Ocean Modelling 29 (2009) 261–268

Contents lists available at ScienceDirect

Ocean Modelling

journal homepage: www.elsevier.com/locate/ocemod

Coupling winds to ocean surface currents over the global ocean

Zengan Deng [a,b,*], Lian Xie [b], Bin Liu [b], Kejian Wu [a], Dongliang Zhao [a], Ting Yu [c]

[a] *Physical Oceanography Laboratory, Ocean University of China, Qingdao, Shandong 266100, PR China*
[b] *Department of Marine, Earth and Atmospheric Sciences, North Carolina State University, Raleigh, NC 27695, USA*
[c] *National Marine Data and Information Service, State Oceanic Administration, Tianjin 300171, PR China*

ARTICLE INFO

Article history:
Received 29 October 2008
Received in revised form 22 April 2009
Accepted 10 May 2009
Available online 18 May 2009

Keywords:
Wind–Current Coupled System
HYCOM
Wind stress
Heat flux
Ocean surface boundary layer
ESMF
Ocean current

ABSTRACT

A Wind stress–Current Coupled System (WCCS) consisting of the HYbrid Coordinate Ocean Model (HYCOM) and an improved wind stress algorithm based on Donelan et al. [Donelan, W.M., Drennan, Katsaros, K.B., 1997. The air–sea momentum flux in mixed wind sea and swell conditions. J. Phys. Oceanogr. 27, 2087–2099] is developed by using the Earth System Modeling Framework (ESMF). The WCCS is applied to the global ocean to study the interactions between the wind stress and the ocean surface currents. In this study, the ocean surface current velocity is taken into consideration in the wind stress calculation and air–sea heat flux calculation. The wind stress that contains the effect of ocean surface current velocity will be used to force the HYCOM. The results indicate that the ocean surface velocity exerts an important influence on the wind stress, which, in turn, significantly affects the global ocean surface currents, air–sea heat fluxes, and the thickness of ocean surface boundary layer. Comparison with the TOGA TAO buoy data, the sea surface temperature from the wind–current coupled simulation showed noticeable improvement over the stand-alone HYCOM simulation.

1. Introduction

Wind stress plays a central role in the understanding and modeling of the air–sea interaction processes. Historically, there have been extensive studies on the wind stress drag coefficient, C_d, a parameter which is closely related to the expression of wind stress. Many formulas describing the relationship between the drag coefficient and wind speed have been derived in the past, such as those by Garratt (1977), Donelan et al. (1997) and Yelland et al. (1998). Traditionally, wind stress coefficient C_d is often expressed as a linear function of wind speed. However, large scatter in the observed drag-wind speed relationship suggests that C_d is also influenced by other environmental variables, such as wave status and ocean surface currents. Determination and quantification of the factors influencing the drag coefficient remain an important contemporary issue in air–sea interaction research. Possible impacts of ocean currents on drag coefficient were discussed in both theoretical studies, e.g., Hwang (2005), and shown for various regional cases of the ocean, e.g., Wuest and Lorke (2003). Owing to its critical role at the air–sea interface, accurate parameterization of the wind stress is necessary for assessing the reliability of many oceanic and atmospheric model predictions, with which we could study the sea state, large-scale ocean circulation, and climate variability. Most studies often approximate the surface wind stress τ exerted by the atmosphere on the ocean as a function of the 10-m wind velocity, U_{10}, which neglects the effects of ocean surface currents. However, in regions of strong currents, e.g., Kuroshio and Gulf Stream, it is important to introduce the difference in near-surface winds and surface ocean currents into the wind stress calculation (Bye, 1967; Rooth and Xie, 1992; Xie and Rooth, 1994). Kelly et al. (2001) confirmed that the scatterometer-derived wind stress over the tropical Pacific is a function of air–sea velocity difference. Introducing the ocean surface velocity into the stress formulation has been shown to significantly improve the modeling of wind stress in the tropical Atlantic Ocean (Pacanowski, 1987), and in the tropical Pacific Ocean (Luo et al., 2005). The effect of ocean surface currents on the bulk formulation of wind stress and heat flux in a 1/5° resolution model of the North Pacific was investigated by Jordan and Thompson (2006), and large impacts are found on wind stress, heat flux, and wind power input to the ocean. Kara et al. (2007) concluded that the effects of ocean surface velocity lead to a reduction of sea surface drag coefficient C_d, by about 2% globally. Recently, Deng et al. (2008) quantitatively investigated the effects of ocean surface currents and waves on drag coefficient and wind stress by using the outputs from HYCOM and wave model WaveWatch III.

Although including the effects of ocean surface current velocity can lead to significant improvements in wind stress calculation, it is not well known what quantitative impacts such improvements in wind stress may have on ocean circulation. Xie and Rooth

* Corresponding author. Address: Physical Oceanography Laboratory, Ocean University of China, No. 238 Songling Road, Qingdao 266100, Shandong, PR China.
E-mail address: dengzengan@gmail.com (Z. Deng).

1463-5003/$ - see front matter © 2009 Elsevier Ltd. All rights reserved.
doi:10.1016/j.ocemod.2009.05.003

(1994) investigated the effect of the current-induced stress perturbation which appears as a damping mechanism for the ocean circulation in an idealized coupled atmosphere–ocean model, but without modification to the drag coefficient. Changes in surface wind stress result in changes in upper ocean dynamics, such as effects on ocean surface currents, and ocean surface boundary layer depths. Our experimental work focuses on the analysis and quantification of the effects of the coupling between the ocean surface currents and wind stress as well as heat flux on air–sea interaction processes and the upper ocean responses by incorporating the improved drag coefficient as well as thermal exchange coefficient in the coupling system. The coupling between wind and ocean surface currents is carried out using the Earth System Modeling Framework (ESMF). The global distribution of the resulted changes in sea surface temperature and upper ocean mixed layer depth are presented and evaluated.

2. The Wind stress–Current Coupled System

2.1. The ocean current model

The HYbrid Coordinate Ocean Model (HYCOM) (Bleck, 2002; Bleck et al., 2002; Wallcraft et al., 2005) is configured to combine three vertical coordinate types (Z-levels, isopycnal layers and σ-levels) into a hybrid coordinate system. It is isopycnal in the open, stratified ocean, but uses the layered continuity equation to make a dynamically smooth transition to a terrain-following coordinate in shallow coastal regions, and to Z-level coordinates in mixed layer and/or unstratified seas. The hybrid coordinate extends the geographic range of applicability of traditional isopycnic coordinate circulation models toward shallow coastal seas and unstratified parts of the world ocean. It maintains the significant advantages of an isopycnal model in stratified regions while allowing higher vertical resolution near the surface and in shallow coastal areas, hence providing a better representation of upper ocean physics. HYCOM is designed to provide a major advance over the existing operational global ocean prediction systems, since it overcomes design limitations of the single coordinate systems as well as limitations in vertical and horizontal resolution. The resulting hybrid system should be a more streamlined system with improved performance and an extended range of applicability. Barnston et al. (1999) compared 8 dynamical and 7 statistical models and concluded that properly forecasting *ElNiño* and *LaNiña*, particularly the transition between them remains a challenge. Kara et al. (2008) successfully captured the westward extent of the SST cooling within the eastern equatorial Pacific during the transition period from the 1997 *ElNiño* to 1998 *LaNiña* one of the largest short term events ever observed, by using HYCOM, suggesting that HYCOM shows some advantages in comparison with some of the traditional coordinate ocean models.

2.2. The ocean surface current-dependent wind stress and heat flux algorithm

We first seek a parameterization for C_d assuming that this quantity depends only on the value of wind speed at the height of 10 m above the sea surface. Donelan et al. (1997) presented an algorithm for wind stress calculation:

$$C_d = 10^{-3}(0.07|\vec{U}_{10}| + 0.95), \tag{1}$$

$$\tau = \rho_a C_d |\vec{U}_{10}| \vec{U}_{10}, \tag{2}$$

where $\vec{U}_{10}$ is 10-m wind speed. By considering the effect of ocean surface currents (Xie and Rooth, 1994), (1) and (2) became

$$C_{d^*} = 10^{-3}(0.07|\vec{U}_{10} - \vec{U}_C| + 0.95), \tag{3}$$

$$\tau_* = \rho_a C_d |\vec{U}_{10} - \vec{U}_C| \vec{U}_{10} - \vec{U}_C), \tag{4}$$

where $\vec{U}_C$ is ocean surface current velocity and ρ_a is air density. The value of U_C can be obtained from HYCOM, while the wind stress τ_* is provided as a forcing to HYCOM. It is also important to note that, the relative wind speed $\vec{U}_{10} - \vec{U}_C$ is also utilized in heat flux calculation. In HYCOM, the total ocean–atmosphere heat flux H_{tot} consists of sensible heat flux H_s, latent heat flux H_l and net radiation flux H_r:

$$H_{tot} = H_s + H_l + H_r, \tag{5}$$

$$H_s = C_{Pair} E_x (T_s - T_a), \tag{6}$$

$$H_l = E_x L (H_u - E_V), \tag{7}$$

$$E_x = \rho_a C_T W, \tag{8}$$

where C_{Pair} is the specific heat of air, T_s is sea surface temperature, T_a is the air temperature in the atmospheric boundary layer, L is latent heat of vaporization, H_u is specific humidity, E_v is evaporation, E_x is the exchange coefficient, ρ_a is air density, C_T is heat transfer coefficient and W is wind speed. In our coupled model, W refers to the relative wind velocity $\vec{U}_{10} - \vec{U}_C$, which includes the effects of ocean surface currents. The heat transfer coefficient C_T can be partitioned into two parts on the basis of Monin–Obukhov similarity (MOS) theory,

$$C_d = c_d^{1/2} * c_d^{1/2}, \tag{9}$$

$$C_T = c_t^{1/2} * c_d^{1/2}, \tag{10}$$

and C_T now is a function of C_d. On neutral condition, c_t is expressed as (Panofsky and Dutton, 1984; Geernaert, 1990),

$$c_t^{1/2} = \frac{a\kappa}{\log(z/z_{OT})}, \tag{11}$$

according to Coupled Ocean–Atmosphere Response Experiment air–sea flux algorithm COARE3.0 (Fairall et al., 1996, 2003), z_{OT} is given by

$$z_{OT} = \min[1.1\times10^{-5}, 5.5\times10^{-5}(z_O u_*/\upsilon)^{-0.6}], \tag{12}$$

$$z_O = \frac{\alpha u_*^2}{g} + \frac{0.11\upsilon}{u_*}$$

$$\alpha = \begin{cases} 0.011 & U_{10} < 10\ \text{m/s} \\ 0.011 + 8.75\times10^{-4}(U_{10} - 10) & 10 < U_{10} < 18\ \text{m/s}, \\ 0.018 & U_{10} > 18\ \text{m/s} \end{cases} \tag{13}$$

$$u_*^2 = C_d * W^2, \tag{14}$$

where κ is *von Karman* constant (0.4), a the difference in scalar and velocity *von Karman* constants, z_{OT} the thermal roughness length, z_O the velocity roughness length, υ the kinematic viscosity for air and α the Charnock constant. Kara et al. (2005) obtained the heat flux exchange coefficient C_T, based on COARE3.0 algorithm, expressed as a polynomial function of wind speed and air–sea temperature difference and relative humidity,

$$C_T = \begin{cases} C_{T0}(W) + C_{T1}(W)(T_a - T_s) + C_{T2}(W)(T_a - T_s)^2 & -8 \leqslant T_a - T_s < -0.75\,^\circ\text{C} \\ C_{T0}(W) + C_{T1}(W)(T_a - T_s) & -0.75 \leqslant T_a - T_s < 0.75\,^\circ\text{C}, \\ C_{T0}(W) + C_{T1}(W)(T_a - T_s) + C_{T2}(W)(T_a - T_s)^2 & 0.75 \leqslant T_a - T_s < 7\,^\circ\text{C} \end{cases} \tag{15}$$

where T_a is air temperature and T_s sea surface temperature. In relation (15), all of the parameters C_{T0}, C_{T1} and C_{T2} are functions of wind speed W, which are detailed and specified by Kara et al. (2005) (Table 1). As indicated by (3), (9), (10), (11), and (15), C_T and C_d are in keeping with the MOS relationship and both vary with wind speed. The advantage of using above polynomial-based thermal exchange coefficient is that, as proposed by Kara et al. (2005), it agrees well