ISSN 0001-4338, Izvestiya, Atmospheric and Oceanic Physics, 2015, Vol. 51, No. 3, pp. 337–341. © Pleiades Publishing, Ltd., 2015.
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Wind Waves as an Element of a Hydrodynamic Coupled Ocean–Atmosphere Model

D. V. Chalikov* and K. Yu. Bulgakov

Shirshov Institute of Oceanology, St. Petersburg Branch, Russian Academy of Sciences, ul. Pervaya Liniya 30, St. Petersburg, 119053 Russia

**e-mail: Dmitry-chalikov@yandex.ru*

Received October 23, 2014

Abstract—A new approach to the coupled simulation of the ocean, atmosphere, and sea waves based on a detailed simulation of ocean-surface processes is described. The role of surface waves and the chain of energy and momentum transformation are briefly described: energy and momentum are transferred by wind to waves, turbulence, and surface currents through the field of surface pressure and tangential stress. Both energy and momentum accumulated in waves are transferred (with a lag) also to currents and turbulence within the ocean surface layer. The possibility of coupling atmosphere, ocean, and sea-wave models is considered.

Keywords: waves, ocean, atmosphere, simulation, interaction

DOI: 10.1134/S0001433815030032

1. INTRODUCTION

The improvement of models describing a hydrometeorological medium becomes a currently central problem which calls for considerable resources and new technologies [1]. On the whole, the scientific basis for such models has been developed; however, scientific recommendations for the description of physical processes occurring in the atmosphere and ocean cannot completely be used because of the insufficient capacity of computing aids. At the same time, adequate practical algorithms are yet to be formulated for many important physical processes. One of the most important disadvantages of the coupled ocean–atmosphere simulation is the absence of a physical approach to the coupling of atmospheric and oceanic models through a detailed description of ocean-surface conditions—wind waves [2, 3].

Wind waves affect the intensity of interaction and, tending to lag, they accumulate high momentum and energy. The wave field gains energy and momentum from the atmosphere, transports them long distances, and gives them up to currents and the atmospheric boundary layer. Calculations of heat, momentum, and energy fluxes should be based on the theory of the wave boundary layer, the lower part of the near-water layer within which vertical momentum and energy fluxes are produced due to waves. This theory has sufficiently been developed [4–7]. In the wind–wave–oceanic upper layer system, there is a complex cycle of energy and momentum transformation: the wind transfers energy and momentum to waves through the field of surface pressure and tangential stress. Collapsing and volumetrically dissipating waves transfer momentum to currents and energy to currents and turbulence. Waves and currents can also exchange momentum through radiation stresses. All these types of transformation can be calculated separately. Wave dissipation may occur far from the region of wave generation so that there are two friction-stress vectors at every point, and, thus, the ocean–atmosphere interaction is nonlocal. Such energy and momentum transformations occur everywhere, especially in the regions of large horizontal gradients of surface currents and in regions of continental shelves, in which the velocity of currents produced by waves may reach an order of 1 m/s.

Figure 1 shows estimates of the wind-velocity dependence of the momentum and energy of waves and wind for the case of steady-state roughness and currents. The wave momentum and energy were calculated using a conformal wave model, and the momentum and energy of currents were estimated under the assumption that the currents are uniform in depth within a mixed layer 50 m in thickness and the current velocity amounts to $0.03U_{10}$ (U_{10} is the wind velocity at a height of 10 m). One can see that the momentum of waves is smaller than that of currents by approximately an order of magnitude. However, the wave momentum may be so high that, at its fast transfer to currents, the current velocity may reach 1 m/s. Such phenomena occur when large waves move into shallow waters, in which these waves begin to actively collapse. At a wind velocity of 18 m/s, the energy of waves begins to exceed that of currents. These effects

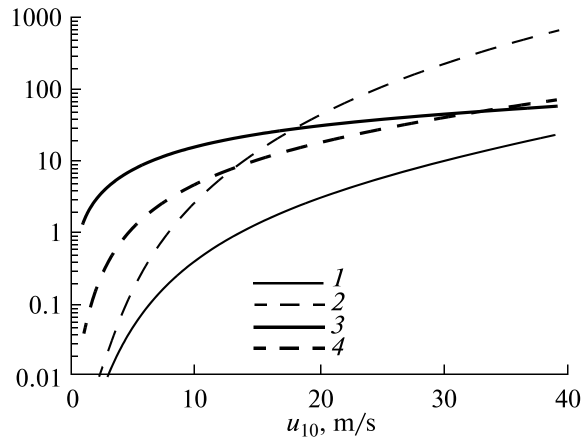

Fig. 1. Wind-velocity dependence of momentum (m^2/s) and wave energy (m^3/s^2): (*1*) wave momentum, (*2*) wave energy, (*3*) current momentum, and (*4*) current energy.

are especially manifested in ocean (sea) models with a high resolution. This circumstance implies that a valid ocean–atmosphere interaction model should include a surface-wave model as a coupling element. It is important that the fast transport of energy and momentum makes this interaction nonlocal. Waves can accumulate a significant amount of energy and momentum, transport them for long distances, and contribute them to the currents and energy of the ocean surface layer. Thus, currents can be generated by both stresses induced by wind and by collapsing waves. These stress vectors may be directed at any angle to each other. All the above effects increase with improved model resolutions.

2. MODELS

The main prognostic variables of a typical weather (climate) forecast model are as follows: components of wind velocity and disturbances in potential temperature, geopotential, surface pressure, and specific humidity. Parameterizations, i.e., algorithms for describing small-scale processes in terms of basic large-scale variables, are used for reproducing processes (such as turbulent exchange, cloud microphysics, radiation transport, and others) which cannot explicitly be reproduced because of their scales. The most important component of coupled atmosphere–ocean models is a boundary-layer model. Most models use the Monin–Obukhov similarity theory. Almost none of the models make distinctions between land and sea: some modifications are used only for the roughness parameter. However, the roughness parameter is known to vary within a wide range, and its value depends not only on wind velocity, but also on the state of the sea surface. The relationship between waves and the atmospheric boundary layer may be described in detail with the aid of a wave boundary layer (WBL) model.

2.1. WBL Model

The WBL is defined as an atmospheric near-water layer within which a noticeable portion of vertical momentum and energy fluxes is due to wave-induced variations in wind velocity and pressure. The height of the WBL is close (in order of magnitude) to the significant wave height H_s (which is the mean height of the one third of the highest waves); i.e., it does not exceed a few meters. The lower part of the WBL adjoins the ruffled surface and its upper part passes into the Monin–Obukhov boundary layer. In most cases, the WBL height H_a does not exceed the height of the dynamic sublayer; therefore, the direct influence of stratification is negligibly small in the WBL. A WBL model (developed in [6]) is based on the evolutionary equations for the two components of wind velocity $\mathbf{u}$, kinetic turbulence energy e, and dissipation ε:

$$\frac{\partial \mathbf{u}}{\partial t} = \frac{\partial}{\partial z}(\tau + \tau_w), \quad (1)$$

$$\frac{\partial e}{\partial t} = D_e + P - \varepsilon, \quad (2)$$

$$\frac{\partial \varepsilon}{\partial t} = D_\varepsilon + \varepsilon_e(c_2 P - c_4 \varepsilon), \quad (3)$$

where τ and τ_w are the momentum fluxes caused by turbulence and wave components; D_e and D_ε are the vertical diffusions of turbulence energy and dissipation; and P is the rate of the production of turbulence energy, which is determined by the scalar product $P = \frac{\partial \mathbf{u}}{\partial z}(\tau + \tau_w)$. This system of equations may be supplemented with temperature and humidity equations and relations for calculating the coefficients of turbulent exchange. The fundamental difference between the WBL and an ordinary boundary layer is in the occurrence of a new variable—the wave momentum flux τ_w produced by wave fluctuations in wind velocity and pressure [6]. Temperature, turbulence energy, and the profile of wind velocity are determined from the solution of Eqs. (1)–(3); then one can calculate the flux of energy transferred to waves.

According to the Miles classical theory [8], the Fourier components of surface pressure p_0 are related to the Fourier components of surface elevation

$$p_k + ip_{-k} = (\beta_k + i\beta_{-k})(h_k + ih_{-k}), \quad (4)$$

where β_k and β_{-k} are the real and imaginary components of the β-function (i.e., the Fourier coefficients for cosine and sine, respectively, and for p_k, p_{-k}, h_k, and h_{-k}). These coefficients depend on the dimensionless virtual frequency $\Omega = \frac{\omega U}{g}\cos(\varphi - \theta)$. It is natural that the height at which wind velocity is specified should be different for different frequencies; therefore, Ω is determined as follows:

$$\Omega = \frac{\omega U(\lambda/2)}{g}\cos(\varphi - \theta) = \frac{U(\lambda/2)}{c}\cos(\varphi - \theta), \quad (5)$$

where U is the wind velocity at the height $z = 1/2$, where the wave length is $\lambda = \frac{2\pi g}{\omega^2}$ and the phase velocity is $c = \frac{g}{\omega}$.

Formula (4) is used in simulating waves in order to calculate surface pressure and energy and momentum fluxes transferred to waves. For spectral calculations, it is necessary to have only the imaginary component of pressure (below, this component will be denoted by β without index), which is approximated by the expression

$$\beta = \begin{cases} b_1 + d_1(\Omega - \Omega_1) & \Omega < \Omega_1 \\ b_0 + a_0(\Omega - \Omega_0) + a_1(\Omega - \Omega_0)^2 & \Omega_1 \geq \Omega < \Omega_2 \\ b_2 + d_2(\Omega - \Omega_2) & \Omega \geq \Omega_2 \end{cases} \quad (6)$$

where the numerical parameters are determined as follows: $\Omega_0 = 0.7$, $\Omega_1 = -19.3$, $\Omega_2 = 20.7$, $\Omega_5 = 21.2$, $a_0 = 0.02277$, $a_1 = 0.09476$, $b_0 = -0.02$, $b_1 = = 37.43$, $b_2 = 38.34$, $b_4 = 0.07$, $d_1 = -3.768$, and $d_2 = 3.813$.

The spectral velocity of the flux of energy transferred to waves $F_e(m^3s^{-1})$ and the spectral components of the momentum-flux vector F_x and F_y are calculated from the formulas

$$F_e(\omega, \phi) = g\omega\beta(\Omega)S_e(\omega, \phi), \quad (7)$$

$$F_x(\omega, \phi) = k_x\beta(\Omega)S_e(\omega, \phi)\cos\phi,$$
$$F_y(\omega, \phi) = k_y\beta(\Omega)S_e(\omega, \phi)\sin\phi. \quad (8)$$

According to [6], the profile τ_w can be calculated from the formula

$$\tau_w = \int \mathbf{k} g\beta_{-k}(\tilde{\Omega}_k)S(\omega, \theta)\Delta\omega \times \exp(-G(\bar{\omega})kz)(d|k|)(d\omega), \quad (9)$$

where $S(\omega, \theta)$ is the wave spectrum (ω is frequency and θ is the angle between wind and wave-mode directions) and $G(\omega/\omega_p)$ is the function close to 1 (ω_p is the wave-crest frequency). The WBL model and the approximation of the β-function are described in detail in [6, 7]. Thus, the WBL model establishes a direct relation between the WBL structure and the wave spectrum. The model components may be dependent on the spectral resolution of a wave model.

2.2. A Wave Model

Basically, all available wave-forecast models (in a varying degree) are based on the WAM model [9–11] developed at the European Center for Medium-Range Weather Forecasts in the 1980s. Later, different modifications of this model were developed and adapted to shallow waters in the ocean and lakes and to different regions. A typical wave model is based on the equation for the evolution of the spectral energy density $S(\omega, \theta)$ (or the wave action S/ω).

$$\frac{\partial S(k, \omega)}{\partial t} = A_w + N + I_w - D_w, \quad (10)$$

where A_w, I_w, and D_w are the advection, influx, and dissipation of wave energy, respectively, and N is the rate of the wave-spectrum evolution due to nonlinear interactions. The energy advection A_w in every spectral range is determined by group velocity. The energy influx I_w is calculated within the scope of the WBL model (see equations in [1, 2]). It is considered that the wave-energy dissipation D_w is caused by collapsing waves, orbital-motion dissipation, near-bottom friction, and irreversible energy flux into the spectral subgrid region. The results of a comparison between model and observational wave-field data show that, despite a significant idealization of wave processes, current wave models describe the evolution of waves quite adequately.

2.3. An Oceanic Upper Layer Model

The evolution of temperature T and turbulence energy e in the oceanic upper layer is described by the system of differential equations:

$$\frac{\partial T}{\partial t} = A_T + \frac{\partial}{\partial z}K_T\frac{\partial T}{\partial z}, \quad (11)$$

$$\frac{\partial e}{\partial t} = A_e + \frac{\partial}{\partial z}K_e\frac{\partial e}{\partial z} + P + B - \varepsilon_e, \quad (12)$$

where A_T and A_e are the T and e advections determined from the three-dimensional current field, z is the vertical coordinate, K_T is the coefficient of vertical turbulent exchange, P is the energy influx due to the turbulent dissipation of currents and orbital wave motions, B is the term describing the buoyancy effects, and ε_e is the rate of turbulence energy dissipation. The divergence of the vertical flux of heat takes into consideration the heat exchange with the atmosphere and deep layers. The divergence of the vertical flux of turbulence energy takes into a consideration the influx of turbulence energy from collapsing waves D_w (see Eq. (5)).

$$\left(K_e\frac{\partial e}{\partial z}\right)_{z=0} = \int_{\omega,\theta} D_w d\omega d\theta, \quad (13)$$

where the integration is performed over the entire range of wave numbers and angles.

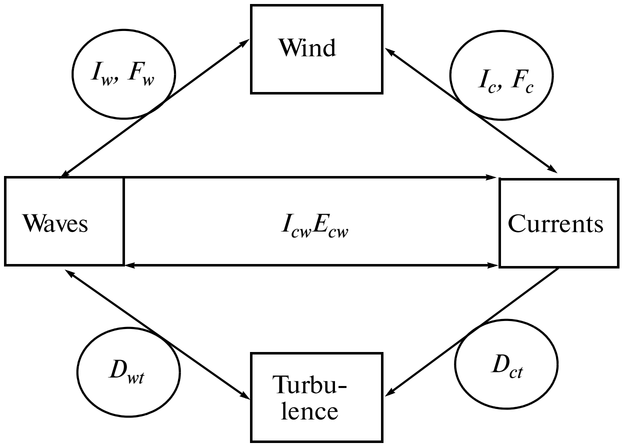

Fig. 2. Scheme illustrating the momentum and energy exchange between wind, currents, and waves. The rectangles correspond to the interacting objects and the elements of interaction are given in circles. See explanatory notes in the text.

2.4. Dynamic Wind-Wave–Current Interaction

There are several interaction mechanisms when wind blows over the sea surface (see Fig. 2 which shows the main components of the system and their interaction). Momentum is usually transferred from wind to currents through a thin near-surface layer. The wave spectrum is usually very wide and reaches high frequencies. Prognostic wave models usually have a rather low resolution; therefore, one has to assume that the form of the wave spectrum in its subgrid region is known (for example, this form satisfies the ω^{-5} law). One has also to assume that subgrid (i.e., short) waves dissipate on the spatial scales of a grid used in a numerical model of waves and currents and transfer the total momentum to currents. The momentum flux from wind to currents I_c can be calculated using the conventional formula

$$\boldsymbol{\tau}_c = C(\omega_r)|U_{10}|\mathbf{U}_{10}, \tag{14}$$

where the friction coefficient C depends on the limiting frequency of the wave spectrum ω_r. The energy flux from wind to currents F_c is $\boldsymbol{\tau}_c \cdot \mathbf{u}_0$ (u_0 is the surface-current velocity). The spectral energy flux to waves F_w is determined from (7) and the momentum flux is determined from (8). The total rate of energy exchange between wind and waves (I_w and F_w) is calculated though the integration of (7) and (8) over the entire range of frequencies and angles. Note that this exchange may change its sign: in the case of weak wind and developed waves, momentum and energy can be transferred from waves to wind. If wind and waves are differently directed, the wind attenuates waves, i.e., it decreases their energy and momentum. In most cases, the momentum transferred to waves amounts to only a few percent of the total momentum flux; however, since waves are conservative, the wave momentum can accumulate and be transferred for long distances. The total momentum flux towards currents and wind $\tau_c + I_c$ is equal to the momentum flux τ_H at the upper boundary of the WBL, which is calculated from the standard relation $\boldsymbol{\tau}_c = C_{10}|U_{10}|\mathbf{U}_{10}$, where C_{10} is the total coefficient of resistance.

Turbulence within the oceanic upper layer is supplied by energy through the dissipation of shear currents, the volume dissipation of orbital wave motions, and breaking waves. These latter transfer momentum and energy to currents and turbulence with the rates I_{cw} (vector) and E_{cw} and partially to waves on other frequencies (the last mechanism remains to be explored).

$$I_{cw} = \int_{\omega,\theta} g\mathbf{k}DS(\omega,\theta)d\omega d\theta \tag{15}$$

where D is the dissipative function which is similar to the β function. The portion of energy transferred to currents can be calculated from the formula

$$E_{cw} = I_{cw}\mathbf{u}_0.$$

The remaining energy is transferred to turbulence in the oceanic upper layer and is taken into account in the boundary condition for the diffusion term in Eq. (12). In regions with strong horizontal gradients, waves and currents can exchange energy and momentum through radiation stresses (see [12]).

The volume influx of turbulence energy P is calculated from the formula

$$D = D_{wt} + D_{ct}, \tag{16}$$

where D_{ct} is the standard term calculated from the components of the tensor of the rates of deformation of horizontal wind velocity components and D_{wt} is the rate of the volume dissipation of wave energy due to the interaction between orbital velocities and the components of the vortex of nonpotential motions [13, 14]:

$$P_{wt} = 3.87\times10^{-7}H_s^{1/2}g^{3/2}\exp(0.506\tilde{z} + 0.0057\tilde{z}^2), \tag{17}$$

where $H_s = 4(\int Sd\omega d\theta)^{1/2}$ is the height of significant wave, $\tilde{z} = z/H_s$ is the dimensionless depth, and g is the gravitational acceleration.

Thus, the intensity of turbulence in the oceanic upper layer is associated with a wave model through the surface influx of energy due to collapsing waves and its volume influx due to the dissipation of orbital wave motions. In simulating processes in shallow waters, one should take into consideration the near-bottom friction in a wave model and the influx of turbulence energy to the near-bottom wave boundary layer.

CONCLUSIONS

This study does not claim to be exhaustive; however, it contains information sufficient for formulating a model based on atmospheric, oceanic, and wave models. Details in coupling these models may further be specified by including additional factors, for example, the momentum and energy exchange between waves and currents due to radiation stress. Current wave models of the WAVEWATCH type [9] are fully adapted for use in the available coupled atmosphere–ocean models. To adequately account for the above effects, it is necessary that both spectral and angular distributions be significantly improved. At the same time, one can hardly expect that in some time a non-local wave model will be capable of resolving the entire range of wind waves. Therefore, the problem of separating momentum and energy fluxes transferred to waves and currents calls for further special investigations.

The coupled simulation of waves and the oceanic upper layer (without an atmospheric model so far) has recently started at several world scientific centers. Scientists have started to develop such schemes at several institutes in different countries, for example, at Miami University under the guidance of M. Donelan [8] and at the Rutgers University under the guidance of I. Ginis in the United States and at the Max Planck Institute in Germany. Such models are also being constructed at the US Naval Research Laboratory. This topic is new; therefore, there are few works published in journals. However, many publications are available in electronic form. All works of such a kind are based on the WAVEWATCH model [4–6] and on both oceanic and atmospheric models of different levels of complexity.

ACKNOWLEDGMENTS

This work was supported by the Russian Foundation for Basic Research (project no. 14-05-00422).

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Translated by B. Dribinskaya