The Dynamical Coupling of a Wave Model and a Storm Surge Model through the Atmospheric Boundary Layer

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ABSTRACT

The effect of a wave-dependent drag coefficient on the generation of storm surges in the North Sea is studied. To this end, a barotropic storm surge model is driven by stresses that explicitly depend on the ocean waves. To estimate the effects of waves on the boundary layer the theory of Janssen is used. In this theory the aerodynamic drag depends on the fraction of the stress carried by the waves. For waves limited in growth by time, fetch, or depth this gives an enhancement of the drag. The importance to surges of radiation stress is also investigated.

The coupled wave and storm surge models have been tested for three recent storm periods. The calculations with a Smith and Banke stress relation underestimate the surges by 20%. The calculations with the wave-dependent drag give a significant improvement. When corrected for the effects of an external surge, the storm surge caused by a fast moving depression was overestimated slightly. In this case, the generation of the storm surge was dominated by relatively young waves. In the other two cases, the wave-dependent stress reproduced the overall level of the surges within a few percent. The radiation stress increased the surge some 5% during one storm, but the effect was negligible during the other two storms.

Most of the improvements can be reproduced by assuming an overall increase of the dimensionless roughness parameter. If the Smith and Banke relation is replaced by a Charnock one with a dimensionless constant of $\alpha = 0.032$, the difference in water level between this formulation and the wave-dependent calculation is smaller than the uncertainty in the observations. Different basins would, however, give rise to different choices of α. Therefore, a wave-dependent drag is to be preferred for storm surge modeling.

1. Introduction

The stress on the water that generates storm surges depends on the wind speed and roughness of the water. This roughness depends on the waves that are present on the water surface. Since waves are generated by the wind, the roughness of the water surface can be expressed to a fairly good approximation in terms of this wind. Charnock (1955) derived on dimensional grounds an implicit relation between the roughness and the wind. This leads to a roughly linear relation between the drag coefficient and the wind. This type of drag coefficient is commonly used in storm surge modeling.

Other variables enter the problem, however, if waves are not fully determined by the wind. If the growth of the waves is limited by fetch, time, or depth, these quantities will influence, through the waves, the interaction between air and sea. The drag coefficient cannot be fixed completely by the wind and depends explicitly on the sea state. In recent years, several authors have published experimental evidence of the wave state dependence of the drag coefficient. Geernaert (1990) shows that drag relations depend on the depth of the basin in which they were measured. Among others, Maat et al. (1991) and Janssen (1992) find that the roughness depends explicitly on the wave age.

Wolf et al. (1988) reported on a first attempt to couple dynamically a wave and a surge model. To calculate the drag from wave parameters they used the theory of Kitaigorodskii (1973). This theory gave values too high for the drag coefficients. They had to be divided by a factor 2.5 to give a reasonable mean stress.

In this paper, we will look at the effects of a wave-dependent drag on storm surges in the North Sea. A third-generation wave model will be used to calculate the waves and the wave dependence of the drag coefficient. To calculate the drag coefficient from the sea state the theory of Janssen (1991) will be used. The wave-dependent drag coefficient will be used to force a barotropic storm surge model. The results will be compared with two sets of reference runs for three recent storms. In the first set of reference runs we used the drag relation of Smith and Banke (1975). This relation was chosen because it is commonly used in storm surge calculations. As will be shown in section 5, calculations with the Smith and Banke relation systematically underestimate the surge. For this reason, a second set of reference runs was created with a wave-independent drag relation that did not have this flaw. This was done by using the Charnock relation and tuning the dimensionless constant in such a way that it

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reproduced the surge for one storm exactly. Comparing the surges calculated with this drag relation with the ones calculated with the coupled model will reveal the effects of the wave-enhanced drag more clearly.

That waves influence the apparent roughness of the sea surface is a consequence of the fact that waves have momentum. Since growing waves extract their momentum from the atmospheric boundary layer, they change the vertical distributions of turbulence and the wind profile. Ultimately this causes an enhancement of the apparent roughness of the sea. The impact on storm surges, in the North Sea, of this change in the apparent roughness of the sea surface due to waves is the main subject of this paper. Another consequence of waves carrying momentum is that propagating waves cause an excess momentum flux. In the balance equations for the total mean momentum this gives rise to a radiation stress term. Since this effect is so closely related to the main topic of this paper, its impact on storm surges has also been assessed by including the radiation stress term in the storm surge model.

There is both theoretical (e.g., Christoffersen and Jonsson 1985) and experimental (e.g., Gross et al. 1992) evidence that long waves in shallow water cause an enhancement of the bottom stress felt by the currents. The orbital motions of the long waves cause an alternating current near the bottom. This current causes a thin boundary layer (typically a few centimeters) in which the level of turbulence is increased. The apparent bottom roughness felt by the current is increased in this way. Several problems arise when one wants to account for the effect of waves on the bottom roughness. The roughness of the bottom is not a well-known parameter in a storm surge model. Usually a quadratic drag law is assumed, with a drag coefficient that is either constant or a weak function of depth. This parameterization implicitly assumes a roughness length that is proportional to the depth. The magnitude of the drag coefficient is usually estimated by tuning the amplitude of the tide in the model against time series based on harmonic analysis of tidal observations. It is believed by the authors that before the modification of the bottom roughness by waves can be taken into account, more accurate estimates of the bottom roughness, without waves present, have to become available. The assumption that the roughness is proportional to the depth is a very crude one.

In the next section, the theory of Janssen (1991) will be outlined. The governing equations of the wave model and the storm surge model will be discussed in section 3 and 4, respectively. In section 5, the results of the coupled model are compared with observations and with results of reference runs. A summary of the conclusions will be given in the last section.

2. Wave-dependent roughness

In this section the effect of waves on the roughness of the sea surface is discussed. First an expression for the drag will be presented in case the influence of waves is negligible. Then a simple assumption will be made regarding the effect of waves on the boundary layer. This leads to a wind profile that deviates from a logarithmic profile and to a wave-dependent drag. Finally, an outline will be given of the theory of Janssen (1991).

In all three cases to be discussed we will neglect the influence of viscosity. We also assume that a wind is blowing in the positive X direction, which is independent of time and the horizontal spatial coordinates. For such cases the Reynolds equation for the X component of the wind reduces to $\partial\tau/\partial z = 0$. In general, the total stress is the sum of a turbulent part and a wave-induced part: $\tau = \tau_t + \tau_w$. In all cases the turbulent stress is parameterized with a mixing-length hypothesis:

$$\tau_t = \rho_a(\kappa z)^2\left(\frac{\partial U}{\partial z}\right)^2, \qquad (1)$$

where $\kappa = 0.4$ is the von Kármán constant and $U(z)$ the wind speed at height z. The three approaches discussed below differ only in their assumptions for the wave-induced stress τ_w.

First we will neglect the influence of waves. Since viscous effects are ignored, the stress τ in the lower part of the boundary layer will be dominated by turbulence: $\tau = \tau_t$. Integration of this equation, with the help of (1), gives the well-known logarithmic wind profile:

$$U(z) = \frac{u_*}{\kappa}\ln\left(\frac{z}{z_0}\right), \qquad (2)$$

where $u_* = \sqrt{\tau/\rho_a}$ is the friction velocity and the integration constant z_0 is called the roughness length. On dimensional grounds, Charnock (1955) argued that this roughness length above the sea should be given by $z_0 = \alpha u_*^2/g$. If the dimensionless constant α is given, an implicit relation is found between the friction velocity and the wind at a certain height. From (2) one can see that the drag coefficient, defined by $C_D = u_*^2/U(10)^2$, is fully determined by the roughness length.

We will now make a simple approximation of the influence of waves on the boundary layer. Due to correlations between pressure and wave slope, wave-induced motions will contribute to the stress: $\tau = \tau_w + \tau_t$. Far away from the surface the wave stress τ_w goes to zero. Since the mixing length hypothesis (1) is still assumed to be valid for the turbulent stress, the velocity profile will be logarithmic if z is large enough for τ_w to be ignored:

$$U(z) = \frac{u_*}{\kappa}\ln\left(\frac{z}{z_e}\right), \quad z > \lambda, \qquad (3)$$

where λ is the height for which $\tau_w \ll \tau$. The value of the effective roughness z_e depends on the wave stress. From numerical simulations of the turbulent boundary layer, Makin and Chalikov (1986) found that the profile of the wave stress can be well approximated by a step function (Chalikov and Makin 1991). They assume $\tau_w(z) = \tau_w > 0$ for $z < \lambda$ and $\tau_w(z) = 0$ for $z \geqslant \lambda$. The wind profile for $z > \lambda$ is given by (3). For $z < \lambda$, another logarithmic wind profile is found:

$$U(z) = \frac{\sqrt{\tau - \tau_w}}{\sqrt{\rho_a}\kappa} \ln\left(\frac{z}{z_0}\right), \quad z < \lambda. \qquad (4)$$

Continuity of the wind profiles (3) and (4) at $z = \lambda$ demands that

$$z_e = z_0(\lambda/z_0)^{1-\sqrt{1-\tau_w/\tau}}. \qquad (5)$$

The height λ is equal to the significant waveheight: $\lambda = H_s$. For given H_s and z_0, the larger the fraction of the stress going into the waves, the larger the effective roughness z_e will be. The drag coefficient at height $z \gg \lambda$, which is roughly proportional to the logarithm of z_e, will be enhanced. This example illustrates qualitatively that the drag coefficient is sea state dependent. For young wind sea, when waves extract a large amount of momentum from the atmosphere, the drag coefficient will be enhanced.

The theory described in Janssen (1991) also starts with the assumption that the waves contribute to the stress close to the surface: $\tau = \tau_w + \tau_t$. Instead of assuming a profile for $\tau_w(z)$ and deriving from this a velocity profile, Janssen (1991a) follows this line of reasoning the other way around. In agreement with his numerical results (Janssen 1989), he assumes the wind profile is given by

$$U(z) = \frac{u_*}{\kappa} \ln\left(\frac{z + z_e - z_0}{z_e}\right). \qquad (6)$$

From this it is possible to derive the profile of the wave stress. From Fig. 1 it can be seen that this wind profile shows the same characteristics as the one derived with the step-function approximation for the wave stress. For the turbulent part of the stress we again assume the mixing length hypothesis (1) to be valid. Since the drag coefficient depends on the effective roughness z_e, we would like to derive an expression for z_e. If the wind profile (6) is differentiated, squared, and compared with the expression for the turbulent stress (1), we find for $z = z_0$:

$$z_e = \frac{z_0}{\sqrt{1 - \tau_w/\tau}}, \qquad (7)$$

where $\tau_w = \tau_w(z_0)$. Comparing both approaches we see that Janssen used the length scale $z_e - z_0$ to parameterize the height dependence of τ_w. This leaves only one variable to be parameterized: the roughness length z_0. For this Janssen assumes a Charnock-like relation $z_0 = \tilde{\alpha} u_*^2/g$ is valid. The value for $\tilde{\alpha}$ is tuned in such a way that $z_e = \tilde{\alpha} u_*^2/g$ for old wind sea. In this way a model is constructed that makes the stress behave according to Charnock if τ_w/τ is small but enhances the stress if τ_w/τ approaches one.

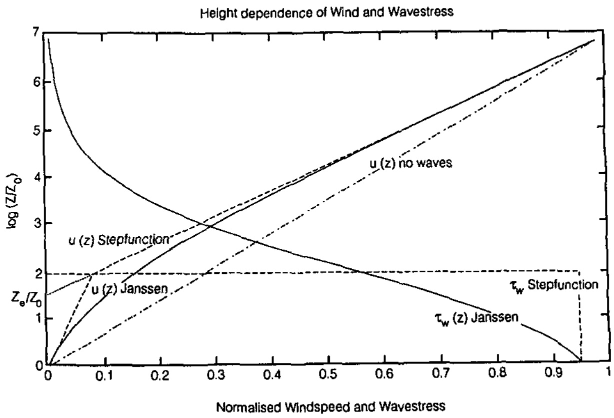

FIG. 1. Profiles of wind speed and wave stress. If the influence of waves is neglected, a logarithmic wind profile is assumed. Representing the wave stress by a step function gives rise to a wind profile with an effective roughness z_e larger than z_0. The wind profile according to Janssen (1991a) with the same effective roughness is also shown, as is the corresponding wave stress profile.

In the last two theories discussed above, the wave dependence of the drag coefficient follows from assumptions concerning the wave-induced stress in the lower part of the boundary layer. In this way the wind profile is consistent with the surface stress and the wave stress. In the past, several authors have followed a less fundamental and more straightforward approach. They parameterized the dimensionless roughness length or the drag coefficient in terms of wave parameters like the steepness and wave age. Several examples of this approach can be found in the review article by Geernaert (1990). Unlike the quasi-linear approach of Janssen (1991), however, these theories do not affect the growth of waves.

3. The wave model

To calculate the wave state on the North Sea, a regional version [see Burgers (1990) and Fig. 2] of the WAM wave model has been used. This model solves the energy balance equation for waves:

$$\frac{\partial F(f, \theta)}{\partial t} + \nabla \cdot (\mathbf{c}_g F(f, \theta)) = S_{in} + S_{dis} + S_{bot} + S_{nl}, \qquad (8)$$

where $F(f, \theta)$ is the wave spectrum (in m^2 s), f is the frequency of a wave component, and θ its direction. In this study, we disregard the influence of currents on waves, as can be seen from the energy balance equation (8). Each wave component travels with its *group velocity:*

$$\mathbf{c}_g = \frac{\partial \omega}{\partial \mathbf{k}}, \qquad (9)$$

with $\omega = 2\pi f$. On the right-hand side of (8), four so-called *source functions* represent the wind input, the dissipation due to whitecapping, the bottom dissipa-

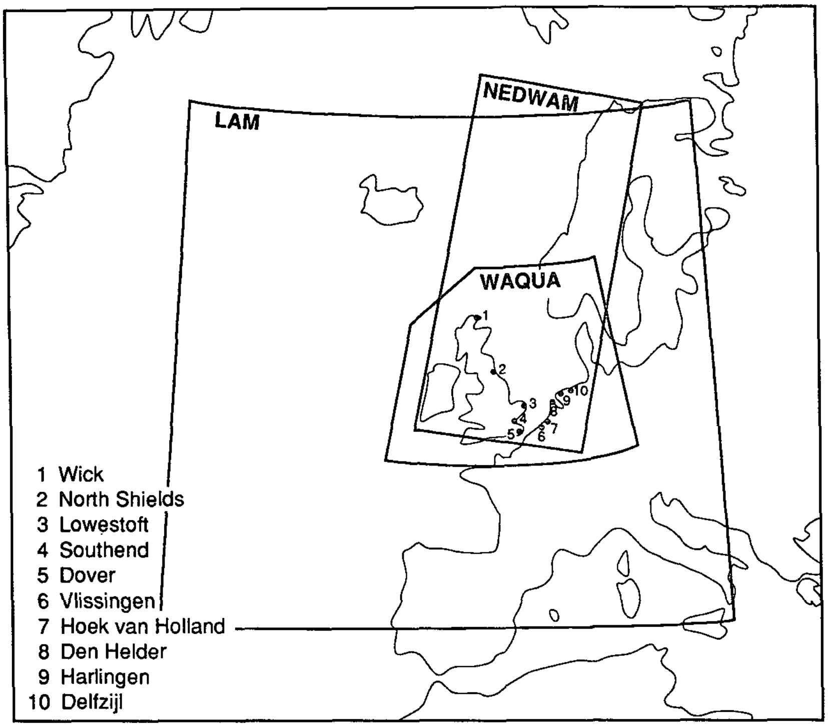

FIG. 2. Map of the three models involved: the meteorological model LAM, the wave model NEDWAM, and the storm surge model WAQUA.

tion, and the nonlinear wave–wave interactions, respectively. Two source functions, S_{in} and S_{dis}, were changed with respect to WAMDI Group (1988) to fit the scheme of Janssen. The approximations used to calculate the nonlinear energy transfer S_{nl} are described by Hasselmann et al. (1985). The bottom dissipation is given by

$$S_{bot}(f, \theta) = \frac{0.076}{g} \frac{k}{\sinh(2kD)} F(f, \theta), \quad (10)$$

where $k = |\mathbf{k}|$ is the wavenumber of a wave component. This formulation is equivalent to the one proposed by JONSWAP Group (1973). The dissipation due to whitecapping is parameterized in the spirit of Hasselmann (1974) but tuned to give reasonable wave heights combined with the input source term discussed below:

$$S_{dis}(f, \theta) = -2.25\bar{\omega}(E\bar{k}^2)^2\left(\frac{k}{\bar{k}} + \frac{k^2}{\bar{k}^2}\right)F(f, \theta), \quad (11)$$

where E is the wave variance in meters squared, $\bar{\omega}$ is the mean angular frequency, and $\bar{k}$ is the mean wavenumber. Following the scaling arguments of Miles (1957), the input source term can be written as

$$S_{in}(f, \theta) = \omega\epsilon\beta\left(\frac{u_*}{c}\right)^2 \cos^2(\theta - \phi)F(f, \theta), \quad (12)$$

where $\epsilon = \rho_a/\rho_w = 1.25 \times 10^{-3}$, $c = \omega/k$ is the phase speed of a wave component, and ϕ is the direction of the wind. The friction velocity u_* is defined as $u_* = \sqrt{\tau/\rho_a}$. In the theory of Janssen (1991), the Miles parameter β is a function of two dimensionless parameters: the wave age u_*/c and the wind profile parameter gz_0/u_*^2. To determine the input source function, information about the surface roughness length z_0 is required. This surface roughness will depend on the input source function, making the resulting set of equations implicit. First, the expressions for the input source function will be given. Using the fact that observed wind profiles are described very well by (6) and fitting the input source term to observations (Janssen 1991a), the Miles parameter can be written as

$$\beta = \frac{1.2}{\kappa^2} \mu \ln^4\mu, \quad \mu \leq 1. \quad (13)$$

This parameterization of the Miles parameter is in agreement with numerical calculations based on the theory of Miles. The dimensionless critical height μ is given by

$$\mu = \frac{gz_e}{c^2} e^{\kappa c/|u_* \cos(\theta - \phi)|}. \quad (14)$$

The effective roughness z_e and friction velocity u_* follow from the theory of Janssen (1991) described in the Introduction. In this theory, the wave stress $\tau_w = \tau_w(z_0)$ is required. Using conservation of momentum, this wave stress equals the amount of momentum going into the waves due to wind:

$$\boldsymbol{\tau}_\omega(z_0) = \rho_\omega \int_0^\infty df \int_0^{2\pi} d\theta \omega S_{in}(f, \theta) \frac{\mathbf{k}}{k}. \quad (15)$$

Now the set of equations for u_* and z_e is closed. Usually the U_{10} and the wave state $F(f, \theta)$ are given, and the problem is to find a proper u_*. First, the wave stress τ_w is calculated with the integration (15), using u_* and z_e from the previous time step to determine the input source function S_{in}. Then τ, z_e, and z_0 are found by iteration of Eqs. (6) (for $z = 10$ m), (7), and the Charnock-like relation $z_0 = \tilde{\alpha} u_*^2/g$. For old wind sea ($\tau_w/\tau \approx 0.5$) we want the coupled model to give the same drag coefficient as a Charnock theory without wave effects. This leads to $\tilde{\alpha} = 0.01$.

The geographic grid of NEDWAM contains 612 grid points, with a grid spacing of approximately 75 km. The wave spectrum is defined on a wave–vector grid of 25 frequencies by 12 directions. The lowest frequency is given by $f_1 = 0.0418$ Hz, every subsequent frequency is 10% higher than the previous one. The highest frequency is $f_{25} = 0.4117$ Hz. Since the computation of S_{nl} and τ_w requires information about frequencies beyond f_{25}, the spectrum for frequencies $f > f_{25}$ is assumed to be proportional to f^{-5}:

$$F(f_m, \theta) = \left(\frac{f_m}{f_{25}}\right)^{-5} F(f_{25}, \theta), \quad m > 25. \quad (16)$$

4. The storm surge model

The motions of tides and surges are assumed to be described by the depth-averaged Reynolds equations. The averaging procedure is performed in such a way that the mass flux by the mean velocity **u** is the sum of the mass flux by the Eulerian mean velocity and the Stokes drift caused by wave motions (Hasselmann 1971; Battjes 1974). In Cartesian coordinates the equations are given by

$$\frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} - \gamma v = -g\frac{\partial \zeta}{\partial x} - \frac{1}{\rho}\frac{\partial p_a}{\partial x} + \frac{1}{\rho h}\left(\tau^x - \tau_b^x - \frac{\partial S_{xx}}{\partial x} - \frac{\partial S_{xy}}{\partial y}\right), \quad (17)$$

$$\frac{\partial v}{\partial t} + u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + \gamma u = -g\frac{\partial \zeta}{\partial y} - \frac{1}{\rho}\frac{\partial p_a}{\partial y} + \frac{1}{\rho h}\left(\tau^y - \tau_b^y - \frac{\partial S_{yx}}{\partial x} - \frac{\partial S_{yy}}{\partial y}\right), \quad (18)$$

where u, v are components of the depth-averaged velocity **u**; t is time; γ the Coriolis frequency; g the gravitational acceleration; ζ represents the sea surface elevation; ρ is the density of water; p_a the atmospheric surface pressure; τ^x, τ^y are components of the surface stress $\boldsymbol{\tau}$; τ_b^x, τ_b^y are the components of the bottom stress $\boldsymbol{\tau}_b$; and $h = D + \zeta$ is the instantaneous water depth. The radiation stress $S_{ij}(i, j = x, y)$ represents the contribution of the wave motions to the mean horizontal flux of horizontal momentum. Later in this section it will be expressed in terms of the wave spectrum F. Integration of the continuity equation over the depth combined with the kinematic boundary conditions at the surface and the bottom gives the third equation to be solved:

$$\frac{\partial \zeta}{\partial t} + \frac{\partial (hu)}{\partial x} + \frac{\partial (hv)}{\partial y} = 0. \quad (19)$$

The storm surge model WAQUA/CSM-16 (see Verboom et al. 1992) solves this coupled set of partial differential equations to obtain u, v, and ζ on every grid point. The wind stress and atmospheric surface pressure are supplied by a meteorological model. Usually, the meteorological model supplies wind speeds at a height of 10 m. From this $\mathbf{U}_{10}$ the surface stress is calculated by

$$\boldsymbol{\tau} = \rho_a C_D |\mathbf{U}_{10}| \mathbf{U}_{10}, \quad (20)$$

where C_D is parameterized according to Smith and Banke (1975): $C_D = (0.066|\mathbf{U}_{10}| + 0.63) \times 10^{-3}$. This parameterization will be replaced by a more elaborate scheme that takes waves into account. The bottom stress in (17) and (18) is parameterized by

$$\boldsymbol{\tau}_b = \rho C_b |\mathbf{u}| \mathbf{u}. \quad (21)$$

In theory, the bottom drag coefficient C_b depends on the roughness of the bottom, the depth, and the velocity profile. By tuning the model, the following parameterization was found to reproduce optimally the tidal elevations in the southern part of the North Sea:

$$C_b = \begin{cases} 2.55 \times 10^{-3}, & D \leqslant 42 \text{ m} \\ 9.81(D + 20)^{-2}, & 42 < D \leqslant 66 \text{ m} \\ 1.33 \times 10^{-3}, & D > 66 \text{ m}, \end{cases} \quad (22)$$

with D given in meters.

The storm surge model WAQUA/CSM-16 solves Eqs. (17), (18), and (19) for u, v, and ζ on a staggered C grid using an alternating direction implicit scheme with a time step of 10 minutes. The model covers the whole northwest European continental shelf with a grid spacing of $\frac{1}{4}°$ in the W–E direction and $\frac{1}{6}°$ in the N–S direction. This results in a grid spacing of approximately 16 km and 4213 active points. The water levels on the open boundaries, located in deep (>200 m) water, are prescribed by ten tidal constituents.

The radiation stress can be expressed in terms of the wave spectrum F as calculated by the wave model (Battjes 1974):

$$S_{ij} = \rho g \int_0^{2\pi} \int_0^\infty \left\{ \frac{c_g}{c}\frac{k_i k_j}{k^2} + \left(\frac{c_g}{c} - \frac{1}{2}\right)\delta_{ij} \right\} \times F(f, \theta) df d\theta. \quad (23)$$

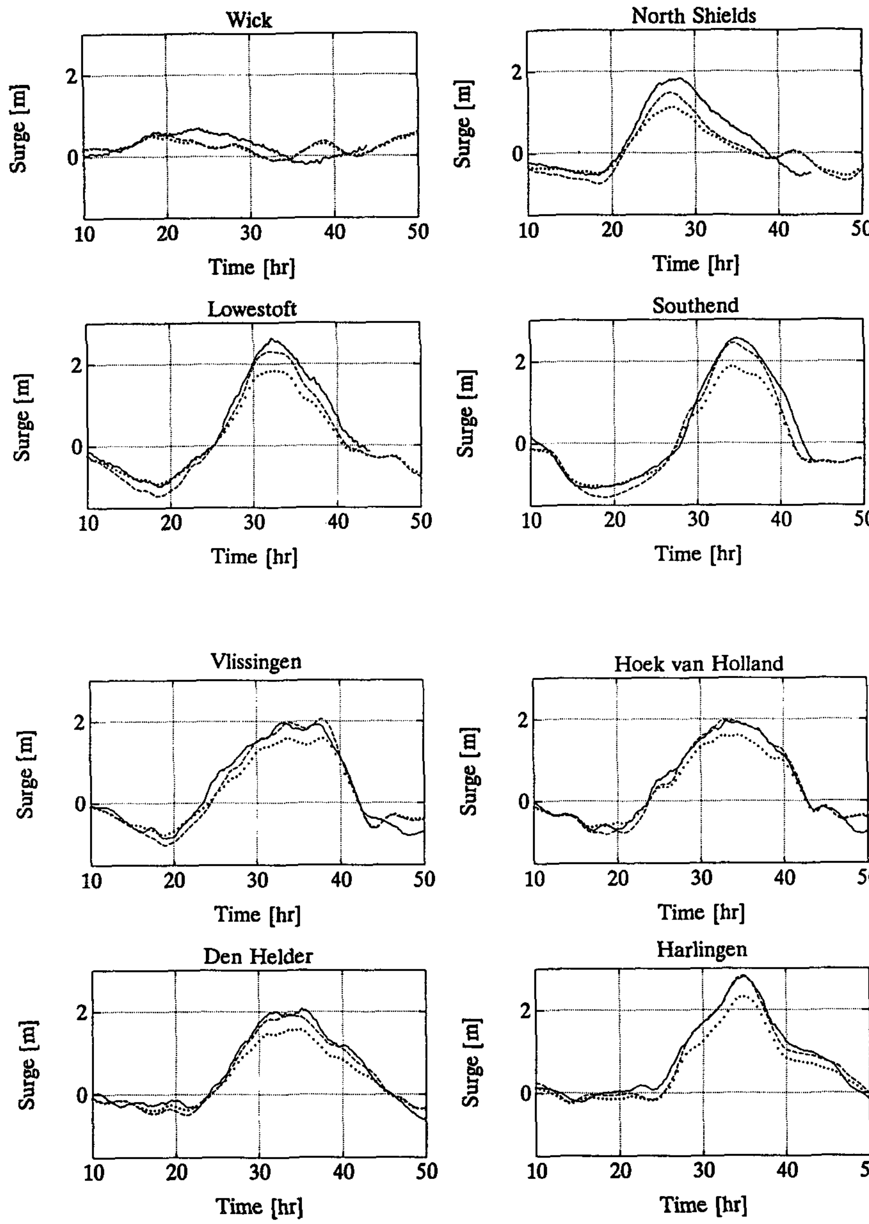

FIG. 3. Residual elevations at several stations along the English and Dutch coasts for the February 1989 storm. The full line represents the observations, the dotted line corresponds to the calculation with the Smith and Banke drag relation, and the dashed line shows the results of the calculation with the wave-dependent drag coefficient. The times on the x axis are given with respect to 0000 UTC 13 February.

In deep water, when $c_g/c = 1/2$ applies, this expression simplifies to

$$S_{ij} = \frac{1}{2}\rho g \int_0^{2\pi}\int_0^{\infty} \frac{k_i k_j}{k^2} F(f, \theta) df d\theta. \qquad (24)$$

In the southern part of the North Sea, with a typical depth of 40 m, the deep-water assumption breaks down for long swell waves. In this case the full expression (23) should be used. Since we do not expect the radiation stress to be the leading forcing term to drive the storm surge model, we will assume that the deep-water assumption applies everywhere. The integrals (24) are evaluated at every propagation time step of the wave model, that is, each 20 minutes, and the results are accumulated. Every 3 hours a mean radiation stress is calculated, which is subtracted from the wind stress. The resulting stress is used to force the storm surge model.

In the original version of the storm surge model, WAQUA/CSM-16 (without wave effects), the winds at 10 m and the surface pressures are interpolated from the grid of the meteorological model LAM (resolution 55 km) to the WAQUA/CSM-16 grid. On the surge model grid the stresses are calculated using the relation proposed by Smith and Banke (1975). In the coupled setup the 10-m winds from the meteorological model are interpolated to the NEDWAM grid with an interval of 3 h. Using the scheme described in the previous section, the friction velocity u_* is calculated. Every 3 h, the drag coefficient $C_d = U_{10}^2/u_*^2$ is interpolated back to the meteorological grid. In regions where the wave model and the meteorological model overlap, this wave-dependent drag coefficient is used to calculate the stress. Outside these overlap regions the Smith and Banke relation is used. The resulting stress is interpolated to the surge model grid. So in the original version, wind velocities are interpolated from the meteorological grid to the surge model grid, whereas in the coupled version, stresses are interpolated. To estimate the influence of this difference, we performed a calculation where the winds were transformed in stresses using the Smith and Banke relation on the meteorological grid, which were interpolated to the surge-model grid. Compared to the original calculation, this resulted in surges that were 1%–2% higher. Compared to other sources of errors like the winds from the meteorological model and the water levels at the open boundaries, the error due to interpolation can be neglected.

5. Results

In the first part of this section we will concentrate on what turns out to be the main effect of waves on surges: the wave-dependent drag. In the model calculations discussed in this part, the effects of the radiation stress and the Stokes drift will be neglected. This assumption will be justified in the latter part of this section.

The surges calculated with a wave-dependent drag coefficient were compared with observations for three recent storms: 14 February 1989, 27 February 1990, and 13 December 1990. These three storms have different characteristics. The depression causing the storm of February 1989 moved rapidly to the east over the northern part of the North Sea. Due to its pace this storm can be associated with relatively young waves. By contrast, the depression of December 1990 moved to the south slowly. During this storm waves became fully developed under high wind conditions. The storm surge of February 1990 was generated by a chain of depressions. These depressions did not allow the winds on the North Sea to drop under 20 m s^{-1} for several days. This resulted in a surge that lasted more than 4 days in the Southern Bight. So this February 1990 is an example of a storm where we can expect relatively old waves to be important.

In Fig. 3, the observed and measured surges of eight stations are shown for the storm of February 1989. In all cases except Wick, the reference run with the Smith and Banke drag relation underestimates the surge substantially. The wave-dependent drag parameterization of Janssen seems to perform well for this storm. This is also reflected in the estimates of the model of the highest water level that occurred at each station. As can be seen in Table 1, the model estimates of the highest water level reached during the storm are within 10 cm for stations located along the Dutch coast. When using the Smith and Banke drag relation, the model underestimates the highest water level 30 cm or more, depending on the location.

In the case of a storm that moved a little slower, December 1990, the wave-dependent parameterization also performs well. The time series of this surge for the Dutch stations are shown in Fig. 4. From Table 1 the highest water level reached is estimated with an accuracy of about 20 cm. This is less accurate than in the case of the February 1989 storm but still an improvement of about a factor 2 compared with the model using the Smith and Banke drag relation. That the parameterization of Janssen does not just enhance the drag but also takes the wave state into account can be seen if the drag coefficient is looked at directly. In Fig. 5, snapshots of the drag coefficient as a function of the wind speed at 10 m are shown at two different moments during the storm. The upper picture shows the situation

TABLE 1. Observed and calculated surges along the English and Dutch coasts for the storms of February 1989 and December 1990. The locations of the stations are marked in Fig. 2. The surge, defined as the difference between the highest water level that occurs and the highest water level expected due to the tide, is given in meters.

	14 February 1989				12 December 1990			
Station	Obs	Sm&B	Wave	Charn	Obs	Sm&B	Wave	Charn
Wick	0.45	0.39	0.41	0.41	0.44	0.31	0.35	0.35
North Shields	0:18	0.01	0.00	0.01	0.86	0.60	0.81	0.73
Lowestoft	1.54	1.01	1.41	1.31	1.12	1.06	1.42	1.28
Southend	1.38	0.73	0.97	0.93		0.83	1.14	1.00
Dover	0.53	0.30	0.55	0.50	0.71	0.91	1.20	1.11
Vlissingen	1.59	1.38	1.63	1.62	1.43	1.17	1.54	1.40
Hoek van Holland	1.80	1.54	1.91	1.84	1.63	1.16	1.49	1.40
Den Helder	2.03	1.56	1.90	1.86	1.85	1.40	1.67	1.70
Harlingen	2.25	1.74	2.08	2.10	2.29	1.82	2.12	2.21
Delfzijl	1.99	1.55	1.77	1.84	2.44	1.47	1.74	1.75

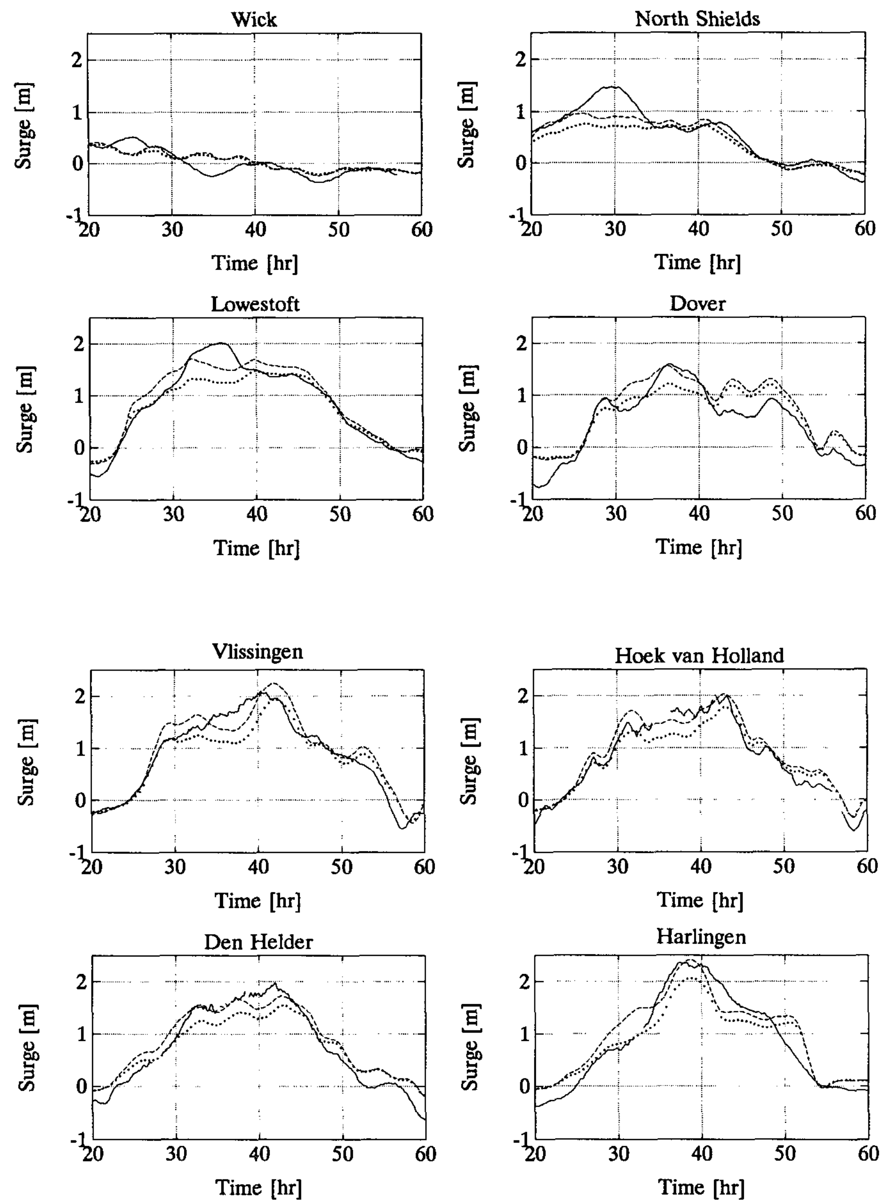

FIG. 4. Residual elevations at stations along the English and Dutch coasts for the December 1990 storm. The different line styles are the same as in the previous figure. The times are given with respect to 0000 UTC 11 December 1990.

at the beginning of the storm. Young wind sea with a wave age $c_p/u_* < 10$ gives rise to drag coefficients of up to $5\ 10^{-3}$, two times the value of Smith and Banke at the same wind speed. The picture below shows the situation 9 h later. The wind has decreased only a few percent, but the drag coefficient has been reduced to slightly more than the Smith and Banke values. This is consistent with the fact that waves with ages $c_p/u_* < 10$ have disappeared.

In the third case, February 1990, the coupled model also gives an improvement compared to the reference run; however, for certain stations, Den Helder and Harlingen, the surge is underestimated by about 0.5 m. This is probably due to poor wind and pressure data. The meteorological situation is more complex than during a storm caused by a single depression. This may introduce spatial and temporal variations in the wind field that are not well resolved in the meteorological model. The time series of the water level observations indeed show a lot of variations on short time scales. Also, the fact that the mismatch between the calculations and the observations varies considerably for the different Dutch stations (see Table 2) indicates that small-scale processes are important for this storm.

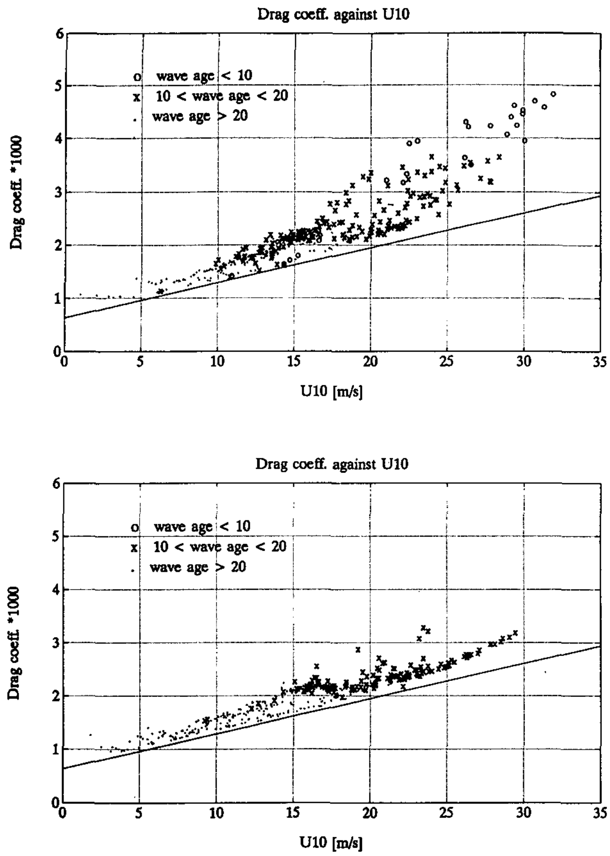

FIG. 5. Scatterplot of the drag coefficient against the wind speed at 10 m. The first plot shows a snapshot of the drag coefficients on all grid points of the wave model at 0000 UTC 12 December 1990; the second 9 h later. In the first plot, many young waves are present. This results in drag coefficients up to almost 0.005. Nine hours later no young waves are present, and the maximum drag coefficient is reduced to less than 0.0035. Wave age is defined as the phase velocity of the peak frequency divided by the friction velocity: c_p/u_*.

It is clear from the results presented so far that the calculations with a Smith and Banke stress relation underestimate the surges substantially. The calculations with the wave-dependent drag do not show this behavior. This leads to the important conclusion that the theory of Janssen gives a good description of the flow of momentum from the atmosphere to the water. It is clear that much of the improvement of the wave-dependent drag can be reproduced by choosing a larger drag coefficient than the Smith and Banke relation. This can be done easily by adopting the Charnock relation and determining the dimensionless roughness length α by tuning the calculated surges to observations. We expect that in a case where the sea is dominated by young wind sea, like the February 1989 storm, the stresses calculated with the wave model are relatively high. If α is tuned to reproduce the elevations of such an event, it can be expected to give elevations too high in cases that are not dominated by young sea, like the December 1990 storm. The calculations for the February 1989 storm with the Charnock relation gave best results for the dimensionless roughness length $\alpha = 0.032$. The elevations calculated with this parameter are shown in Table 1. Also in Table 1 are the results with this tuned relation for the December 1990 case. For the English stations the surges are lower than the surges calculated with the wave-dependent drag. When the surge reaches the Dutch stations this difference has disappeared. In Fig. 6, the surge in Hoek van Holland is plotted as a function of time. In the beginning of the storm the wave-dependent stress causes an elevation 20 cm higher due to the young wind sea. Some 15 h later the sea state is dominated by old waves, which reduces the water level compared to the Charnock calculation. From this figure it is also clear that the difference between the calculations is of the same order as the scatter due to small-scale processes in the observations. So the influence of drag enhancement by young waves can be seen in the calculations, with a typical magnitude of 5% of the total surge. Since this difference is of the same order as the overall error of the model and the scatter in the observations, the observations do not provide evidence in favor or against the drag enhancement by growing waves.

As can be seen in the surge plot for February 1989 (Fig. 3), both the coupled model and the model with the Smith and Banke stress relation underestimate the

TABLE 2. Observed and calculated surges along the Dutch coast of two consecutive high waters during the storm period of February 1990. The surges, defined as in the previous table, are given in meters.

	26–27 February 1990				27–28 February 1990			
Station name	Obs	Sm&B	Wave	Charn	Obs	Sm&B	Wave	Charn
Wick	0.50	0.47	0.47	0.47	0.66	0.65	0.71	0.69
North Shields	0.62	0.50	0.54	0.53	0.53	0.46	0.51	0.51
Lowestoft	1.01	0.92	1.09	1.05	0.59	0.53	0.57	0.62
Southend	0.51	0.29	0.42	0.41		0.08	0.08	0.20
Dover	0.69	0.76	0.89	0.86	0.63	0.52	0.58	0.61
Vlissingen	1.08	1.03	1.21	1.23	1.21	1.02	1.13	1.20
Hoek van Holland	1.29	1.05	1.24	1.24	1.45	1.05	1.18	1.22
Den Helder	1.51	0.91	1.09	1.08	1.70	1.07	1.20	1.26
Harlingen	1.88	1.29	1.55	1.58	2.09	1.53	1.74	1.87
Delfzijl	1.92	1.50	1.63	1.71	2.23	1.67	1.79	1.92

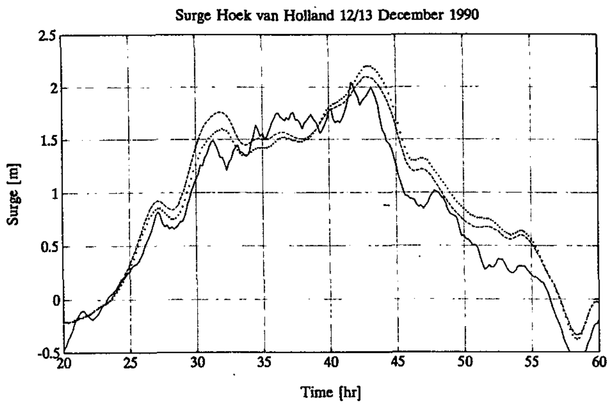

FIG. 6. Residual water level at Hoek van Holland during the December 1990 storm. The full line represents observations, the dashed line the calculation with the wave-dependent drag coefficient, and the dotted line the calculation with the Charnock relation. The wave effect gives the surge an enhancement in the beginning of the storm. Later this enhancement disappears and the Charnock relation, tuned to reproduce the average elevation, produces higher elevations.

surge at Wick with some 20 cm for more than 10 h. This underestimation propagates with the surge wave along the English coast. In the case of the coupled calculation, it disappears in the southern part of the North Sea. This phenomena could be caused by an external surge that entered the model area from the north. If this is the case, the coupled model might overestimate the stress since it is able to completely compensate the surge lack as the surge propagates along the coast. To investigate the influence of the possible external surge on the conclusions stated above, we used a version of the surge model that can adjust the surge wave by assimilating water level measurements. This technique is based on Kalman filtering (Heemink 1990). To make the model adjust to the external surge, we assimilated the measurements at Wick. As can be seen in Fig. 7, the coupled model overestimates the surge in the southern North Sea by some 10% in the case of the February 1989 storm if the possibility of an external surge is taken into account. During the December 1990 storm the external surge was insignificant (see Fig. 7); in the February 1990 case it was completely absent. This leaves the conclusion for these two storms unaltered.

In most cases the radiation stress is negligible compared to the wind stress. In some particular cases, when the swell generated by the storms reaches the shallower parts of the North Sea (the Doggerbank and the German Bight), the radiation stress becomes as large as 0.40 N m^{-2}. Usually it is only a fraction of this value. This results in a small impact on calculated water levels: in the case of the two single-depression storms, February 1989 and December 1990, the impact is less than 5 cm. The water levels during the February 1990 storm showed an increase of 10 to 15 cm when the radiation stress was included in the calculation. So clearly, the effect of radiation stress cannot be neglected in all cases. Due to the poor quality of the meteorological forcing in the case of the complex multiple-depression storm of February 1990, which introduces an error of approximately 30 cm at some stations, it is hard to say if the radiation stress improved the performance of the storm surge model in this case.

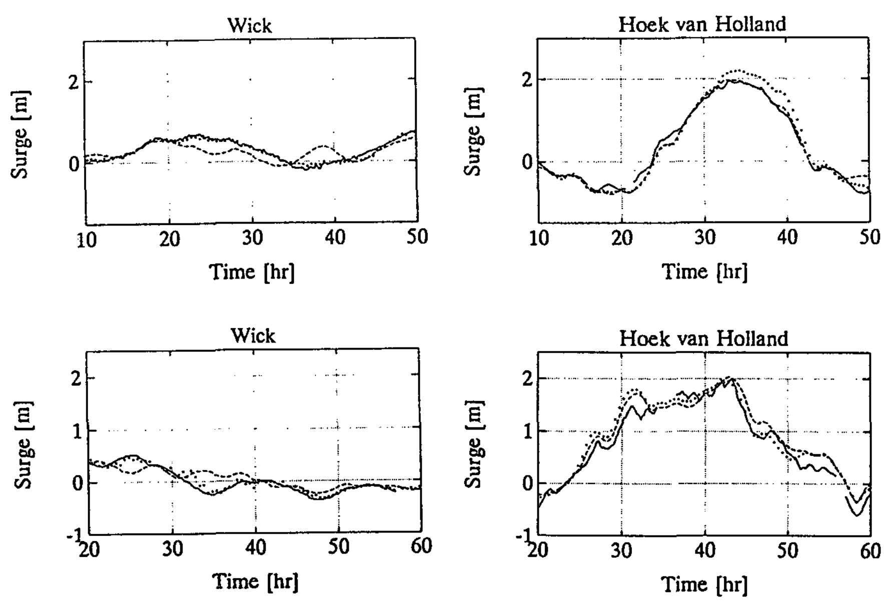

FIG. 7. Residual elevations at Wick and Hoek van Holland for the February 1989 (above) and the December 1990 storms. The full line represents the observed surge. The dashed and the dotted lines represent coupled calculations without and with assimilations of observations at Wick, respectively.

6. Conclusions

The main result of the research described in this paper is that the description of the momentum transfer to waves and currents by Janssen (1991) is consistent with the storm surge elevations observed in the North Sea. This result is obtained without tuning to wave or surge observations in the North Sea.

Comparing the result of the coupled model with a model using a standard Smith and Banke stress relation for three selected storms, we find that the results improve considerably using the coupled model. This improvement depends on the type of storm that caused the surge. In the case of a fast moving depression (February 1989), where the air–sea interaction is dominated by young waves, the coupled model performs best. When the effect of a surge generated outside the model is taken into account, the coupled model slightly overestimates the surge in this case. For a storm that moved more slowly (December 1990) the improvement is also considerable. In the case of a storm caused by a chain of depressions (February 1990) the improvement is less impressive. Since the weather pattern is more complex in the case of multiple depressions, this may be due to a lack of resolution in both the meteorological and the wave model. The storm of February 1990 was the only event where the radiation stress significantly affected the calculated surge. The effect of radiation stress was negligible in the case of the two single depression storms, February 1989 and December 1990.

Most of the improvements of the coupled model can be reproduced by increasing the drag coefficient with respect to the Smith and Banke parameterization. By tuning the roughness parameter of the Charnock relation for the February 1989 storm, we find that the value $\alpha = 0.032$ gives results comparable to the coupled model. Compared to the calculation with the Charnock relation, the wave-dependent drag enhances the water-level rise in the beginning of the surge, causing a difference of about 5% of the total surge. This difference remains within the uncertainty range of the observations. It is important, however, to realize that the tuned Charnock parameter α is not universal. For ocean conditions it will certainly be too large, and for smaller and shallower basins even larger α are appropriate. For this reason the use of a wave-dependent drag for storm surge modeling is to be preferred.

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