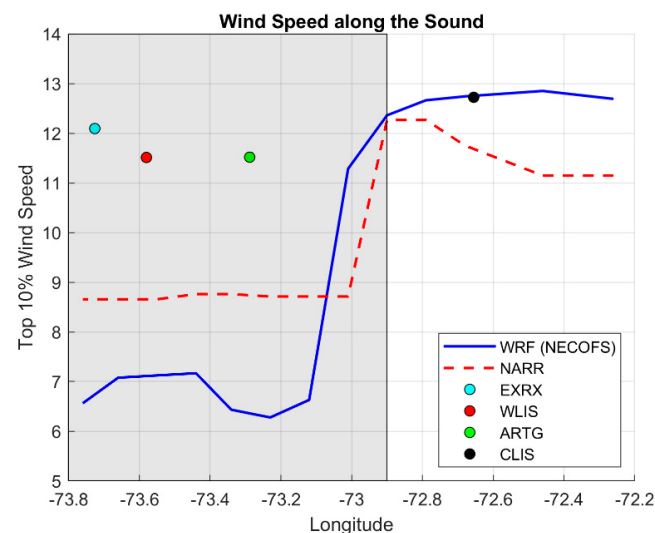


Figure 2. Yearly average of top 10% wind speeds along the Sound. The blue line shows the average of the top 10% wind speeds obtained from the WRF (NECOFS) model. The red dash line shows the average of the top 10% wind speeds obtained by the NARR climatology model. The dots represent the average of the top 10% wind speed at the buoys. The model's resolutions are not adequate to resolve wind fields in the western LIS. NARR low-biased storms in the eastern LIS. Both WRF (NECOFS) and NARR underestimated storms (top 10% wind speeds) in the western area of -72.9 longitude. WRF (NECOFS) was preferred over NARR in this study as it only needs to be assimilated on western longitudes, the gray area.

2.3. Oceanographic Observations

We used wave and meteorological data from the WLIS and CLIS buoy stations (lisicos.uconn.edu) for wave statistics within the LIS. We also explored meteorological data from ARTG and EXRX buoy stations, seen in Figure 1. These buoys are maintained and operated by the University of Connecticut and are equipped with met packages from R.M. Young, which include wind speed and direction, air temperature, barometric pressure, and relative humidity. The CLIS buoy is equipped with a directional wave sensor manufactured by Axys Technologies, sampling every 30 min for 22 min at 4 Hz. A non-directional wave sensor by Neptune Sciences was installed on the WLIS buoy and it sampled every 30 min for 17 min at 2 Hz.

3. Numerical Model

We used the wave module in the Finite Volume Community Ocean Model (FVCOM), described in [16–19], in the LIS region (Figure 1) to quantify the response of the estuary to severe weather events in a changing climate. FVCOM uses a 3D unstructured grid, free surface, and primitive equations to calculate hydrodynamics. FVCOM was coupled with the SWAVE wave module [20], which uses an unstructured finite volume version of the Simulating Waves Nearshore model, SWAN, [21] to resolve coastal scales where shallow water processes predominate. The hydrodynamic and wave models are coupled via radiation stresses, bottom boundary layer mechanics, and surface stresses [22].

The hydrodynamic module in this work builds on that developed by McCardell et al. [23]. By incorporating both hydrodynamic and wave processes in a coupled model, it becomes possible to capture the complex feedback mechanisms and interactions between waves, currents, and water levels.

The hydrodynamic model is forced at the open boundary with tidal components for the region, which were further refined based on tide gauge observations inside the estuary [23]. The riverine discharge was included using United States Geological Survey data. The horizontal grid resolution (Δx , Δy) was approximately 250 m within LIS with 41 sigma layers in the vertical direction (z). Atmospheric forcing was based on the Weather Research and Forecasting (WRF) model simulations from the Northeast Coastal Ocean Forecast System (NECOFS) project [24]. This wind field has a resolution of 9 km, which is insufficient to resolve differences between sea and land in coastal areas in western portions of the Long Island Sound. Therefore, we modified the WRF wind forcing so that simulations of waves in the western Long Island Sound were in better agreement with observations (see Section 3).

The turbulent closure model corresponds to the modified $k - \varepsilon$ for surface waves [25] from the General Ocean Turbulence Model [26,27], where k is the turbulent kinetic energy (TKE) and ε is the rate of dissipation of TKE. The bottom stress is determined by the prescribed bottom roughness length (z_0) and requires the current to match a logarithmic profile at height z_{ab} above the bed (depending on the local vertical model resolution). The surface boundary layer follows a wave breaking injection scheme where the TKE decays by a power law [26].

The SWAVE module is a direct adaptation of SWAN, and the reader is referred to the SWAN user manual, Qi et al. [20], and the FVCOM User Manual for further details on the numerical approach to the solution of the wave action equation. The SWAVE spectral frequency range was set to a range of 0.05–1 Hz with 32 logarithmically spaced frequencies and an angular distribution resolution of 10° (Δdeg) with 36 equally spaced angles. Spectral sources and sinks of wave energy were slightly modified to better capture the wave dynamics inside the estuary.

3.1. Wind Energy Input

The SWAVE wave module was run as a third-generation wave model (GEN3), where both the Komen [28,29] and the Janssen [30] wave growth parameterizations were initially evaluated. The total wind input from wind forcing was prescribed to have a linear and

exponential growth component t , where the total wind input (S_{in}) follows from the addition of the two growth parameterizations:

$$S_{in}(f, \theta) = A + B F(f, \theta) \quad (1)$$

where $F(f, \theta)$ is the frequency–direction sea surface spectrum. Parameter B is the exponential wave growth parameterization, and there are two options for it. The first follows Snyder et al. [28] with the friction velocity modification of Komen et al. [29] from the WAM Cycle 3 model (the WAMDI group, 1988). The second follows Janssen [30] from the WAM Cycle 4 model. We examined both B parameterizations in this study and determined the appropriate coefficient choices for the LIS. For the first expression (WAN cycle 3):

$$B = \max \left\{ 0, 0.25 \alpha \left(28 \left(\frac{U_*}{c_p} \right) \cos(\theta - \theta_w) - 1 \right) \sigma \right\} \quad (2)$$

where α is the ratio of air to seawater density, U_* is the friction velocity, c_p is the wave propagation velocity, θ is the direction of wave generation, θ_w is the wind direction, and σ is the wave frequency. Parameter A represents the linear growth parameter of Cavaleri and Malanotte-Rizzoli [31], which was active during all simulations.

$$A = \frac{c}{2\pi g^2} \left[(U_* \max(0, \cos(\theta - \theta_w)))^4 \right] \exp \left\{ - \left(\frac{\sigma}{\sigma_{PM}} \right)^{-4} \right\} \quad (3)$$

where c is constant and equals 1.5×10^{-3} , g is the gravitational acceleration, and σ_{PM} is the peak frequency as defined by Pierson–Moskowitz [32].

3.2. Energy Dissipation: Whitecapping

Wave dissipation is not well understood, and available parameterizations are highly empirical. The model was run using two parameterizations: (1) the generalized Komen et al. [29] wave breaking parameterization to complement the Snyder et al. [28] wave growth parameterization and (2) the Janssen [30] wave breaking parameterization and wave growth parameter. The spectral source term for both parameterizations can be stated as [21,33]:

$$S_{diss}(f, \theta) = \Gamma f_m \left(\frac{k}{k_m} \right) F(f, \theta) \quad (4)$$

where Γ is a steepness (s)-dependent coefficient [30]:

$$\Gamma = C_{ds} \left[(1 - \delta) + \delta \left(\frac{k}{k_m} \right)^n \right] \left(\frac{s}{s_m} \right)^p \quad (5)$$

where C_{ds} is an empirical coefficient of proportionality, s corresponds to the overall steepness parameter, p is the power of wave steepness impact, n is the power of wavenumber, k , and $0 \leq \delta \leq 1$ controls the wavenumber impact on whitecapping dissipation. The subscript m denotes an average where k_m follows from:

$$k_m = (\langle 1/\sqrt{k} \rangle)^{-2} \quad (6)$$

and the steepness parameter s follows from:

$$s = E(k_m)^2 g^{-2} \quad (7)$$

where E is the zero-order moment of the spectrum ($F(f, \theta)$).

3.3. Energy Dissipation: Bottom Friction

The spectral dissipation function due to bottom friction effects (S_{bf}) can be written in the general form [34]:

$$S_{bf}(k) = -C_f \frac{k}{\sinh(2kH)} G(k) \quad (8)$$

where k is the wavenumber, H is the water depth, C_f is a friction coefficient with units of velocity, e.g., m s^{-1} , and $G(k)$ is the wavenumber spectrum where $G(k) = F(f) df/dk$. The Madsen wave friction coefficient (f_w) follows as:

$$f_w = \begin{cases} 0.15 & ; \frac{a_b}{k_N} < 1.57 \\ \frac{1}{4\sqrt{f_w}} + \log_{10} \frac{1}{4\sqrt{f_w}} = m_f + \log_{10} \frac{a_b}{k_N} & ; \frac{a_b}{k_N} > 1.57 \end{cases} \quad (9)$$

where the roughness element length (k_N) corresponds to the actual physical roughness length (i.e., associated with the sediment size and distribution), a_b is the bottom excursion amplitude ($a_b = u_b \omega_b^{-1}$), where u_b is the bottom orbital velocity, ω_b is the bottom wave frequency, and m_f is a constant of value -0.08 . Then, the friction coefficient follows from:

$$C_f = \sqrt{2} f_w \langle u_b^2 \rangle^{1/2} \quad (10)$$

3.4. Quadruplet Nonlinear Interaction

Quadruplet nonlinear interactions were activated using the default settings for DIA [35]. This nonlinear wave–wave interaction transfers the wave action from high to low frequencies, redistributing the energy in the wave field. Triad nonlinear interactions were also active during these simulations; however, these interactions mainly affect very shallow water and surf zone regions.

4. Wind Field Modification

Two mesoscale atmospheric models, the Northeast and Coastal Ocean Forecast System (NECOFS) with 9 km resolution and the North American Regional Reanalysis (NARR) with about 32 km resolution, were analyzed for implementing the atmospheric forcing on the model. NECOFS calculates winds and surface heat fluxes using the Weather Research and Forecasting model (WRF). The NARR is produced by the National Centers for Environmental Prediction (NCEP) and assimilates observational data. In our analysis, we found out that both NECOFS and NARR underpredicted the wind speeds in the western area of the LIS (see Figure 2). This underestimation in the western LIS was particularly noticeable for higher wind speeds. For instance, NECOFS's annual means were biased low by a factor of 1.59 for wind speed greater than 4.5 m s^{-1} and by a factor of 1.83 for wind speeds greater than 10 m s^{-1} (Table 1). The NARR bias error in the WLIS region was lower than that of NECOFS, but the NARR bias extended throughout the entire LIS. We, therefore, chose to use WRF (NECOFS) for atmospheric forcing with modification of the western region to remove the underestimation bias.

The LIS is 30 km wide at the center and narrows in the western part to an average width of 10 km (longitude -73.75° to -73.25°). The NECOFS WRF model has a resolution of 9 km, which is not enough to resolve and differentiate the intensity of wind speeds between land and water surfaces in the western LIS. A similar issue was observed with NARR mask land data [36]. Because the LIS is a fetch-limited basin, wave heights are highly dependent on local winds [10]. Therefore, an underestimation of local winds leads to a substantial underestimation of simulated wave height in the western LIS. To improve the wave simulation, we used wind observations from the WLIS and ARTG buoys to adjust the western LIS region's WRF wind field.

Table 1. Wind underestimation by WRF (NECOFS) in the western LIS for each cardinal direction and the mean for all directions. These coefficients are used for modifying wind events in the western LIS for periods when there were no observations available.

	Wind Speed	Wind Underestimation Factors				
		North	East	South	West	Mean
WLIS	>4.0 m/s	1.68	1.71	1.56	1.34	1.59
	>10 m/s	2.09	2.02	1.73	1.51	1.83
ARTG	>4.0 m/s	1.93	1.89	1.96	1.87	1.92
	>10 m/s	2.48	2.45	2.22	2.08	2.27
EXRX	>4.0 m/s	1.96	1.87	1.95	1.85	1.92
	>10 m/s	2.44	2.37	2.13	2.03	2.23

The data collected at the buoy stations were recorded at a height of 3.5 m above sea level. In order to compare the recorded wind speeds with reanalysis data, the recorded values were adjusted to a standard height of 10 m using the law of the wall equation. This adjustment allows for better comparability and consistency with the reanalysis data, which is typically provided at a standard height of 10 m. The drag coefficient from Drennan et al. (2005) was used for the adjustment, ensuring better comparability and consistency with reanalysis data.

The wind field in the western LIS was matched to wind observations at the WLIS and ARTG buoy locations and then corrected elsewhere with the region based on the buoy locations' correction factors. We modified wind speeds for each cardinal wind direction independently. Therefore:

$$C_i^t = U_{i\text{obs}}^t / U_{i\text{WRF}}^t \quad (11)$$

where C_i^t is the correction factor at the buoy, i is the cardinal direction (north, east, south, and west), and t is the time. The correction factors were linearly interpolated between the buoys. The wind speed was modified as follows for the times with observation and without observation:

$$\begin{cases} Um_i^t = C_i^t \times U_i^t & \text{obs.} \\ Um_i^t = C_i \times U_i^t & \text{no obs.} \end{cases} \quad (12)$$

$$C_i = \frac{1}{n} \sum_{i=1}^n C_i^t \quad (13)$$

Here, only the events with wind speed larger than 4.5 m s^{-1} , which is the threshold for surface wave breaking, were corrected. For each time event, the wind speed correction in the field was limited to 1.30 times the observed wind speed at the nearest buoy (maximum spatial wind speed difference over the Sound). The correction of wind vectors was conducted separately for each direction. It is important to note that not all wind events underwent correction, as the correction was applied selectively based on the requirements of the analysis.

Table 2 shows a comparison of wind speeds at WLIS, ARTG, and EXRX before and after wind momentum flux correction. Statistical parameters are shown for all data periods and high wind events ($4 > \text{m s}^{-1}$) separately. The WLIS and ARTG buoys' observations were used for assimilation, and the observed data at EXRX were used for validation. The NECOFS wind data underestimated wind speeds at all buoys. After the calibration of the wind field, the bias decreased by 1.75 m s^{-1} and 2.19 m s^{-1} at WLIS and ARTG buoys, respectively. The RMS error at WLIS was dropped from 3.42 m s^{-1} to 0.60 m s^{-1} and at ARTG from 4.18 m s^{-1} to 0.90 m s^{-1} for wind speeds larger than 4.0 m s^{-1} . The correlation, r , at WLIS improved substantially, from 0.63 to 0.97, while at ARTG it increased from 0.48 to 0.94 for high wind speed events. At the EXRX station, the bias was decreased by 2.6 m s^{-1} ,

the RMS difference dropped by 1.95 m s^{-1} , and the correlation increased from 0.57 to 0.78 for wind speeds larger than 4.0 m s^{-1} . The WLIS and ARTG buoy data are used to estimate the assimilation factors, while the EXRX buoy data were used for assessment. Consequently, the correlation between WLIS and ARTG is higher than the one at EXRX.

Table 2. Wind field at the western LIS was modified with meteorological observations at WLIS and ARTG buoys and validated with meteorological data at the EXRX buoy. The table shows statistical comparison of wind speeds used for forcing the model before and after modification. WRF significantly underestimated the wind speeds in the western LIS. The wind data were modified using wind observations at WLIS and ARTG buoys and validated using the wind observations at EXRX buoy. The modification improved wind speed significantly.

		Before Wind Modification				After Wind Modification			
		N	Bias	RMS	COR	n	Bias	RMS	COR
EXRX	All Data	8761	−2.31	3.30	0.69	8761	−0.80	1.84	0.85
	> 4.0 m/s	5743	−3.45	3.98	0.58	5743	−1.19	2.03	0.78
WLIS	All Data	32,508	−1.93	2.79	0.79	32,508	−0.18	0.74	0.97
	> 4.0 m/s	20,617	−2.86	3.42	0.63	20,617	−0.14	0.60	0.97
ARTG	All Data	21,346	−2.34	3.45	0.56	21,346	−0.15	1.07	0.94
	> 4.0 m/s	13,735	−3.60	4.18	0.48	13,735	−0.22	0.90	0.94

Figure 3 shows scatter plots and statistics of modeled vs. observed significant wave heights at the WLIS location before and after wind modification. Before the adjustment of wind speed, significant wave heights were underestimated in the western LIS. For wave heights greater than 4 m, modeled wave height mean biases decreased from -0.39 m prior to the adjustment to -0.04 m after the adjustments, and the model–data correlation increased from 0.45 to 0.76.

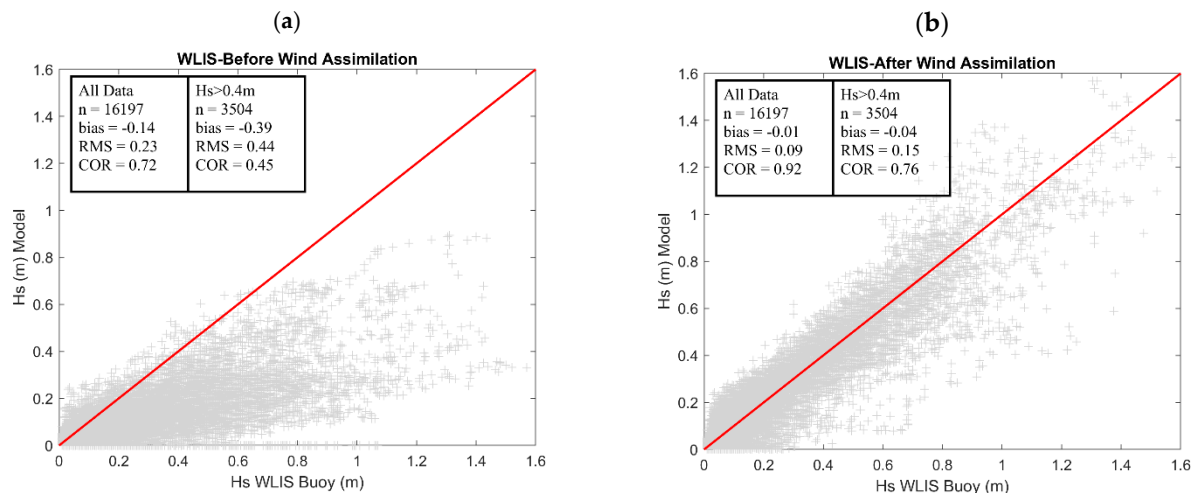


Figure 3. Scatter plots and statistical comparisons of significant wave heights at WLIS (a) before and (b) after wind speed modification. The significant wave height in the western LIS has been improved dramatically by wind modification. The red line shows the best fit, the 45-degree line.

5. Numerical Experiments

We performed numerical experiments to compare the two wave growth and white-capping dissipation parameterizations [29,30] and concluded that, with default coefficients, neither parameterization estimated significant wave heights well for storms peaks. The Komen coefficient values of $C_{ds} = 2.36 \times 10^{-5}$ and $\delta = 0$ (Equations (4) and (5)) resulted in overestimates of the significant wave height while the Janssen defaults of

$C_{ds} = 2.36 \times 10^{-5}$ (4.5) and $\delta = 0.5$ (Equations (4) and (5)) led to an underestimation of the significant wave height. Notice, in the model input, that the C_{ds} for Janssen is introduced as a value normalized by the square root of the wave steepness for a Pierson–Moskowitz spectrum. We also found the Komen scheme to present a larger high bias relative to observations. Nonetheless, it is more sensitive than the Janssen scheme; changing the coefficient in the Komen scheme led to a larger change in the results than in the Janssen scheme. The main difference between the two schemes lies in the parameterization of the exponential wave growth rather than differences in the whitecapping parameterization. The whitecapping in both schemes is almost the same, although Komen is equipped with some additional parameters such as the wavenumber exponent n (Equation (5)). The exponential wave growth parameter B in the Komen scheme, WAM cycle 3, is proportional to the inverse wave age $\left(\frac{U_*}{c_p}\right)$, while in Janssen’s parameterization, WAM cycle 4, it follows a quadratic representation of the inverse wave age with air–sea interaction at boundary layers. We optimized the model by changing both the Komen and Janssen whitecapping parameterizations for the LIS in order to decrease the significant wave height bias to zero for one full year of wave–current modeling. Increasing C_{ds} increases whitecap dissipation leading to smaller wave heights, while increasing δ leads to increased whitecap dissipation at higher frequencies.

Table 3 shows the optimized coefficients for storm peaks of 2014 for both the Janssen and Komen models. As the waves are quite young and therefore dissipative in the LIS, we had to increase the whitecap dissipation rate C_{ds} by 80% to enhance the model performance for capturing peaks in storms. Although both schemes showed similar performance in predicting significant wave heights, the Komen parameterization estimated peak wave periods for storms slightly better. Therefore, Komen was selected as the superior parameterization for stormy conditions.

Table 3. Optimized Komen and Janssen parameterizations for LIS. The C_{ds} definition for Janssen in the model input is the normalized C_{ds} , shown in parenthesis. p is the power of wave steepness impact, n is the power of wavenumber.

Method	C_{ds}	δ	n	p
Komen	4.72×10^{-5}	1.0	1.0	2.0
Janssen	3.19×10^{-5} (3.5)	0.5	-	-

The other dissipative term in these models is the bottom friction parameterization. Our results indicate that this parameter has almost no effect in the Sound near the CLIS and small effects on WLIS buoys (discussed in Section 6) but is an important term in coastal areas of depths less than 10 m. The western end of the LIS has a depth of ~25 m. Because wave peak periods during typical storm events are less than 6 s and wavelengths are less than 50 m, there is partial interaction with the seabed for frequencies lower than the peak frequency.

We used a roughness element length (kN) of 0.02 m to estimate the Madsen et al. [37] spectral parameterization. During all simulations, triad nonlinear wave–wave interactions were on, and the quadruplet nonlinear wave–wave interactions were solved using default DIA coefficients.

We examined the performance of the model in three types of storms. The first category had strong winds from the east. In these simulations, the long fetch leads to locally generated waves and swell from waves generated in the Block Island Sound (BIS). The second group includes westerly storms with limited fetch and only locally generated waves. The third category of storms begin with easterly winds, which then turn westerly, or vice versa. Storms like these occur a few times per year in the LIS. These were problematic for spectral wave simulations in New Haven Harbor, where the fetch length is limited to ~5 km [5].

The wave model results were compared to observations, and the model performance was evaluated based on the best fitted slope, m ; correlation, r ; constant, c ; and the root mean squared error, Equation (14); the model bias, Equation (15), states the under-over prediction and the index of agreement, which is a measure of the skill of the model, Equation (16) [38].

$$rms = \left[\langle (X - x)^2 \rangle \right]^{1/2} \quad (14)$$

$$bias = \frac{\sum_{i=1}^N (X_i - x_i)}{N} \quad (15)$$

$$IA = 1 - \left[\sum_{i=1}^N (X_i - x_i)^2 / \sum_{i=1}^N (|X_i - \langle x \rangle| + |x_i - \langle x \rangle|)^2 \right] \quad (16)$$

where brackets denote temporal averages, and X and x correspond to model and observations, respectively.

The model results were assessed for the moderate to strongest yearly easterly and westerly storms. In addition, the performance of the model was tested for two storms with a sudden wind direction shift. The latter case occurs when a low-pressure storm center passes close to the LIS. Simulated significant wave height in the LIS matches with observed significant wave height in WLIS and CLIS (Table 4). The peak wave period has a slightly lower correlation but is in good agreement with the observed values at buoys (Table 5).

Table 4. Statistical parameters at WLIS and CLIS for the significant wave height for three easterly storms, three westerly storms, and two direction-switching storms. The model could accurately simulate significant wave heights at both WLIS and CLIS.

Station	n	Bias (m)	rms (m)	m	r	c (m)	IA
WLIS	2237	0.07	0.11	1.01	0.95	0.08	0.95
CLIS	2235	0.06	0.16	0.97	0.92	0.08	0.95

Table 5. Statistical parameters at WLIS and CLIS for peak wave period for three easterly storms, three westerly storms, and two direction-switching storms.

Station	n	Bias (s)	rms (s)	m	r	c (s)	IA
WLIS	1134	0.04	0.46	0.68	0.76	1.24	0.84
CLIS	1740	−0.13	0.58	0.67	0.74	1.23	0.83

6. Wave Formation and Dissipation

The formation and dissipation of surface gravity waves in the LIS have not been extensively studied. By analyzing wind and wave observations in the western LIS, Shin et al. [39] concluded that, under westerly winds, waves follow well the established fetch- and duration-limited empirical formulae of Bretschneider [40] and Wilson [41]. Shin et al. [39] also showed that an “effective fetch” should be used to account for the narrowness of the Sound for winds from the east. For an easterly storm in the LIS, the fetch length is longer at WLIS (~130 km) than at CLIS (~50 km). To understand the need for the “effective fetch”, we investigated the terms that mainly control wave generation and dissipation and examined the effect of storm duration on wave growth in the LIS. We first simulated a synthetic storm with a constant wind along the longest fetch direction of the LIS (75°) using the optimal model parameters shown in Table 3. We used a wind speed of 12 m s^{−1}, which is the mean wind speed for the top 10% of storms in the LIS (Figure 2).

To examine the terms influencing the saturation state of waves, four different forcing and dissipation combinations were simulated. Figure 4 compares the results of saturation

state obtained from these simulations. The four combinations are (1) all wave dissipation and growth terms, (2) no refraction, bottom friction, or shoaling, (3) no nonlinear wave–wave interactions, and (4) no whitecapping dissipation. The comparison of the significant wave height results shows that the dissipation terms have a very significant role in the development and limitation of wave heights in both CLIS and WLIS. When all dissipation parameters and wave growth terms are enabled (the black solid and dashed lines in Figure 4), there is little difference between the significant wave heights at CLIS and WLIS during either the developing or saturation state. Wave growth stopped quickly, after 8 h, at both locations at similar significant wave heights. This number is 4 h smaller than the 12 h obtained by Shin et al. for the saturation time by analyzing observations. By 2 h after the onset of winds, significant wave heights reached 80% of their fully saturated state (8 h values). Removing the quadruplet nonlinear wave–wave interaction effects increased the time required to reach a developed sea state by 50% from 8 h to 12 h (black vs. green curve in Figure 4). Nonlinear interactions transfer wave energy to less dissipative frequencies and are, therefore, important for wave field energy growth.

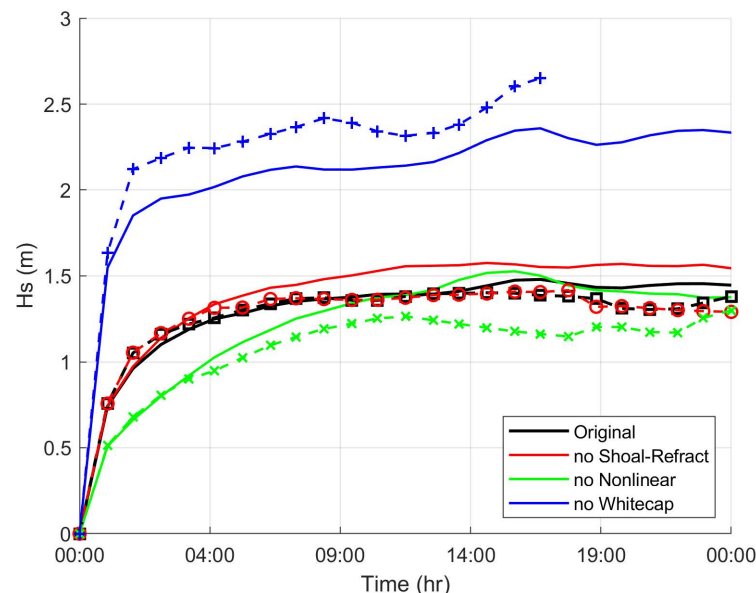


Figure 4. Modeled significant wave heights at WLIS (solid lines) and CLIS (dashed lines with marker) for an easterly storm with constant wind speed and unlimited duration for four configurations: (1) original forcing, black lines, (2) no shoaling or refraction by removing depth limitations, red lines, (3) no quadruplet nonlinear wave–wave interactions, green lines, (4) no whitecapping, blue lines. Quadruplet nonlinear wave–wave interaction has a major effect on the saturation time.

Whitecapping is the most important energy loss mechanism, and it dissipates most of the wave energy in the Sound. Shoaling and refraction are secondary effects. Quadruplet nonlinear wave–wave interaction does not produce or dissipate wave energy by itself. But by transferring energy from higher (dissipative) frequencies to lower frequencies, they lead to a decrease in the dissipation rate by whitecapping. To examine the variation of these terms, we also compared the spatial distributions of surface dissipation (e.g., whitecapping) versus bottom dissipation (e.g., friction, refraction, or shallow water breaking). We ran the model in six configurations: (1) comprehensive dynamics, (2) with no dissipation (growth only), (3) whitecap dissipation off, (4) nonlinear wave–wave interaction off, (5) no bed or shoaling effects, and (6) tidal forcing off to investigate the effect of tide on waves. For case 5, we turned off bottom friction and shallow water breaking and set the bathymetry to a constant depth of 100 m to remove the effects of refraction. Using the results of these model runs, we compared the dissipation due to surface vs. seabed effects as well as the change in wave energy density by tide and nonlinear wave–wave interaction.

Figure 5a shows the total mean wave energy density input per surface area for an easterly storm (75° from true north) with a constant wind speed of 12 m s^{-1} and unlimited duration. The wave energy input in the western LIS is less than that west of the widest area even though fetch length is larger for an easterly storm. For example, the fetch length for a northeastern storm from 75° at WLIS is 130 km, while at CLIS, it is 50 km. Figure 5a shows that the maximum total mean wave energy density input (4838 J m^{-2}) occurs not at the WLIS location (with the greatest fetch) but at midway between WLIS and CLIS. The interaction of wind and waves shown by $\cos$ in the Equations (2) and (3) indicates that the wave energy generated 60° away from the wind direction is equal to half of the wave energy generated along the wind direction. Therefore, the interaction of wind and waves operates in a way that the area with largest exposure can receive more wave energy density. The LIS is a narrow but relatively long estuary whose width varies from 5 km at the western end to about 30 km in the central area. Therefore, the wave climatology at any location in the LIS depends on a range of fetch values or exposure to a cumulative direction. When a northeasterly storm blows over the LIS's central portions, waves are generated from a broad range of directions with relatively large fetch. In the western LIS, waves are generated from a more limited set of directions. When all the energy in the storm is accumulated, the energy density input into the central region can be greater than that in the west. That is why areas with higher exposure, such as the wider region, in the LIS have greater energy density input than narrower regions. The maximum wave energy density input into the Sound for this simulation is between Port Jefferson, Long Island, and Stratford, Connecticut, where the combined effect of wind from all fetch directions is greatest.

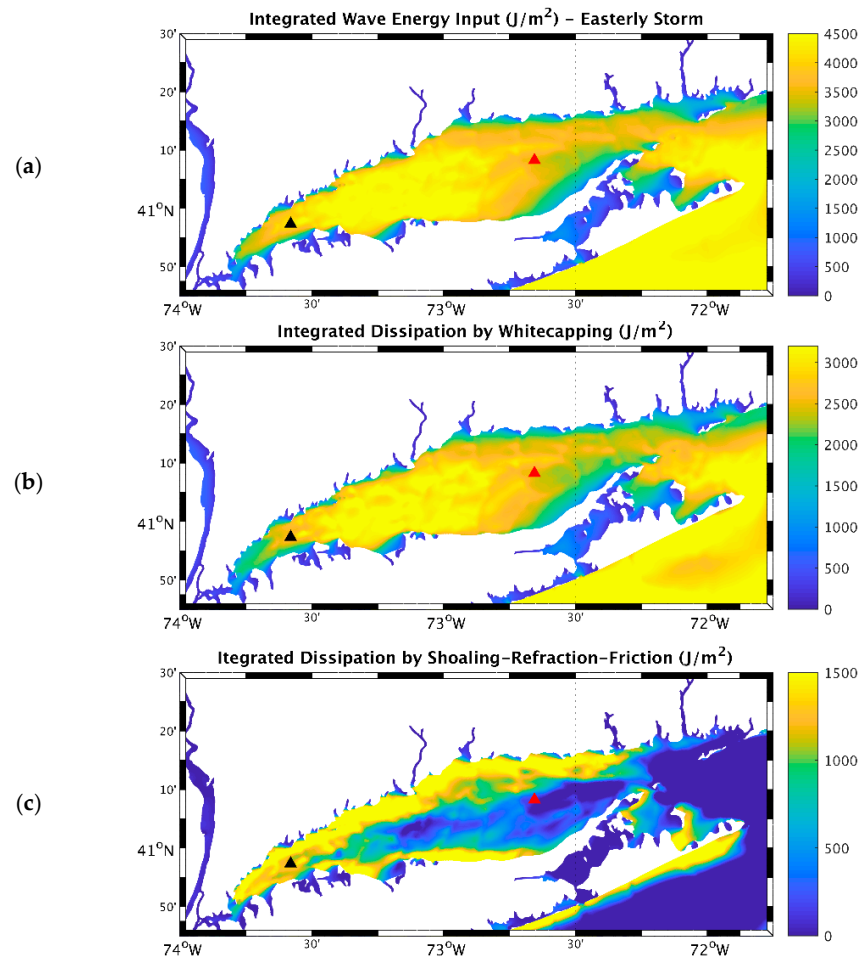


Figure 5. (a) Distribution of total mean wave input energy density per horizontal unit area in the LIS for an easterly storm with wind speed 12 m/s . (b) Total dissipation by whitecapping. (c) Dissipation by the seabed (e.g., shoaling, refraction, and friction). The locations of the WLIS and CLIS buoys are shown by the black and red triangles, respectively.

Figure 6 shows the directional wave spectra at WLIS, CLIS, and the point between them with the model's maximum wave density energy input (see Figure 5). Directionally, the CLIS spectrum is broader than the WLIS spectrum. The WLIS spectrum only has wave action from the 80° direction but with a broader frequency range. The broader frequency spectrum at WLIS is due to the longer travel of waves, which leads to a greater frequency transfer (nonlinear quadruplet interaction effect). Figure 6b shows that the spectrum at the intermediate location (which experiences the greatest wave energy during northeasterly storms) is broader in angle than either WLIS or CLIS because wave energy can enter from a wider range of directions with relatively large fetch lengths.

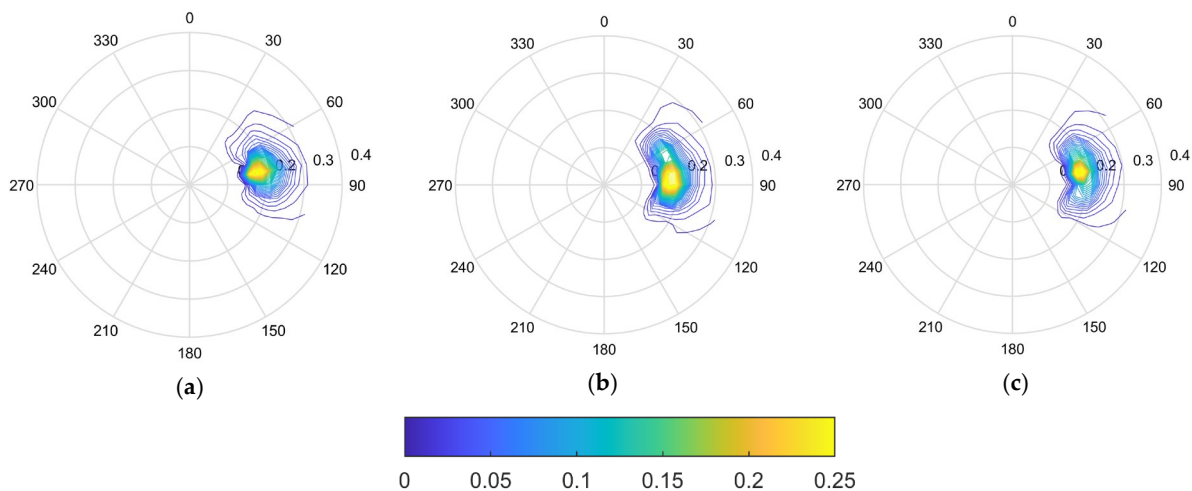


Figure 6. Directional wave spectrum for an easterly storm at (a) WLIS, (b) -73.1 (maximum wave height area), (c) CLIS. The wave action spectrum is monodirectional at WLIS due to narrow geometry. It is directionally broader at CLIS and broadest at a region between them where the maximum wave heights occur.

The dissipation terms are also critical to the spatial distribution of waves in the LIS and are the primary determinant of maximum significant wave heights. Figure 5b shows, for an easterly storm, the distribution of the rates of dissipation by whitecapping. Figure 5a,b shows that the areas of maximum whitecap dissipation match closely with the maximum wave energy input region: more wave energy leads to greater whitecapping dissipation. From Figure 5b–c, we can see that whitecapping is the primary dissipative mechanism in the LIS. Waves in the LIS are young and primarily generated locally with a range of relatively high frequencies; therefore, they are highly dissipative.

Bottom effects (shoaling, friction, and refraction) dissipate wave energy and transfer it to the coast. As shown in Figure 5c, the maximum wave energy density dissipation due to these terms happens near the coast. The WLIS station location is located in a narrow section of the LIS where dissipation by friction and refraction at the Long Island and Connecticut shores leads to a divergence of energy, or refraction, near the thalweg where the WLIS station is located. This divergence of energy takes place at mid-depth and is most prominent in the narrow section of the LIS where the effects of both northern and southern shores are combined and refraction creates a narrow spectrum, Figure 6a. In addition, the divergence of energy, or refraction, should be higher for lower-frequency waves which refract in deeper areas. Comparing Figure 5c, the bottom dissipation is the predominant dissipation in shallow water regions.

When quadruplet nonlinear wave–wave interaction is enabled in the model, wave heights reach the steady state more quickly due to the nonlinear interactions transferring energy from higher frequencies to more stable lower frequencies. Additionally, although quadruplet nonlinear wave–wave interaction does not produce or dissipate wave energy, it can affect other terms' dissipation rates. For instance, when quadruplet wave–wave interaction transfers wave energy from higher to lower frequencies, this causes a reduction

in whitecapping-driven dissipation. On the other hand, as the waves are transferred to lower frequencies, the shoaling, refraction, and bottom dissipation increase. Figure 7a shows the change in energy density due to nonlinear wave–wave interaction. The reduction in dissipation due to nonlinear effects is most significant in the central region because waves in this region are younger (less transferred frequency) than those in the west during easterlies. The results also indicate a dissipation increase due to refraction and bed friction enhancement in a relatively small region in the west, the blue region in Figure 7a. The western region has the lowest wave frequencies as more frequencies were transferred by quadruplet nonlinear wave–wave interaction in the west than east; therefore, low-frequency waves are affected by the bottom in deeper water.

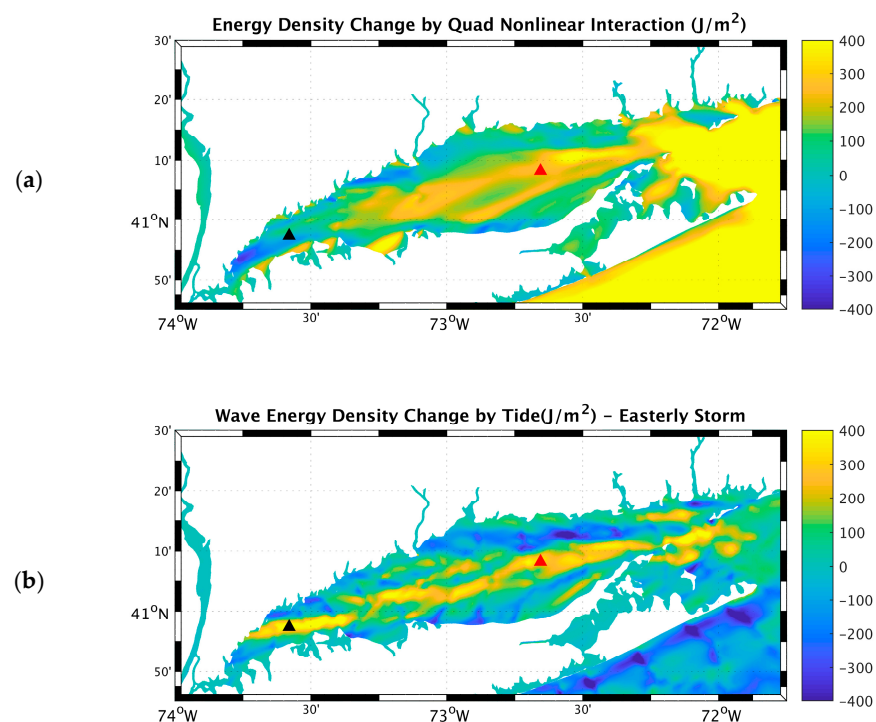


Figure 7. (a) Wave energy density change by nonlinear wave–wave interaction. (b) Wave energy density changes by tide.

Tidal currents and water level oscillations can refract and potentially amplify or diminish the dissipation by bathymetry. A recent study by Mengual et al. [42] suggested a 20–30% increase in wave height without tidal ebb current simulation at the mouth of the Tagus Estuary in Portugal. In this study, we faced limitations with the model employed, FVCOM-SWAVE, which prevented us from isolating the effects of tidal currents from tidal levels. To investigate the effect of tidal currents on waves, we should be able to run the wave model, SWAVE, alone, without coupling it to FVCOM. The wave model should be forced with tidal levels over the domain and no current. However, it seems that SWAVE cannot be run without coupling it with FVCOM. As a result, we were unable to investigate the effects of tidal currents alone on waves. Consequently, when we removed tidal forcing at the boundary, we encountered a decrease in the height of waves at both WLIS and CLIS. The effect of tides in the model, as shown in Figure 7b, leads to some change in the wave energy density. This change is in the order of magnitude of the energy density change due to quadruple nonlinear wave–wave interactions. However, the pattern of energy density change caused by tides in the LIS does not follow a clear trend, making it complex to analyze. To understand these effects better, it is essential to conduct a comprehensive investigation into the influence of tidal currents and levels on wave behavior. As shown in Figure 7b, the energy density change caused by tides is highest in the central deeper part of the LIS. Conversely, the most significant negative changes occur at the steepest slopes,

where refraction becomes a prominent process. Additionally, there are specific patterns in the southern part of Long Island that merit further investigation in future studies.

Following the scaling introduced by Kitaigorodski [43] and the data sorting method of Shin et al. [39], we compared the dependence of dimensionless wave energy and periods on dimensionless fetch length. The recorded data are for the period of 2006 to 2019. For easterly winds, $75^\circ \leq \theta_{\text{wind}} \leq 85^\circ$, Shin et al. [39] used a fetch length of 100 km for the WLIS buoy, and for westerly winds, $235^\circ \leq \theta_{\text{wind}} \leq 255^\circ$, a fetch length of 10 km was adopted. The black lines in Figure 8a,b show the relations of Kahma and Calkoen [44] and the blue circles and squares show the data of Shin et al. [39] for winds from the east and west, respectively. Clearly, for easterly winds, the wave energy is a factor of between two and four less than predicted by the Kahma and Calkoen [44] scaling for stable conditions. However, the scaling is consistent with data for westerly winds. The model results at the WLIS buoy location for the easterly and westerly storm simulations are shown by the red circles and squares. The data were obtained at WLIS from a one-year realistic wave modeling of 2014. These results are in broad agreement with the data and inconsistent with scaling when the wind is from the east. Also, the model results are relatively in agreement with the narrow fetch observation of Kahma and Petterssen [45,46]. These data suggest that, for the same wave age, the peak frequency change is less than wave energy change. However, some of the model results show some low peak frequencies that have not been observed.

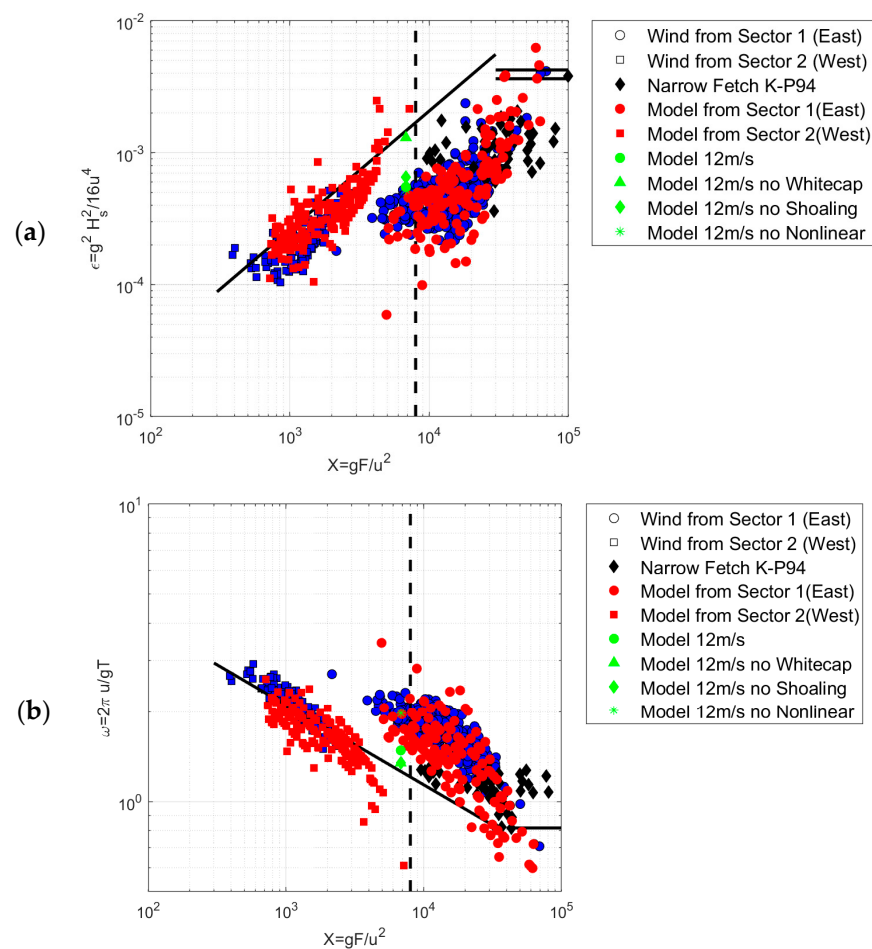


Figure 8. (a) The black line shows the dependence of the dimensionless wave field energy, estimated as $\epsilon = g^2 H_s^2 / 16 u^4$, on the dimensionless fetch, $X = gF / u^2$, as predicted by Kahma and Calkoen [44] under stable conditions. The black dashed lines show $X = 8000$, the upper bound of their observations. The horizontal lines show the limiting value for large X estimated by Alves et al. [47]. The circles and squares show ϵ estimated from the segment average data (blue) and model (red) for winds

from sector 1 (east) and sector 2 (west) using fetch estimates $F_1 = 100$ and $F_2 = 10$ km, respectively. The green dots show the ε for the unrealistic easterly storm with 12 m s^{-1} wind speed. The black diamonds show the narrow fetch observations (approximately) of Kahma and Petterssen [45,46]. (b) shows the same information for the dimensionless angular frequency $\omega = 2\pi u/gT$.

The green triangles in Figure 8a,b show the model results when it is forced with a constant easterly 12 m s^{-1} wind and with dissipation by whitecapping turned off. With less dissipation, the model is more consistent with the scaling. We concluded that the representation of whitecapping dissipation is critical to properly represent the effect of fetch and geometry in a small basin such as the LIS.

The overall wave height reduction by dissipation driven by large-scale wave breaking (i.e., whitecapping) was estimated at approximately 52% across the LIS, Figure 9a. This reduction factor varied from 30% in the western LIS to 60% in the eastern LIS, with most of the LIS between 45% and 55%. The wave height reduction factor due to whitecapping is not correlated with whitecapping energy dissipation (Figure 9a). The reduction factor is high both in regions of freshly generated waves such as rivers and coasts and higher energy developed regions where the whitecapping energy dissipation is maximal.

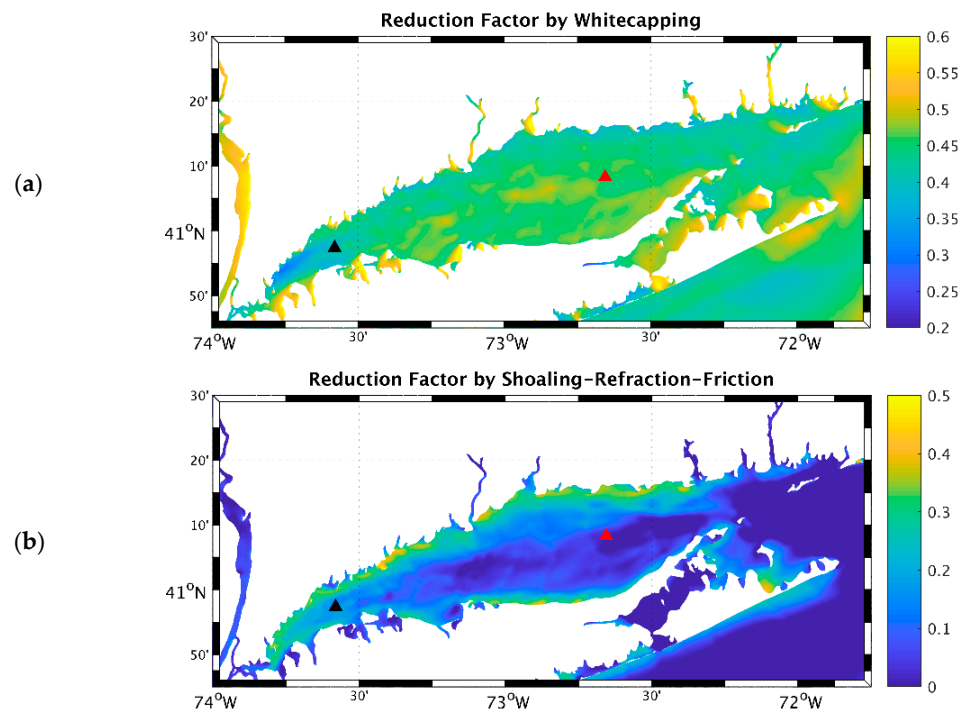


Figure 9. Wave height reduction due to dissipation by (a) whitecapping and (b) seabed interactions.

Figure 9b shows the reduction in maximum wave heights due to bottom effects. Shoaling and refraction are the most significant dissipative terms in shallow waters. The coastal area experienced a maximum reduction factor of 50% and an average reduction of 36% due to the dissipation by shoaling. The middle of the Sound is also affected by bottom dissipation effects but only by a reduction factor of 0 to 15%.

Although not presented here, wave energy density distributions during westerlies are similar to those during easterlies and discussed above. During westerlies, the Sound's wider areas experience more energy input than at narrow areas. For similar reasons, the spatial distributions of whitecapping and bottom dissipation energies during westerly events are similar to the easterly results shown above.

7. Impact of Swell Waves

In the present study, swell waves were initially disregarded due to the prevailing belief that they could not reach the western and central regions of the LIS. However, in response to a reviewer's request and the documentation of low-frequency waves by Shin et al. [39], a conservative approximation was conducted to assess the feasibility of swells in the western and central LIS. Additionally, a new set of observational evidence was provided.

The mouth of the LIS is narrow and located in the shadow of Block Island, Montauk Point, and Fisher Island. Swells propagating from 290 ± 20 degrees have the potential to reach the eastern part of the LIS by passing through the 7 km passage between Gull Island and Fishers Island. These swells may be diffracted about 45 degrees, then they move toward the central and western LIS. When the passage is much larger than the wavelength, similar to the situation in the LIS or a semi-infinite breakwater, the diffraction factor for swell wave height (the ratio of incident wave height to diffracted wave height) is smaller than that of sea waves behind the obstacle, see Figure 6 in [48]. For the LIS case, the wave height diffraction factor for a direction 45° away from the incident wave propagation direction is about 0.15, as determined by using the Goda (1978) diagram for a semi-infinite breakwater (Figure 6 in [48]). Additionally, the wave energy scattered behind the obstacle exhibits a greater ratio for low wave frequencies compared to high wave frequencies [48]:

$$S = \begin{cases} S_{max} \left(\frac{f}{f_p} \right)^5 & f \leq f_p \\ S_{max} \left(\frac{f}{f_p} \right)^{-2.5} & f \geq f_p \end{cases} \quad (17)$$

S represents wave spectra, f denotes wave frequency, and f_p is the peak wave frequency. Equation (17) indicates that wave energy spectra for lower frequencies are scattered more strongly (power 5) than those for higher frequencies (power 2.5), leading to the disappearance of low frequencies as the distance from the wave passage point increases, Figure 6 in [48]. Consequently, the wave period decreases behind the obstacle for diffracted swell waves; for instance, at a distance of 10 wavelengths from the passage, the wave period reduces to approximately 0.7 times that of the incident wave, as depicted in Figure 6 in [48] due to wave energy scattering. A comparison between the dominant wave periods observed in the eastern LIS and Block Island Sound (BIS) using a set of observations gathered in 2013 to investigate suitable sites for disposing dredge material confirms this finding [49]. Figure 10 illustrates the difference in observed dominant wave periods between the eastern LIS and BIS. As evident from Figure 10, there is a significant decrease in dominant wave periods from up to 12 s to up to 8 s when moving from the BIS to eastern LIS.

Consequently, utilizing this methodology, a 1 m wave height and period of 10 s in the BIS will be diffracted 45 degrees, transforming into a 0.15 m wave height and wave period of 7 s in the eastern LIS. As swell waves progress to higher frequencies after diffraction in the mouth of the LIS, they become more susceptible to strong dissipations, such as whitecapping, refraction, and bed friction. As they move towards the central and western LIS, the actual wave heights at WLIS become much less than 0.15 times the incident swell waves in the BIS.

It is crucial to note that this approximation is very conservative. Half of the passage between Gull Island and Fishers Island is less than 20 m deep, making it shallow water for swell waves. Consequently, this shallowness can lead to strong dissipation and refraction of the swell waves due to shallowing. On the other hand, the spectral wave models, such as SWAVE (SWAN) used in this study, neglect the diffraction process. Although some spectral models, like SWAN (SWAVE), incorporate an approximation for diffraction introduced by Holthuijsen et al. [50], this approximation has minimal impact on simulated wave heights, as demonstrated in Figure 13 of Ilia et al. [5]. Moreover, the reliability of this approximation for swell waves is lower [51]. Also, these models do not seem to be able to reproduce wave energy scattering due to diffraction.

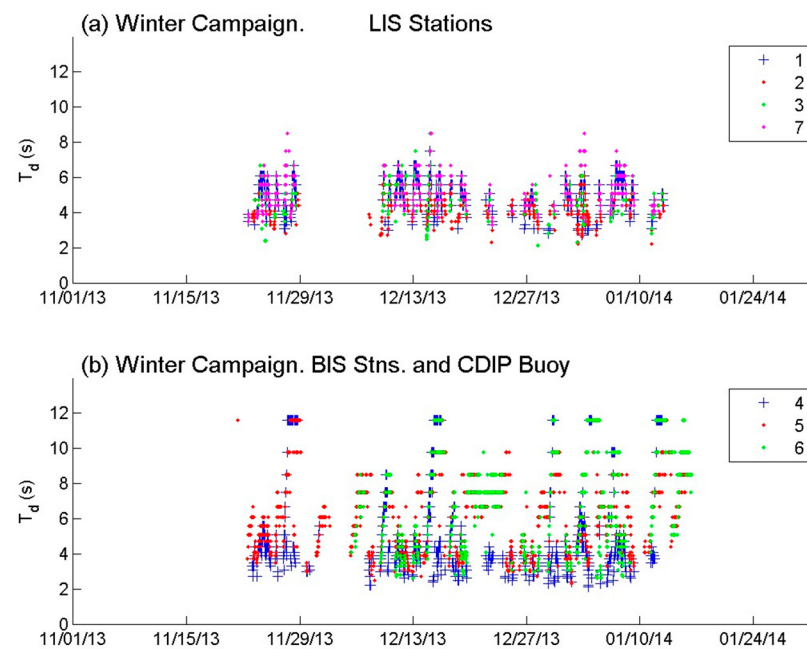


Figure 10. Observed dominant wave period at the (a) LIS and (b) BIS stations and CDIP buoy. Stations 1, 2, and 3 are in the eastern LIS; station 7 is at the passage between Fisher Island and Hull Island; stations 4, 5, and 6 are in the BIS [49].

It should be emphasized that for a detailed study of swell waves in a complex setting like the LIS, implementing a Boussinesq wave model and/or observed wave spectra in the LIS is recommended. Unfortunately, spectral wave observations are not available, and implementing a full Boussinesq wave model for the entire LIS can be exceedingly expensive. Nevertheless, as mentioned earlier, swell waves do not appear to play a critical role in the central and western LIS (with wave heights much less than 0.15 m due to dissipation and energy scattering). Therefore, neglecting these waves should not introduce a considerable bias in this study.

8. Conclusions

To simulate the LIS wave field structure under different forcing scenarios, we used the FVCOM circulation model coupled with SWAVE (a finite volume version of SWAN). We found that the accurate simulation of wind measurements in the western LIS required the introduction of empirical corrections to an operational regional atmospheric model (WRF). To match wave observations in stormy conditions, we also had to increase the dissipation rate by whitecapping by 80%, setting $C_{ds} = 4.72 \times 10^{-5}$, in the Komen parameterization. We also increased the effect of wave frequency on whitecapping by setting $\delta = 1.0$. In addition to improving our ability to forecast wave conditions in the LIS, we also used the model to examine the relative importance of dissipation and production terms to wave development in the LIS.

By comparing model predictions to buoy-based observations, we found that available mesoscale atmospheric models underestimated wind speeds in the western LIS by 50 to 90%. It seems likely that is due to their coarse spatial resolution so that the models do not adequately differentiate land from the water where the LIS is narrow. To provide more accurate wind fields to force the wave model, we developed correction factors for wind speed from WRF (NECOFS) in the western area of the Sound. After this modification, simulated significant wave heights in the western LIS improved dramatically, with a decrease in the RMS error of 0.09 to 0.23 m. Although we could implement corrections based on LIS buoy data, performing high-resolution atmospheric modeling for the LIS to resolve the wind field over the western LIS is advisable.

We performed several numerical experiments to evaluate the model's sensitivity to wave growth, whitecapping, and friction parameterizations. First, we evaluated the performance of the Janssen [30], WAM Cycle 4, and Komen et al. [29], WAM Cycle 3, wave growth and whitecapping parameterizations. We modified both parameterizations to improve the wave model in LIS. The optimized parameters we used for the LIS were significantly different from those used in other studies. For example, we had to increase whitecapping dissipation in the Komen parameterization by increasing the C_{ds} parameter to 4.72×10^{-5} . We also increased the effect of wavenumber on whitecapping dissipation to 1.0 to match simulated peak wave periods with observations. After these adjustments, both the Komen and Janssen parameterizations captured observed wave heights in the LIS at both CLIS and WLIS. However, the Komen parameterization was slightly superior in resolving peak wave periods. We also performed an analysis on easterly, westerly, and easterly to westerly (and vice versa) storms. The model had excellent performance for all three categories, although peak wave periods for westerly storms were slightly underpredicted at WLIS, where the fetch length is less than 15 km.

We forced the coupled model with a constant wind from the east (a synthetic easterly storm) to investigate the consequence of different choices for wave growth and dissipation representations on the saturation state. The waves in the LIS reached 80% of the saturation state after 2 h. However, full saturation took 8 h. Moreover, the removal of quadruplet nonlinear wave–wave interaction effects prolonged the time required to achieve a fully developed sea state from 8 h to 14 h (50%) and 80% saturation state from 2 h to 4 h (100%), emphasizing the importance of quadruplet nonlinear interactions in facilitating wave energy transfer to less dissipative frequencies and contributing to wave field energy growth.

The analysis of the special structure of the wave energy density revealed that the maximum storm mean wave energy density occurred between WLIS and CLIS, rather than at the location with the greatest fetch. This finding highlights the complex interaction between wind and waves, where the area with the largest exposure can receive higher wave energy density. Wider regions with higher exposure to directions that wind comes from exhibited greater energy density input. The examination of directional wave spectra further demonstrated it. The intermediate location experienced a broader directionally spectrum due to the entry of wave energy from a wider range of wind directions with significant fetch lengths. On the other hand, WLIS had a broader frequency spectrum compared to CLIS, demonstrating that waves at CLIS experienced less transfer of high frequencies to lower frequencies and, therefore, are more dissipative.

The results revealed that, for easterly winds, the wave energy was significantly lower than predicted by the method of Kahma and Calkoen [44] under stable conditions, while the model was consistent with data for forcing with westerly winds. Model results with dissipation by whitecapping turned off showed better agreement with the scaling method. This emphasizes the importance of accurately representing whitecapping dissipation in capturing the effects of fetch and geometry in the LIS. Good agreement with Kahma and Petterssen [47] suggests that, for the same wave age, the peak frequency change is less than the wave energy change. Certain discrepancies were found in the form of low peak frequencies that were not present in the observed data or data from Kahma and Petterssen [46].

The analysis of wave energy density changes caused by tides revealed a complex pattern without a specific trend, emphasizing the need for further research. Understanding the comprehensive impact of tidal currents and water levels on wave dissipation and energy density requires a thorough investigation, considering the limitations in parameterization. The removal of tidal forcing resulted in a wave energy increase for the middle of the LIS and a decrease in high-slope regions. Future studies should include investigation of specific patterns in the southern part of Long Island to enhance our understanding of wave dynamics in this region.

Author Contributions: Conceptualization, A.I.; methodology, A.I.; modeling, A.I.; validation, A.I.; formal analysis, A.I. and J.O.; data curation, A.I. and J.O.; writing—original draft preparation, A.I. and A.C.-L.; writing—review and editing, A.I., A.C.-L., G.M. and J.O.; visualization, A.I. and J.O.; supervision, J.O.; funding acquisition, J.O. All authors have read and agreed to the published version of the manuscript.

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Data Availability Statement: The buoy data used in this study can be found at UConn’s ERDDAP site: http://merlin.dms.uconn.edu:8080/erddap/tabledap/WLIS_Wave.html accessed on 1 January 2019 and http://merlin.dms.uconn.edu:8080/erddap/tabledap/CLIS_Wave.html. The NECOFS WRF data can be accessed at http://www.smast.umassd.edu:8080/thredds/catalog/models/fvcom/NECOFS/Archive/meteorology_v1/catalog.html.

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Conflicts of Interest: There are no conflict of interest.

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