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Operational storm surge modelling in the Western Black Sea: one way coupling with a wave model

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Abstract. We present the progress in the operational modelling of storm surges along the Western Black Sea coast with a focus on the Bulgarian coast. A one way coupling of such model with an operational wave model is implemented. The wave model provides information about the wave induced stress on the sea surface and the wind drag coefficient. The validation of the system of these two models during several storms with strong winds and surges above 1m shows that the influence of the wave induced stress is not significant because of the spatial resolution of the storm surge model. The influence of the sea state on the wind drag coefficient is significant and leads to improvement: prevention of overestimation of the surge for severe storm cases. Two alternative approaches to parameterize the wind drag coefficient dependency of the wave parameters are tested: direct usage of the friction velocity from the wave spectra and the use of wave steepness output from the wave model. The use of the wave steepness information in the storm surge model leads to improvements of the simulation results, while the usage of the friction velocity leads to further overestimation of the surges. The approach with the wave steepness was implemented in the storm surge forecasting system of NIMH-BAS.

Keywords: storm surge, Black Sea, wind drag coefficient, waves

1. INTRODUCTION

Storm surges in the Black Sea are events that happen with much lower frequency and amplitude than in the ocean, however during the last 50 years there are more than 20 events with sea level rise more than 80 cm for the Bulgarian coast and about 10 cases with peak level rise of more than 1m and the most extreme recorded one is the storm surge during the storm of February 1979 when the sea level rise reached 1.5m at some locations around the Bulgarian coast and caused very significant damages. Other

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notable storms are the storms of 1976, 1977, 1981, 1996, 1997 1998, 2006, 2010, 2012 and others. The first numerical prediction of storm surges by a hydrodynamic model and a prediction by statistical modeling was implemented by Mungov back in the 80's. After 1990 the storm surge model of Meteo France (Mungov and Daniel, 2000) was adapted to the Black Sea in the frame of the bilateral cooperation between Meteo France and NIMH. It was found that the model generally underestimates the storm surges, fact explained with the low spatial and temporal resolution of the atmospheric forcing. Another approach to the storm surge prediction by using neural networks technique was tested by Pashova and Popova (2011). The present article describes an approach to implement a one way coupling of the storm surge model with the operational wave model of NIMH (National Institute of Meteorology and Hydrology) by exchange of information about the sea state and wave radiation stress from the wave model to the storm surge model. As it was found that the influence of the wave radiation stress on the storm surge prediction is negligible, the study is focused mainly on the influence of the sea state parameters on the wind drag coefficient and therefore on the sea level predictions of the model. We evaluate the influence of the replacement of the bulk definitions with definitions of the wind drag coefficient depending on the wave parameters- wave age or wave steepness. The article may be considered as a continuation of the work of Mungov (Mungov and Daniel, 2000) on the operational storm surge modeling in the Western Black Sea, showing the recent improvements of this operational system.

2. DEPENDENCE OF THE WIND DRAG COEFFICIENT ON THE SEA STATE PARAMETERS

The wind driven surge is determined by the transfer of momentum from the atmosphere to the sea surface that generates a drift current. The transfer of momentum is expressed in the storm surge model using the wind shear stress (the shear force per unit of the contact area):

$$|\tau| = \rho_a * C_d * U_{10}^2 = \rho_a * u_*^2 \tag{1}$$

where C_d is the wind drag coefficient, ρ_a is the air density and U_{10} is the wind speed at 10m height above the sea surface. According to Monin- Obukhov similarity theory of the atmospheric boundary layer:

$$C_d = k^2 * \ln^{-2}\left(\frac{z}{z_0}\right) \tag{2}$$

where k is the von Karman constant, z is 10m in the case of the usage of 10m winds and z_0 is the aerodynamic roughness length of the sea surface. One may naturally expect that the sea surface roughness length depends on the sea state. According to Charnock model (Charnock, 1955):

$$z_0 = \alpha * \frac{uz}{g} \tag{3}$$

where α is the Charnock parameter, varying from 0.012 to 0.0185 according to different estimations and experiments.

Typically in the wave and circulation models the value of C_d is parameterized using linear functions of the 10m wind speed (the so called bulk formulations):

$$C_d = 0.001 * (a + b * U_{10}) \tag{4}$$

The coefficients *a* and *b* in 4) are usually determined by experimental in situ measurements in different oceans, wind speed ranges and atmospheric conditions. According to Wu (1975) *a* = 0.8 and *b* = 0.065 and according to Smith and Banke(1975) *a* = 0.63 and *b* = 0.066. The bulk formulations mean that the drag coefficient rises continuously with the rise of the wind speed. However the experimental measurements show that above 30m the behaviour of the drag coefficient is totally different. The studies of experimental results in situ in tropical cyclones and laboratory experiments by Powell et al (2003), Donelan et al (2004) and others show that above some wind speed (30m/s according to Powell and probably with different thresholds in different areas) the drag coefficient is not increasing, but stays constant and in extreme hurricane conditions above 45m/s even decreases due to the interaction of the wind with waves. Instead of usage of a coupled ocean-boundary layer model, Zijlema, van Vledder and Holthuijsen (2012) recently proposed the usage of a bulk parameterization (further referred as Zijlema formula in the study) that is based on second order polynomial fit:

$$C_d = 0.001 * (0.55 + 2.97 * U - 1.49 * U^2) \tag{5}$$

$$\overline{U} = \frac{U_{10}}{U_{ref}} \tag{6}$$

with a reference wind speed U_{ref} = 31.5m/s. Such function is rising to a maximal value at the reference wind speed and after that is decreasing.

Most of the studies (theoretical and experimental) on the dependence of the drag coefficient focus on two main approaches- expression of C_d as a function of the wave age

$$\beta = \frac{c_p}{u_*} \tag{7}$$

where c_p is phase speed of the waves at the spectral peek, or as a function of the wave steepness or as a combination of wave parameters. Stewart (1974) concluded that in the Charnock model the Charnock parameter is a function of the wave age and the function is

$$\alpha = f(\beta) = \alpha_1 * \beta^m \tag{8}$$

The values of α_1 and *m* are obtained by experiments. Negative value of m means that the growing wind sea (young waves) is rougher than the developed (mature) wind sea. Positive m means the opposite- the mature wind sea is with higher roughness. The field observations conclude that *m* is definitely negative.

Taylor and Yelland (2001) found that if the wind speed is above 12 m/s the dependence of C_d on the wave age becomes poor and argued that especially above that wind speed the dependence of the drag coefficient on the wave steepness is far more clearer. In a recent study Wang et al (2013) found, studying experimental measurements that the dependence of the drag coefficient on the wave age is in a good agreement with observations only if the friction velocity is below 0.5. Above that value wave steepness (or an expression based on Reynolds number) describes the behaviour of C_d much better.

Guan and Xie (2004) described a physical interpretation of the existing linear parameterizations of C_d as a function of the wind speed. They combined the logarithmic law of the wind within the atmospheric boundary layer with the Charnock relation and expressed the drag coefficient as a function of the wind speed with the Charnock's "constant" as a parameter. This function is nearly linear within the usually measured ranges of C_d with a slope determined by the value of α (eq.3). If α is not a constant, but a function of the wave parameters, after invoking the 3/2 power low (Toba, 1972):

$$\frac{gH}{u_*^2} = 0.062\left(\frac{gT}{u_*}\right)^{3/2} \tag{9}$$

This way the authors came to a relation for the wind drag coefficient that is a linear function of the wind speed with a slope determined by a function of the wave steepness:

$$C_d = 0.001*(0.78 + 0.475 * f(\delta) * U_{10}) \tag{10}$$

where $f(\delta)$ is equal to $0.85^B A^{1/2}\delta^{-B}$ and the meaning of *B* is the same as the meaning of *m* in eq.8. The coefficients *A* and *B* are experimentally determined. Some of the proposals are listed in Guan and Xie (2004) and in the present study we use the *A* and *B* proposed by Smith et al (1992) with $A = 0.5$ and $B = -1$. In the case of Smith's definition we are coming to the next relation:

$$C_d = 0.001*(0.78 + K * \delta * U_{10}) \tag{11}$$

where *K* is a tunable parameter. In the case of Smith's definition it is 0.4, note however that the value of 0.475 is a result of a statistical procedure performed by Guan and Xie that will make the relation to provide much lower values of the drag coefficient than the values provided by Wu and Smith and Banke formulations under moderate wind conditions, therefore the user may tune up *K* in case of usage of (11) to his own model and wind input. It is also possible to use other definitions of *B* which are negative but not equal to one (therefore the drag coefficient does not depend exactly linear to the wave steepness) – the ranges of *B* by various authors will lead to a dependence of δ^C with *C* varying from 1 to 2.

From the practical point of view there are two main ways to take into account the dependence of the drag coefficient on the sea state in a storm surge model – to use the wave steepness and the relation of Guan and Xie in coupled or non-coupled models (the wave steepness is a standard output of the wave models) or to use directly the drag coefficient as it is calculated in the source terms (actually meaning that we are using a drag coefficient dependent on the wave age). In the wave models different parameterizations of the source terms are used and while in some of them the drag coefficient (and so the friction velocity) are calculated using bulk formulations, others are computing the friction velocity in the source terms – for instance the source term of WAM cycle IV using the Janssen's theory (Janssen, 1991).

3. METHODS AND DATA

The storm surge model described in this study is the storm surge model of METEO FRANCE which operational implementation to the Black Sea is described by Mungov and Daniel (2000). The bathymetry is with 1/30° spatial resolution and the grid is a regular longitude-latitude grid.

The model is two-dimensional depth integrating model with equations written in spherical coordinate system:

$$\frac{\partial U}{\partial t} = fV - \frac{g}{R\cos\varphi}\frac{\partial \eta}{\partial \lambda} - \frac{1}{\rho R\cos\varphi}\frac{\partial P_a}{\partial \lambda} - \left[\frac{U}{R\cos\varphi}\frac{\partial U}{\partial \lambda} + \frac{V}{R}\frac{\partial U}{\partial \varphi}\right] + \frac{\tau_{sx} - \tau_{bx}}{\rho H} + A_H \nabla^2 U$$

$$\frac{\partial V}{\partial t} = fU - \frac{g}{R}\frac{\partial \eta}{\partial \varphi} - \frac{1}{\rho R}\frac{\partial P_a}{\partial \varphi} - \left[\frac{U}{R\cos\varphi}\frac{\partial V}{\partial \lambda} + \frac{V}{R}\frac{\partial V}{\partial \varphi}\right] + \frac{\tau_{sy} - \tau_{by}}{\rho H} + A_H \nabla^2 V$$

$$\frac{\partial \eta}{\partial t} = \frac{1}{R\cos\varphi}\left[\frac{\partial}{\partial \lambda}(UH) + \frac{\partial}{\partial \varphi}(VH\cos\varphi)\right]$$

containing advection terms, horizontal turbulence terms, nonlinear bottom friction terms, and surface friction terms. U, V are the depth integrated current velocities, f is the Coriolis parameter, R is the earth radius, P_a is the atmospheric pressure, H is the water depth, η is the water free surface elevation, ρ is the water density and A_H is the horizontal diffusion coefficient, τ_{sx} and τ_{sy} are the components of the wind stress and τ_{bx}, τ_{by} – the components of the bottom friction stress.

The wind stress components are:

$$\tau_{sx} = \rho_A C_d \lfloor W_{10} \rfloor W_{10x}\ , \quad \tau_{sy} = \rho_A C_d \lfloor W_{10} \rfloor W_{10y}$$

where ρ_A is the air density, W_{10x} and W_{10y} are the components of the 10m wind speed W_{10}.

The bottom friction stress components are:

$$\tau_{bx} = \rho C_b \left(U^2 + V^2\right)^{1/2} U\ , \quad \tau_{by} = \rho C_b \left(U^2 + V^2\right)^{1/2} V$$

where C_b is the bottom drag coefficient which is set in our model to 0.0015 over the shelf and 0.000015 over liquid bottom. We integrate not down to the actual bottom of the sea but to the mixed layer for the month taking into account the very stable stratification of the Black Sea and treating the surge as a long wave propagating only in the upper dynamical layer of the sea. The mixed layer depth of the Black Sea is very shallow when compared with the other European seas (in order of 40-60m during the winter, compared with the order of 300m for the Mediterranean sea). The data for the monthly mixed layer depth of the Black Sea was taken by the study of Kara et al (2005).

The wave model that is used in the study is SWAN version 40.91.ABC (Booij et al 1999). The computational domain is the same as the computational domain of the storm surge model. SWAN is configured to provide an output of the wave steepness and the friction velocity (applicable only if WAM cycle 4 source terms are set in the namelist). The atmospheric forcing data is available with a temporal resolution of one hour. SWAN provides the wave parameters also with one hour interval between the outputs.

The storm surge model reads the wave model output of friction velocity or steepness and in the case of usage of SWAN the data is directly processed (because the SWAN grid is the same as the storm surge model grid) by the subroutine that calculates the wind drag coefficients and the wind stress components.

Data for comparisons of the surge model output is available for two cases: one for the tide gauge Irakli (the middle part of the Bulgarian coast) and one for the tide gauge Ahtopol (southernmost part of the coast). Both tide gauges are located in areas where the coastline is almost straight and simple bottoms with uniform slope and therefore no significant local topographic effects are expected. The data is from the studies of Mungov (2000) and Andreeva (2011).

The data available at the present moment is the maximal daily surge levels. Hourly data is not available and so the possible time shifts of the peaks of the storms cannot be determined, but due to the fact that the tide in Black Sea is below 8 cm this is not of primary importance and the most important parameter for the operational storm surge forecast is the maximum daily sea level rise.

The fields wind components at 10m height and the mean sea level pressure for some of the most significant cases of a storm surge at the Bulgarian coast are provided by the Bulgarian regional atmospheric modeling group by a downscaling of the ERA-Interim (Dee et al, 2005) reanalysis using the ALADIN model (Bubnova et al 1995; Bogatchev, 2008). For more details about this specific downscaling see Galabov et al (2015). The domain covers the entire Black Sea. The integration starts several days before the beginning of the historical storm at calm weather and so we may expect the sea level at the beginning of the run of the storm surge model to be equal to the mean undisturbed level.

4. RESULTS

The first presented case is the most important one, because it is the highest ever recorded in the Black Sea storm surge (for more details about this storm see Galabov et al (2015)). Storm caused a surge of 1.43 peak value at the tide gauge Irakli and more than 1.5m at the tide gauges in the ports of Varna and Burgas.

Initially we tested the approach to integrate not to the bottom but to the mixed layer depth.

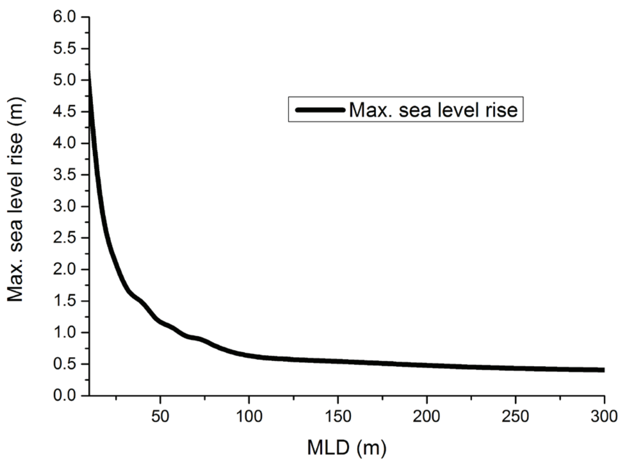

Fig. 1. Dependence of the maximum sea level rise of the depth of integration for the storm of February 1979

We varied the depth of integration and the result is shown on Fig.1. As it may be seen, when integration is down to the bottom of the sea (more than 2000m) the model fails to reproduce any surge above 40cm. On the other hand when the integration is down to the mixed layer depth (which is 40-50m for February) we obtain realistic simulation result comparable with the tide gauge measurement when using the Smith and Banke formulation of the wind drag coefficient. In the study of Krestenitis et al (2012) the usage of Smith and Banke formulation of C_d results in complete failure to reproduce properly even significantly lower surges (below 1 m) and they are tuning their model with a very high values of C_d – 8.10^{-3} when the wind speed is above 20 m/s. The problem is that while it is possible to tune up a model this way, such values of C_d are physically meaningless and taking into account the mentioned previously experiments of Powel and others, such values are much higher than the highest possible in the oceans. Moreover while this is a temporal solution to fit a model to observations, even with such high C_d the model will fail to reproduce the extreme surges like the one in 1979 and after all the important surges are those above 80 cm, while all surges below 50 cm are without practical importance for the operational practice. The conclusion is that the correct way to do storm surge modeling in the Black Sea is to consider the shallow mixed layer and to integrate only to MLD. It is important also to note that climate projections of the storm surges in the Black Sea cannot be performed by the usage of such two dimensional models because of the argument, that the surge is depending strongly on the mixed layer depth. A future change of the temperature of the upper layer of the sea will inevitably change the mixed layer depth and in the case of warming it will lead to a shallower depth of the mixed layer during the winter months and then even weaker storms may result in much higher surges (see Fig. 1 for instance for MLD = 30m for the storm of 1979). Therefore any such projections that ignore the change in the sea stratification underestimate the future extreme events and possibly the underestimation is very serious. This means, that the ONLY way to do a projection of the future storm surges in the Black Sea is to use baroclinic hydrodynamic models and such two dimensional models as the presented one are not applicable for such tasks in order to be able to account the possible increase of storm surge danger even when the storms are weakening.

The first case for which we evaluate the influence of different approaches to define the drag coefficient is the storm of February 1979, which is the highest recorded surge that caused damages and coastal inundation. For a description of this storm see Galabov et al (2015) and Belberov et al (2009). We compare the model simulations with the daily maximum surge, measured at Irakli tide gauge. Fig. 2. shows the performance of the model using three different bulk parameterizations of C_d– the widely used Wu and Smith-Banke formulations and the second order polynomial proposed by Zijlema et al (2012). Both Wu and Smith-Banke formulations result in overestimation of the maximum level by 10–15cm on 19.02.1979 (the peak of the storm) which is in the order of 10% overestimation. The quadratic relation of Zijlema leads to underestimation (that in principle is possible to avoid by rescaling of the relation). The next experiment is using the two approaches to account the dependence of the C_d on the sea state parameters. When we use the drag coefficient obtained by the usage of friction velocity from SWAN (using the Janssen wind input) we overestimate the surge even higher than the overestimation by the bulk formula. The usage of the formulation based on the wave steepness leads to successful prevention of the overestimation for the highest surge during this storm. The simulations on 20.02.1979 are not successful but this is because of the wind fields and after all the most important is the proper simulation of the storm peak on 19.02.1979 for which the C_d based on wave steepness performs best.

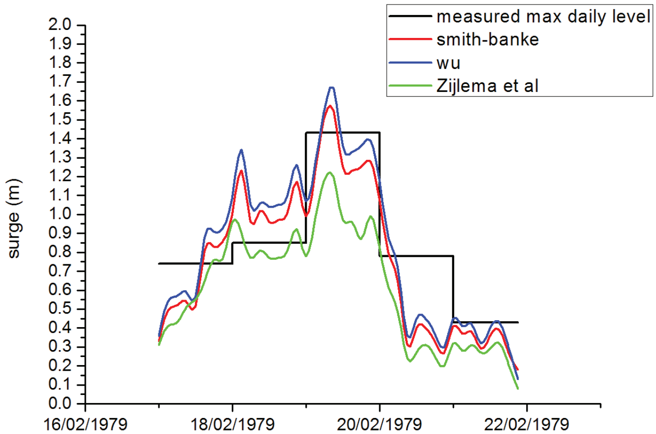

Fig. 2. Comparison of the measured daily maximum surge measured at Irakli tide gauge with model simulations using Smith-Banke and Wu bulk parameterizations of C_d and the second order bulk formula of Zijlema et al (2012).

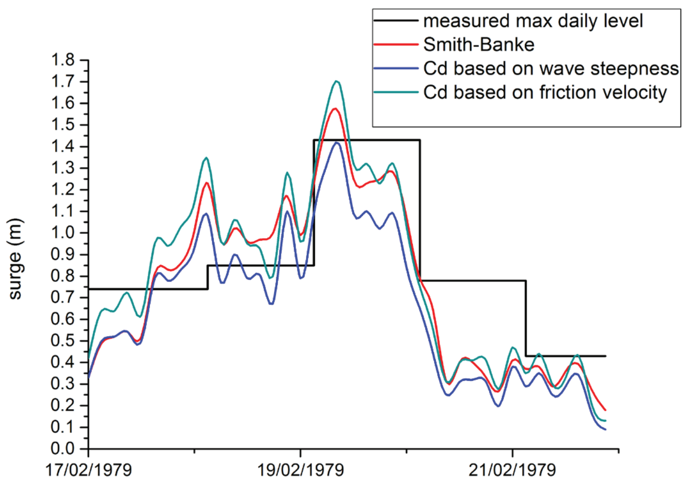

Fig. 3. Comparison of the measured daily maximum surge measured at Irakli tide gauge with model simulations using the default Smith-Banke formula and Cd estimated using the wave steepness and C_d based on the friction velocity obtained by the SWAN wave model.

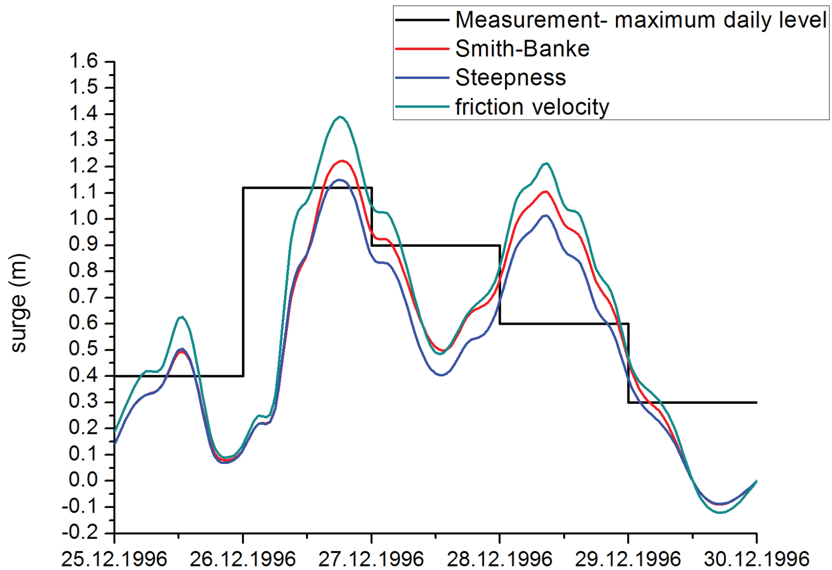

Fig. 4. Comparison of the measured daily maximum surge during the storm of December 1996 measured at Ahtopol tide gauge with model simulations using the default Smith-Banke formula and C_d estimated using the wave steepness and C_d based on the friction velocity obtained by the SWAN wave model.

The next case is the storm of 12.1996. The storm surge model performance was studied in the work of Mungov and Daniel (2000) and an underestimation of the model was found, however in the present study there is no underestimation because the resolution of the wind input is higher and the wind interval between the atmospheric input files is 1h instead of 6h in the study of Mungov and Daniel. The results are presented on Fig. 4 and the conclusions are the same as for the previous case. The highest surge was recorded on 26.12.1996. The model simulates a second peak on 18.12.1996 that was not observed, because of the wind fields downscaling. Otherwise, the usage of the friction velocity again leads to high overestimation, while the usage of the wave steepness prevents the overestimation.

The next case is the storm of January 1998. That storm was also studied by Mungov and Daniel (2000). The results of experimental runs are presented on Fig. 5 and the conclusions are identical to the conclusions for the two previous cases. The lowest overestimation was achieved when using the steepness. Outside of the storm peak the use of the wave steepness leads to identical results to the results using the Smith and Banke formulation. For the peak of the storm however the coupled version of the model lowers the tendency of the model to overestimate.

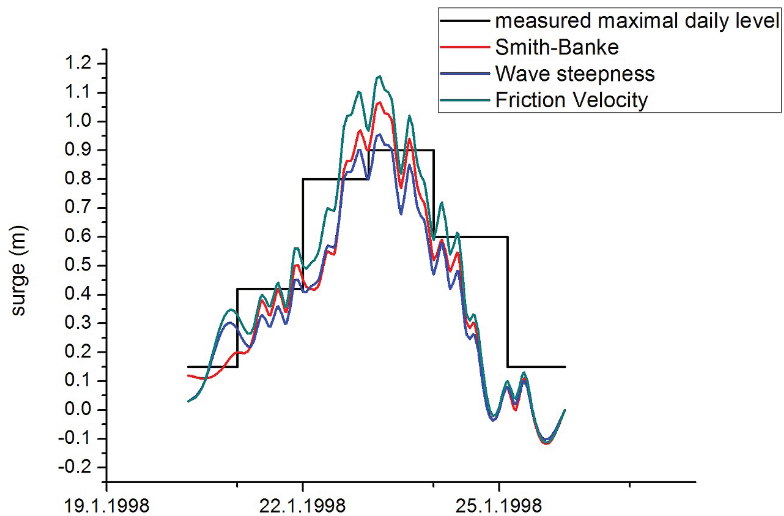

Fig. 5. Comparison of the measured daily maximum surge measured at Irakli tide gauge with model simulations for the storm of January 1998 using the default Smith-Banke formula and C_d estimated using the wave steepness and C_d based on the friction velocity obtained by the SWAN wave model.

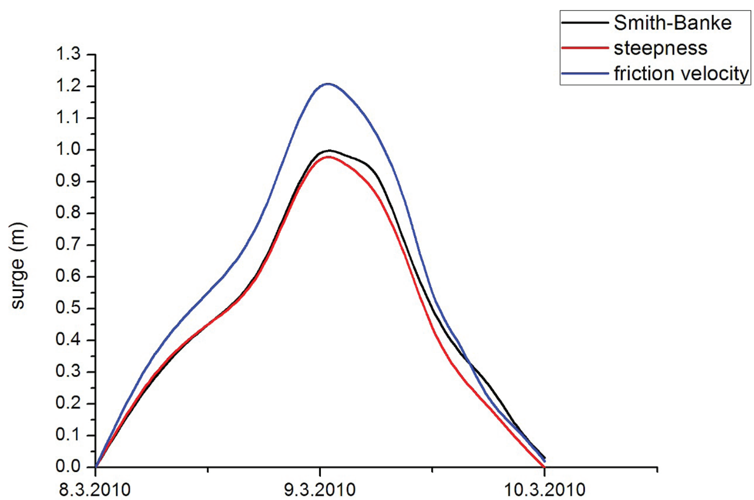

Fig. 6. Model simulations using the default Smith-Banke formula and C_d estimated using the wave steepness and C_d based on the friction velocity obtained by the SWAN wave model- the storm of March 2010.

The next case is a storm during March 2010. For this case there is no significant difference between the result when using the Smith and Banke formula and when using the steepness. The data provided to us by the Bulgarian Oceanologycal Institute (Valchev et al, 2014) shows that the highest hourly value at Shkorpilovtsi was 1 m, so in this case the model does not overestimate when using the bulk formulation of Smith and Banke. The simulation using the wave steepness is with the same maximum value. Therefore in such cases the use of wave steepness does not leade to underestimation and converge towards the true value better than the bulk formulation and not just permanently produce systematically lower values of the surge. This is an obvious advantage of the use of wave steepness based C_d in the model instead of the bulk formulas and definitions based on friction velocity (wave age).

5. CONCLUSIONS

Quantitative conclusions in the present study are difficult, because we have a limited number of daily maximum sea levels. The storm surges in the Western Black Sea are not a frequent event (especially the extreme ones). While it is possible to obtain the statistical parameters for many cases with sea level rise within the limits of 60 cm, such comparisons are without practical importance because such surges does not lead to coastal inundation and significant hazard. The surges of 1 m and above are with critical importance and the studies of the operational modeling in the Black Sea must be focused on such scenarios of high surges. If we realy focus on such cases the conclusions are qualitative rather than quantitative due to their low number. The obvious conclusion however is that the approach using the wave steepness gives much better results than the approach with the friction velocity (which means dependence of the drag coefficient on the wave age). It is likely that the same is valid in other tideless basins with a limited fetch where the wind sea is dominant during the storms that cause significant surges. However, in the oceans it may be the opposite (as it is suggested in some of the referenced studies). There are also hints that the dependence of the drag coefficient on the wave steepness may act as a limiter that prevents the storm surge models from overestimation of the surge. If the model systematically underestimates the surges due to the wind input being with too coarse resolution in space and time, a bulk formulation may be preferable after proper tuning (the quadratic relationship of Zijlema, van Vledder and Holthujsen may be a better alternative than the linear bulk formulations). The approach presented here is allready implemented in the operational system of marine models of NIMH-BAS.

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