A Simple Model for Coastal Sea Level Prediction

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(Manuscript received 31 July 2003, in final form 10 November 2003)

ABSTRACT

Reliable forecasting of wind-forced coastal sea level on the synoptic scale is available for most of the coastal areas of the United States through the National Weather Service Extratropical Storm Surge Model (ESSM). However, in many coastal areas around the world, especially in underdeveloped countries, little if any sea level forecasting is available, despite the often acute need. Here a simple linear-regression model based on modest wind forecast capability and records of local coastal sea level, wind, and pressure is developed and tested. Despite its simplicity, the model is based on robust ocean dynamics, in particular coastal Ekman circulation principles. The empirical model is tested using sea level observations at Atlantic City, New Jersey. The performance of the model is comparable to that of ESSM. For the 2-yr time period 1997–98, ESSM explains 79% of the total observed subtidal frequency sea level while the model presented here explains 74%. The root-mean-square errors in sea level for ESSM and the current model are 0.136 and 0.115 m, respectively. This performance indicates that the empirical model is adequate for general use in regions where reliable sea level forecasts from a circulation model such as ESSM are not available.

1. Introduction

We generally associate severe coastal flooding with hurricane landfalls and the accompanying storm surge. Nevertheless, in many coastal communities, even those subject to hurricanes, the storm of record for flooding is often caused by synoptic-scale wind events, such as winter storms. A typical example is Atlantic City, New Jersey. The costliest flood there was the “northeaster” of March 1962 which devastated the city.

Elevated sea level is not the only troublesome outcome of coastal wind events. Its companion, depressed sea level, can render navigation of coastal bays and harbors difficult and hazardous. Misleadingly called blowout tides in the marine vernacular, sea level depression can have the same magnitude in height as floods, typically 1 m.

Reliable forecasting of wind-forced coastal sea level on the synoptic scale has been available for most of the coastal areas of the United States for years. The National Weather Service (NWS) employs the Extratropical Storm Surge Model (ESSM; Kim et al. 1996; Chen and Shaffer 1999) with wind and pressure forcing from the NWS’s Aviation Model (AVN). The model is a vertically averaged barotropic coastal ocean model that includes specified bottom friction. Forecasts are issued for 12-h intervals out to 48 h.

In many coastal areas around the world, however, especially in underdeveloped countries, little sea level forecasting is available, despite the often acute need. Prospects for obtaining and operating the equivalent to ESSM are remote.

Here we develop and test a simple linear-regression model based on modest wind and pressure forecast capability and records of local coastal sea level, wind, and pressure. Despite its simplicity, the model is based on robust ocean dynamics and, in particular, on coastal Ekman circulation principles.

We will focus on the storm surge response, as it is known in the meteorological community, or on wind-forced response, as it is known in the oceanographic community. Neither our model nor the ESSM addresses the added height from breaking waves on beaches (Dean and Dalrymple 1991). While this contribution can be significant, it varies strongly with small-scale local beach and offshore bar topographic details. Consequently, we neglect these effects within this study. Sea level will also respond to astronomical tidal forces and to the horizontal gradient in water density or buoyancy. The latter response is at very low frequencies, cycles per week or month, and so can be included by resetting the forecast model initial sea level from coastal observation, as is done now in ESSM’s operation. To include tidal
forcing, it is convenient to decompose the total sea level into two parts:

$$\eta = \eta_t + \eta_w$$  \hfill (1)

where $\eta_t$ is the tidal response, a deterministic variable, and $\eta_w$ is the wind-forced response. Because the synoptic wind field has little energy at frequencies greater than diurnal, while tidal response is largely at diurnal and semidiurnal frequencies, we may obtain $\eta_w$ by performing a low-pass filter operation on $\eta$. Conversely, given a forecast of $\eta_w$, we can forecast the total sea level $\eta$ by simple addition. This procedure will be valid provided nonlinear interactions between wind forcing and tidal forcing are weak, as in most applications.

Our statistical model will be based on dynamical principles developed in physical oceanography. Ekman (1905) first derived the solution for the surface boundary layer or Ekman layer and then applied the theory to a coastal setting (see Fig. 1). To recover his solution we first state that all along-shelf ($x$ direction) gradients are negligible and that the subsurface flow has reached a state of equilibrium with the atmosphere. In particular, we neglect all along-shelf and temporal variability, or $\partial/\partial x = 0$ and $\partial/\partial t = 0$. This leads to a balance between wind stress and Coriolis force such that in the surface boundary layer a vertically averaged mass flux of $\tau' / f$ (where $\tau'$ is the surface wind stress and $f$ is the Coriolis parameter) is directed $90^\circ$ to the right of the stress in the Northern Hemisphere. The along-shelf component $\tau''$ will generate a component of the Ekman flux that is blocked by the coast. If this surface flux is offshore, sea level will fall at the coast as mass is exported offshore, and a subsurface onshore flow will develop to balance the offshore surface flux. Where this subsurface flux encounters the coast, it rises or upwells; hence, this mode of coastal circulation is known as coastal upwelling. For the reverse, the surface Ekman flow is onshore, coastal sea level rises, and the subsurface flow is offshore and sinking, a coastal downwelling circulation. Note that in this model the along-shelf wind, not the across-shelf wind, is the critical forcing agent.

As a rule, coastal flooding occurs during the season of weak stratification, because then the strong winds enhance vertical mixing. Usually the shelf dynamics then will be predominantly barotropic. A measure of this condition is the shelf Burger number $S = \alpha N / f$ (Clarke and Brink 1985), where $\alpha$ is the shelf bottom slope and $N$ is the vertical stratification. If $S \ll 1$, the sea level response to wind is mainly barotropic. Examples in which this restriction is found include the Mid-Atlantic Bight (except in summer) and the west Florida shelf (Weisberg et al. 2001). In contrast, where $S = O(1)$, such as off Peru, baroclinic dynamics and remotely forced sea level disturbances are common.

If we keep the steady-state constraint and restrict the dynamics to barotropic $S \ll 1$, but allow for small along-shelf variations, we pass to the model of Csanady (1978), termed the arrested topographic wave model. In
particular, this model requires that $L/\Lambda \ll 1$, where $\Lambda$ is the scale of the shelf length and $L$ is the shelf width. For the site we examine in this study, Atlantic City, New Jersey, the adjacent shelf has a typical width $L = 100$ km and a length $\Lambda = 1000$ km (see Fig. 1). Most continental shelves have this slender configuration, but there are exceptions. In this model the general features of Ekman’s coastal circulation are found, but with increased generality, including along-shelf variations of sea level. The across-shelf momentum equation becomes geostrophic, that is, $fu = -g(\partial \eta / \partial y)$, where $g$ is the acceleration of gravity, $u$ is the along-shelf current, and $y$ is the across-shelf coordinate, positive seaward. The along-shelf momentum equation includes both the wind stress and bottom stress in the along-shelf direction.

If we further relax the constraints and now permit time variability but restrict it to subinertial frequencies $\sigma \ll f$, we find that wave properties emerge, those of coastally trapped waves (Clarke and Brink 1985). The free wave manifestation of these waves travels downshelf, or in the same direction as winds that produce downwelling circulation (the direction we assign to positive $x$ in Fig. 1). Forced wave motion from time-varying and translating wind fields is now also possible. The geometric restriction $L/\Lambda \ll 1$ and the constraint $S \ll 1$ keep the along-shelf variations small and the response predominantly the barotropic local one found by Ekman (Lopez and Clarke 1989), however.

These previous studies have ignored the effect of across-shelf winds; however, Tilburg (2003) has shown that across-shelf winds can also affect coastal sea level. Following the derivations of Sandstrom (1980) and Tilburg (2003), we can estimate the magnitude of the sea level response to winds by examining the vertically integrated barotropic equations of motion governing transport on a continental shelf:

\[
\frac{\partial u}{\partial t} + fu = -g \frac{\partial \eta}{\partial y} + \frac{\tau^v - \tau^b}{\rho D(y)} \quad \text{and} \quad \frac{\partial v}{\partial t} + fv = \frac{\tau^u - \tau^b}{\rho D(y)}, \tag{3}
\]

where $D(y)$ is the bottom depth as a function of across-shelf distance, $\rho$ is the water density, and $\tau^v$ and $\tau^b$ are the bottom stresses. Since there is no flow through the coast, the vertically averaged across-shelf velocity $u$ and the across-shelf bottom stress $\tau^b$ can usually be neglected. Sandstrom (1980) shows that, for wind fluctuations on time scales greater than Csanady’s (1974) frictional adjustment time $[T_f = D(y)\sqrt{\rho(c_s^{-1} + \tau^v)}]$, where $c_s$ is the free surface drag coefficient, the acceleration term $\partial u/\partial t$ in Eq. (3) can be neglected and the along-shelf current $u$ adjusts so that the along-shelf bottom stress will be of the same order as the along-shelf wind stress. Taking both surface and bottom stresses following a quadratic drag law then gives

\[
\rho c_s q u = \rho c_s W W, \tag{4}
\]

where $c_s$ is the bottom drag coefficient, $q$ is the current speed, $W$ is the wind speed, and $W$ is the along-shelf wind. Since $v \ll u$ for the slender shelf configuration assumed, $q \approx u$; hence, $u = W \sqrt{(\rho c_s W)/\rho D(y)}$, which is compatible with the “mariner’s rule” that alongshore current is a few percent of wind speed. We neglect across-shelf variations in the wind. At higher frequencies (i.e., $\sigma \approx 1/T_f$), the acceleration term can no longer be neglected and introduces a lag between $u$ and $W$.

Substituting Eq. (4) into Eq. (2), neglecting $\tau^b$, and assuming a quadratic law for across-shelf surface stress, we obtain

\[
g \frac{d\eta}{dy} = -f W \sqrt{\frac{\rho c_s W}{\rho D(y)}} + \frac{\rho c_s c_s W W}{\rho D(y)}. \tag{5}
\]

This is essentially the Csanady (1978) model but with the inclusion of across-shelf surface stress, which will induce a free surface “setup” to balance $\partial \eta / \partial y$, in part. For a continental shelf characterized by a constant bottom slope $\alpha$ and a vertical coastal wall $D_o$, integration over $y$ of Eq. (5) from $y = 0$ at the coast to the shelf break (where $y = L$ and $\eta$ nearly vanishes when $S \ll 1$; Clarke and Brink 1985) gives a measure of the coastal response to across-shelf winds:

\[
\eta = \frac{f L W}{g} \frac{\rho c_s W}{\rho D(y)} - \frac{\rho c_s c_s W W}{\rho g D(y)} \ln \left(1 + \frac{\alpha L}{D_o}\right). \tag{6}
\]

Equation (6) predicts that the portion of coastal sea level $\eta$ due to along-shelf winds will be proportional to the Coriolis parameter, the square root of the along-shelf wind stress, and the shelf width $L$, while the portion due to across-shelf winds is invariant with $f$ and varies linearly with across-shelf wind stress, inversely with bottom slope, and logarithmically with $(\alpha L)/D_o$, the ratio of the water depth at the shelf break to the coastal wall depth. If we estimate properties for the Atlantic City application as $f = 10^{-4}$ s$^{-1}$, $L = 100$ km, $\rho = 1025$ kg m$^{-3}$, $\rho_o = 1.25$ kg m$^{-3}$, $c_s = 2.5 \times 10^{-3}$, $c_s = 1 \times 10^{-3}$, $\alpha = 10^{-3}$, $D_o = 2$ m, and $g = 10$ m s$^{-2}$, Eq. (6) becomes

\[
\eta_c = 0.022 W \sqrt{\frac{W}{W}} - 0.0001 48 W W, \tag{7}
\]

with sea level in meters and wind speed in meters per second. If we also include the inverse barometer effect (i.e., 1-hPa change in atmospheric pressure $P$ corresponds to approximately 0.01 m change in sea level), Eq. (7) becomes

\[
\eta_c = 0.022 W \sqrt{\frac{W}{W}} - 0.0001 48 W W + 0.01|P - P(t)|. \tag{8}
\]
Thus, an along-shelf wind of 10 m s\(^{-1}\) would give a sea level response of 0.22 m. An across-shelf wind speed of 10 m s\(^{-1}\) would give a sea level response of 0.05 m. The coastal response \(\eta_c\) is a weak function of the coastal wall depth. For an across-shelf wind speed of 10 m s\(^{-1}\), varying the coastal wall depth from 0.8 to 5 m changes the coastal response by less than 0.02 m. Although the along-shelf wind plays the leading role in our statistical model, we will see that both wind components produce a significant sea level response.

2. Numerical model

A numerical model is used in this investigation to examine the relative effects of across- and along-shelf winds, as well as to provide a relatively simple but dynamically accurate comparison for the statistical model. The numerical model used is the three-dimensional Estuarine Coastal Ocean Model (ECOM3d), a primitive equation finite-difference model in sigma (terrain following) coordinates based on the model developed by Blumberg and Mellor (1987). This model and a similar model, the Princeton Ocean Model, have been used extensively to study transport on coastal shelves (e.g., Allen et al. 1995; Austin and Lentz 2002). Consequently the model is only briefly described here. The interested reader is referred to Allen et al. (1995) for a more comprehensive description. For the simulations in this study, the model is run in a two-dimensional (across-shelf and vertical) channel configuration with no stratification. The model topography was selected to compare favorably with the shelf off Atlantic City and is given by 

\[ D(y) = D_0 + ay, \]

where \(D_0 = 2\) m and \(a = 10^{-3}\). An offshore wall is located approximately 150 km from the coast. There is no mass or heat flux through any boundary. Flow at the bottom boundary is governed by a quadratic drag law with a drag coefficient \(c_b = 2.5 \times 10^{-3}\). Vertical mixing is handled by the Mellor–Yamada level-2.5 turbulence closure scheme. Horizontal eddy viscosities are set to a constant \(A_m = 2\) m\(^2\) s\(^{-1}\). Since ECOM3d is a sigma-level model, the vertical resolution is proportional to water depth. For all simulations there are 41 sigma levels, so that the vertical resolution varies from 0.05 m at the coast to 3.75 m in the deepest water. The horizontal grid size for all experiments is \(\Delta x = 1\) km. Finite-difference time steps are 200 s for the baroclinic and 10 s for the barotropic components. Halving the vertical and horizontal grid sizes, halving the time steps, or extending the domain size have no appreciable effect on the simulations.

Two sets of simulations are performed: a series of “steady state” simulations designed to examine the relative roles of across- and along-shelf winds in determining sea level and a simulation driven by observed winds to provide a comparison with our statistical model. Both sets are driven by spatially uniform winds that are ramped up over 1 day. The steady-state simulations are then subjected to a constant wind stress for 9 days, at which time the simulation has reached steady state. The other simulations are driven by observed winds obtained from the Atlantic City meteorological office for 1997–98.

3. The data

For this investigation, our statistical model is used to examine the coastal sea level response at Atlantic City (Fig. 1). This site was selected because of the availability of the predicted sea level response from ESSM. Wind velocities and atmospheric pressure are obtained from the Atlantic City meteorological office of the NWS, while sea level is obtained from the National Ocean Service Water Station on Taj Mahal Pier in Atlantic City. Since the statistical model is designed to predict sea level response at subtidal time scales, sea level, wind velocities, and pressure are processed using a Lanczos low-pass filter with a cutoff frequency of 1/36 h\(^{-1}\) to remove tidal, inertial, and other high-frequency variations.

4. Contributions of across- and along-shelf winds

Subtidal coastal sea level is due to contributions from the along-shelf wind and, to a lesser extent, across-shelf wind stress and atmospheric pressure. Equation (6) provides a convenient analytic tool to examine the relative contributions of each wind component. In Fig. 2 we show the separate coastal sea level response due to the along-shelf wind component (solid black line) and the across-shelf wind stress component (dashed line) com-
FIG. 3. Coastal sea level response (m) at a coastline oriented east-west to a steady 8 m s\(^{-1}\) wind rotated through 360° for three different bottom slopes. The lines represent sea level output from the two-dimensional numerical model. The symbols represent sea level predicted by Eq. (6).

FIG. 4. Correlation between observed sea level and output from the two-dimensional numerical model as a function of modeled coastline orientation. Note that the highest correlation occurs when the modeled coastline orientation is 35° from north.

...puted from Eq. (6). The wind speed is 8 m s\(^{-1}\) and its direction is rotated through 360°. The sea level response due to solely along-shelf winds (i.e., a wind direction of 90° or 270° from north for an east-west coastline) is larger by a factor of 4 than that of solely across-shelf winds (0° or 180°). The combination of both responses (gray line) shows that the maximum response of sea level is due to a wind that is oriented 20° clockwise from the along-shelf direction.

Different bottom slopes (Fig. 3) support significantly different sea level responses. As the bottom slope decreases, the overall response increases because of a greater contribution of the across-shelf wind stress component, and the wind direction responsible for the maximum sea level response is oriented farther from the along-shelf direction. Comparison of Eq. (6) (lines in Fig. 3) with output from the two-dimensional numerical model (symbols) reveals good agreement between the two, confirming that the sea level response to wind forcing can be approximated well by a combination of Ekman transport and across-shelf wind stress.

...Directional results in Fig. 2 highlight the conceptual errors behind the usage of the phrase blowout tides that is common in the marine community. The general notion is that when the wind turns to the offshore direction it “blows” the water off the coast and lowers sea level. Instead, Fig. 2 shows that the strongest drop in sea level for a given wind speed is caused by wind directed nearly alongshore with just a small angle (about 20°) inclined offshore, basically coastal upwelling circulation. In contrast, for wind blowing directly offshore (180° in Fig. 2) the response of sea level is negative but very weak. The corresponding conceptual error is that onshore wind creates coastal flooding. Instead, Fig. 2 shows that maximum sea level elevation comes with the wind inclined only about 20° onshore, or downwelling circulation.

Examination of the coast near Atlantic City (Fig. 1) reveals that the coastline orientation varies between 25° to 50° clockwise from north. We can determine the optimum coastline orientation for our two-dimensional numerical model by examining a series of nearly identical numerical simulations whose only difference is the orientation of the coastline. The coastal sea level response of the simulations, which are all forced with the observed winds at Atlantic City for the time period 1997–98, is then compared with the observed coastal sea level at Atlantic City (Fig. 4) in terms of correlation coefficients. Maximum agreement between the simulated and observed sea levels (correlation \(r = 0.73\)) occurred for a simulated coastline that was oriented 35° clockwise from north. However, examination of Fig. 4 reveals that the predicted sea level is relatively insensitive to small variations in the orientation of the coastline. Coastline variations of \(\pm 15°\) result in a decrease in correlation from 0.73 to 0.67.

If we apply the results from Fig. 2 to Atlantic City, we find the greatest fall in sea level occurs for the wind from approximately 35° + 20° + 180° = 235° (from southwest), whereas there is no response for wind from 325° (northwest). The maximum rise is created by wind from 55° (northeast), whereas there is no response for wind from 145° (southeast). Janzen and Wong (2002) found similar results for sea level at Lewes, Delaware, where the coastline is oriented along 30°. The maximum positive response was from 60° (northeast), or 30° onshore. These directions are counterintuitive to many in the marine community who are not familiar with the action of Coriolis force and who would expect maximum sea level fall for northwest wind (blowout tides).
and maximum rise for southeast wind (onshore flow) on the New Jersey coast.

5. Statistical model

We can construct a statistical model of coastal sea level based on Eq. (6) that is a function solely of the along-shelf component of the wind velocity, the across-shelf component of the wind stress, and sea level atmospheric pressure. The frictional adjustment time on the shelf off Atlantic City (assuming an average coastal depth of 50 m) is approximately 1 day. Since we are interested in time scales of 1–2 days, we might expect the ocean response to lag the wind forcing, and so we include a lag time \( \xi \) in our statistical model:

\[
\eta_i(t) = a W_i(t - \xi) \sqrt{W_i(t - \xi)} + b W_i(t - \xi) W(t - \xi) + c [P - P(t)],
\]

where \( P \) is sea level atmospheric pressure, \( \bar{P} \) is its mean value (1013 hPa), and \( a, b, \) and \( c \) are regression coefficients whose units are seconds, seconds squared per meter, and inverse hectopascals, respectively.

However, the coastline orientation for use in computing the first two components in Eq. (9) is not immediately obvious without the use of a numerical model that may not be available for all coastal regions. Consequently, the sensitivity of the statistical model to the coastline orientation directly determines the efficacy of the model. Estimates of this sensitivity can be made by regression analysis for the time period 1997–98 to find the regression coefficients and lag time in Eq. (9) for a range of possible coastline orientations. Examination of the correlation (Fig. 5a) reveals a remarkable insensitivity to coastline orientation. The optimal lag time for all orientations is 11 h. As expected, the maximum correlation \( r = 0.82 \) occurs at a coastline orientation of 35°, agreeing with the two-dimensional model. However, the differences in correlation due to changes in the coastline orientation are extremely small. The correlation decreases from 0.82 to 0.80 over coastline changes up to 60°. The insensitivity of the statistical model to the coastline orientation is due to the inclusion of contributions from both an along-shelf component \((aW_i(t - \xi)\sqrt{|W_i(t - \xi)|}|W_i(t - \xi)|^{1/2})\) and an across-shelf component \((bW_i(t - \xi)W(t - \xi))\). The relative contributions of these two terms (see Fig. 5b) as a function of coastline orientation reveal that either can dominate the statistical model, depending on the chosen coastline orientation. However, the combined contributions of the wind effects and atmospheric pressure fluctuations (gray line) are relatively constant for all orientations. The insensitivity of the statistical model to the coastline orientation is fortunate. One does not require precise knowledge of this orientation to produce an accurate prediction of coastal sea level.

If we choose the coastline orientation (35°) that results in the highest correlation \( r = 0.82 \) and is consistent with the two-dimensional numerical model, Eq. (9) becomes

\[
\eta_i(t) = 0.025W_i(t - 11) \sqrt{W_i(t - 11)} - 0.00047W_i(t - 11)W(t - 11) + 0.0087[\bar{P} - P(t)],
\]

where \( t \) is in hours. The 95% confidence intervals for \( a, b, \) and \( c \) are 0.0003, 0.0003, and 0.0002, respectively. The lag time of 11 h agrees well with observations of winds and transport on the continental shelf (Garvine 1991), the two wind coefficients in Eq. (10) compare favorably to those in Eq. (8), and the pressure coefficient agrees with the inverse barometer effect.

As expected from Eq. (6), the wind coefficients are dependent on the shelf bathymetry of the coastal region. We perform another analysis for the sea level response during 1997 at Newport, Oregon, where the adjacent coastal region has a bottom slope that is much greater than that of Atlantic City (\( \alpha = 10^{-2} \) vs \( 10^{-3} \)). For a coastline orientation of 180°, regression analysis produces

\[
\eta_i(t) = 0.018W_i(t - 15) \sqrt{W_i(t - 15)} - 0.000013W_i(t - 15)W(t - 15) + 0.0087[\bar{P} - P(t)],
\]
where the pressure coefficient was held to the same value as in Eq. (10). Comparison between Eq. (11) and observations reveals very high correlation \( r = 0.84 \). However, the increased bottom slope and depth of the coastal region have profound effects on the relative contributions of the across- and along-shelf winds. The increased bottom slope results in a much smaller contribution of the across-shelf wind stress \((0.000\, 013 \text{ vs } 0.000\, 47)\), as Eq. (6) suggests. The greater depth increases the frictional adjustment time, which causes an even greater lag time (15 vs 11 h) between wind fluctuations and the sea level response.

6. Comparison with observations

We can now compare the statistical model described by Eq. (10) with the coastal sea level output from the two-dimensional numerical simulation presented in the earlier sections and output from the Extratropical Storm Surge Model. For this study, the two-dimensional numerical model is forced by Atlantic City observed winds, the statistical model by observed winds and pressure, and ESSM by the AVN forecast winds. Coastal sea level from ESSM is currently corrected for low-frequency variations due to gradients in water density or buoyancy (the steric effect) by continuously resetting the initial sea level from coastal observations (Chen and Shaffer 1999). This correction greatly increases the skill of the forecast. Following Chen and Shaffer (1999), we apply this correction to output from all three models. To force the initial sea level toward observations, an anomaly is calculated from the difference of the observed sea level and forecast sea level for a previous period. For a \( t_9 \)-h forecast, the anomaly is calculated as

\[
\eta_{\text{anomaly}}(t) = \frac{1}{T_{\text{ave}}} \int_{t_9-t_{0}}^{t_9-t_{0}+T_{\text{ave}}} \eta_{\text{observed}}(t') \, dt' - \frac{1}{T_{\text{ave}}} \int_{t_9-t_{0}+T_{\text{ave}}}^{t_9} \eta_{\text{forecast}}(t') \, dt',
\]

where \( T_{\text{ave}} \) is the averaging period. For the statistical and two-dimensional numerical model, an averaging period of 2 days provides the most accurate prediction. For ESSM, an averaging period of 5 days provides the most accurate prediction. This anomaly is then added to the forecast from the model to produce a corrected sea level:

\[
\eta_{\text{core}}(t) = \eta_{\text{forecast}}(t) + \eta_{\text{anomaly}}(t).
\]

The NWS provides 24-, 36-, and 48-h forecasts from the ESSM. With the availability of wind and pressure forecast products such as AVN (information online at http://www.nws.noaa.gov/mdl/synop/products.htm) or the European Centre for Medium-Range Weather Forecasts (online at http://www.ecmwf.int), we can produce similar predictions with the statistical model. A crude method for examining the ability of the model to forecast longer time scales is to treat the winds observed at Atlantic City as perfect “forecast winds” and to modify the anomaly calculation. This allows the isolation of the statistical model’s skill in longer-term forecasting without introducing complications due to uncertainty in truly forecast winds. Using this method, the two-dimensional model can also be evaluated using the forecast winds.

The corrected 24-h sea level predictions for ESSM, the statistical model, and the two-dimensional numerical model are all compared with the observed sea level for the winter of 1997/98 spanning from yearday 270 of 1997 to yearday 150 of 1998. A 65-day period from yearday 360 of 1997 to yearday 60 of 1998 is shown in Fig. 6, revealing that the 24-h forecasts from all three models agree quite well with observations. The correlations between the corrected model output and the observed coastal sea level for the winter of 1997/98 are 0.89, 0.86, and 0.82 for the ESSM, the statistical model, and the two-dimensional numerical model, respectively. The rms errors are 0.136, 0.115, and 0.127 m.

Correlation and rms errors for forecasts from 2 to 48 h are compared with the 24-, 36-, and 48-h forecasts from ESSM in Fig. 7. Again, all models produce accurate forecasts out to 48 h and considerably outperform persistence. Comparison of the statistical model with ESSM reveals that the former produces lower rms errors for all forecast lengths but slightly less correlation than the ESSM. The superior performance of the statistical model may reflect the usage of forecast AVN winds and pressure by ESSM instead of the observed winds and sea level pressure used by the statistical model. In all respects the statistical model outperforms the two-dimensional numerical model.

While all three models perform well, the 65-day period shown in Fig. 6 reveals several persistent tendencies.
of the models. The ESSM typically overresponds to wind events with elevations that are higher than the observed and depressions that are lower. This excessive response is especially clear in Figs. 8a and 8b, which show the greatest depression and greatest elevation, respectively, for the entire 2-yr record. In contrast, the statistical model closely tracks sea level fluctuations. These tendencies are reflected in a greater standard deviation for ESSM ($\sigma = 0.26$ m) than the statistical model ($\sigma = 0.21$ m) and the 2D model ($\sigma = 0.20$ m), whose values are very close to the observations ($\sigma = 0.21$ m). The failure of the statistical model and the two-dimensional model to predict the large sea level depression at yearday 364 (Fig. 8a) may reflect the influence of remote upshelf forcing that propagates downshelf as a coastally trapped wave. Such forcing is included in ESSM but not in our simple two-dimensional models, which lack any along-shelf variations.

7. Conclusions

We have addressed sea level response to synoptic-scale wind forcing. For the majority of the U.S. coastal waters, reliable sea level forecasts are available out to 48 h through the ESSM model. For underdeveloped and developing countries, however, infrastructure for such forecasting is not available. Hence, there is a need for a simple empirical forecast model that can make efficient use of local wind, pressure, and sea level observations and, perhaps, of forecast coastal winds. We developed and tested such an empirical model based on principles of coastal Ekman circulation. This model employs locally observed or forecast coastal winds and pressure and uses regression analysis where the subtidal frequency response is set by the along-shelf wind (not wind stress), the across-shelf wind stress, and the sea level atmospheric pressure. A lag in sea level of about 0.5 day from the wind forcing improves the model performance.

We found support for this empirical model formulation from results of a contemporary numerical ocean model, ECOM3d, applied to an idealized coastal setting, which omitted all along-shelf variations. Analysis of this model showed that both across-shelf and along-shelf wind components contributed to the sea level response, but typically the along-shelf component makes the larger contribution. For a given wind speed, the wind direction giving the largest drop in sea level was nearly along-shelf with the coast to the left but with a modest across-shelf inclination of about 20°. Similarly, the largest rise in sea level is produced by wind directed about 20° onshore. These results highlight the conceptual error behind the phrases blowout tides and onshore flow that are common in the marine community, which assume that the across-shelf wind is the primary forcing agent.

We tested the empirical model using sea level observations at Atlantic City. The storm of record there for coastal flooding was a northeaster that struck in March of 1962. Model performance was insensitive to the precise assignment of the local directions of the along-shelf and across-shelf wind components within about ±30°. For the sea level, pressure, and wind records of 1997–98 the National Weather Service’s Extratropical Storm Surge Model explained 79% of the total subtidal frequency sea level observed ($r = 0.89$) while our empirical model explained 74% ($r = 0.86$). The root-mean-square errors in sea level were 0.136 and 0.115 m, respectively.

This performance of the empirical model, simple as it is, seems adequate for general use and implies that
the shelf and coastline topography at Atlantic City form a regime in which the local wind forcing dominates. Although the model is based on barotropic dynamics and is less accurate during summer, the winds are typically weaker during summer, which results in only small wind-forced sea level variations. During periods of strong winds (and large wind-forced coastal sea level variations), the water column mixes, stratification decreases, and barotropic dynamics become applicable, increasing the accuracy of the model. Examinations of the empirical model at a different location, Newport, reveal similar performance, although the region’s greater bottom slope results in a much smaller contribution from the across-shelf winds and a greater lag. In some other applications, especially where along-shelf gradients in bathymetry are large (e.g., downshelf of the Monterey Bay, California, region) or stratification persists during strong winds (e.g., Peru), remote as well as local wind forcing is likely to be critical to reliable sea level forecasting.

Acknowledgments. We thank Alan Blumberg for providing the numerical ocean model ECOM3d and Jye Chen and Wilson Shaffer for providing the output from the ESSM. We also thank Michael Whitney and J. T. Reager for assistance in processing the observed winds and sea level. We greatly appreciate the comments on earlier drafts of this manuscript by K. Brink, D. Chapman, A. Clarke, C. Janzen, S. Lentz, S. Morey, A. Yanovsky, and two anonymous reviewers. This study was supported by the Sea Grant program of the National Oceanic and Atmospheric Administration under Grant R/F-10.

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