Ocean current and wave effects on wind stress drag coefficient over the global ocean

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[1] The effects of ocean surface currents and dominant waves on the wind stress drag coefficient ($C_D$) are examined over the global ocean. Major findings are as follows: (1) the combination of both ocean wave and current speeds can result in reductions in daily $C_D$ (>10%), but the notable impact of the latter is only evident in the tropical Pacific Ocean; (2) the presence of waves generally makes winds weaker and $C_D$ lower almost everywhere over the global ocean; (3) strong ocean currents near the western boundaries (Kuroshio and Gulf Stream) do not substantially influence $C_D$ since the winds and currents are not always aligned; and (4) the change in speed used in bulk flux parameterization also causes large changes in fluxes. Globally, the combined outcome of ocean currents and waves is to reduce $C_D$ by about (2%), but spatial variations (0% to 14%) do exist. Citation: Kara, A. B., E. J. Metzger, and M. A. Bourassa (2007), Ocean current and wave effects on wind stress drag coefficient over the global ocean, Geophys. Res. Lett., 34, L01604, doi:10.1029/2006GL027849.

1. Introduction

[2] The momentum exchange through wind stress at the atmosphere and ocean interface is of importance for many purposes, including air–sea interaction studies, climate studies, ocean modeling, and ocean prediction on various time scales. The total wind stress magnitude ($\tau$) at the ocean surface is typically calculated from the square of the wind speed at 10 m above the sea surface ($V$), the density of air ($\rho_a$), and a dimensionless drag coefficient ($C_D$) using $\tau = \rho_a C_D V^2$. [Fairall et al., 2003]. Turbulent energy fluxes are proportional to $V$. The change in fluxes due to the change in $V$ is easily estimated. The dependence of $C_D$ on sea surface currents and ocean waves will be examined herein.

[3] Possible impacts of ocean currents and wind waves on $V$ and $C_D$ were discussed in both theoretical studies [e.g., Hwang, 2005], and various regions of the global ocean [e.g., Wuest and Lorke, 2003]. In regions of strong currents (e.g., Kuroshio and Gulf Stream), it may not be simply the wind speed that is important for determining $C_D$, but the difference in near–surface winds and surface ocean currents. On the other hand, based on the authors’ knowledge there is no quantitative study examining spatial and temporal variability of wind and wave effects on $C_D$ over the entire global ocean. Such an investigation is essential because climate studies are often concerned about large–scale processes. Given the strong sensitivity of $C_D$ to water vapor effects at very low wind conditions [e.g., Kara et al., 2005], one would also need to determine the role of ocean currents and waves at these very low wind conditions.

[4] It may be important to take ocean current and wave effects into account in determining $C_D$ over the global ocean. However, experimental measurements for ocean currents and waves are rarely available and those that are available do not have sufficient temporal and spatial resolution to determine their global distribution. Local process studies over many parts of the global ocean often exclude such current and wave effects on wind stress through $C_D$. In some regions, such effects might be so small (i.e., weak ocean currents and negligible wave heights) that they can be considered insignificant. If so, there is no need to include the impact of such factors, eliminating the need to obtain local current speed and wave height information at a specific time and place.

[5] Examining the spatial and temporal distribution of the influence of ocean currents and waves on $C_D$ requires reliable global datasets. The quality of readily available archived numerical weather prediction (NWP) products, such as European Centre for Medium–Range Weather Forecasts (ECMWF) and the Navy Operational Global Atmospheric Prediction System (NOGAPS) has greatly improved since 1990s. They even provide high temporal resolution (e.g., 3–6 hourly) output over the global ocean. Thus, using the near surface meteorological variables from the existing NWP centers, $C_D$ including air–sea stability can be determined. As to waves (i.e., significant wave height, dominant period, etc.), and ocean currents (speed and direction), their global coverages can also be obtained from wave models and ocean general circulation models (OGCMs) at high temporal resolution (see section 2).

[6] Given the need for a quantitative analysis of the impact of ocean currents and waves on $C_D$ over the global ocean, the main focus of this paper is two–fold: (1) to present spatial variations of daily and monthly mean changes in the wind speed at 10 m above the sea surface and corresponding $C_D$ when including vector averages of ocean currents and waves, and (2) to determine regions in the global ocean where surface currents and waves can have significant influence on $C_D$.

2. Methods and Data

different formulations, Bonekamp et al. [2002] found that either a wave–age or wave–steepness dependent Charnock parameter was marginally superior to a linear dependence on wind speed. However, both sea state parameterizations perform better than linear bulk formula for most wind sea–dominant data sets. In fact, the mean is fairly well represented by the bulk formula, while the variability may not.

In this paper, a bulk parameterization that takes full account of stability in calculating $C_D$ is used. Such a parameterization is presented by Kara et al. [2005]. It is based on the state–of–the–art Coupled Ocean–Atmosphere Response Experiment (COARE) bulk algorithm (version 3.0), employing a turbulence theory based on the iterative estimations of the scaling variables to determine stability–dependent $C_D$ [Fairall et al., 2003]. $C_D$ is expressed as polynomial functions of air–sea temperature difference, using air temperature at 10 m, $V_a$ at 10 m and relative humidity at the air–sea interface to include air–sea stability. Due to deficiencies in the COARE algorithm itself at high winds, a constant $C_D$ is used in the parameterization for winds >20 m s$^{-1}$.

It is normally assumed that the stress direction is equal to the wind direction; however, both currents and waves can modify the stress direction [Grachev et al., 2003; Drennan and Shay, 2006; Bourassa, 2006]. We account for this directional change in calculation of the magnitude of the stress.

Global data sets used for calculating $C_D$ are as follows: (1) Near–surface atmospheric variables including 10 m wind speed from 1° × 1° NOGAPS; (2) wave information from 1° × 1° Wave Watch 3 (WW3), a third generation wave model; and (3) ocean currents from an eddy–resolving 1/12° × 1/12° cos (latitude) OGCM, the HYbrid Coordinate Ocean Model (HYCOM). Details of all data are publicly available online, https://www.fnmoc.navy.mil/PUBLIC/ for data sets 1 and 2, and http://hycom.rsmas.miami.edu/ for data set 3. Simulated ocean currents were binned to 1° squares, so that $C_D$ could be calculated on

![Figure 1](image1.png)

**Figure 1.** Spatial variations of wind drag coefficient over the global ocean on 1 Aug 2005 (00Z). Note values in the color bar must be multiplied by 10$^{-3}$. Effects of ocean currents and waves are excluded in calculating these drag coefficients. The regions where ice exists are masked out (shown in gray).

![Figure 2](image2.png)

**Figure 2.** Daily snapshot and monthly mean of (from top to bottom) wind speed at 10 m above the sea surface, ocean current speed and dominant wave speed based on the orbital velocity (see text for calculations) at (a) 1 Aug 2005 (00Z), and (b) Aug 2005. The ice mask (gray) at high latitudes (which is not the focus of this study) is based on the NOAA ice climatology in August.
the same grid. The binning was necessary to have a consistency in grid resolutions of each data set.

3. Impact of Currents and Waves on the Drag Coefficient

[11] In order to explore possible influences of ocean currents and waves on $C_D$, in the COARE–based $C_D$ parameterization (see section 2), $V$ is simply replaced by vector averages of $V-VC$ ($V-VW$), providing an insight of the effects of current speed (wave speed). Here, $V$ is the wind speed relative to the sea surface, $VC$ is the ocean current speed at the sea surface, and $VW$ is the wave speed. For simplicity, we drop vector notation from each term. Note that the vector averages are formed after $V$, $VC$ and $VW$ are decomposed to their components in directions. $V$ is also replaced by $V-VC-VW$ to determine the impact of both currents and waves on $C_D$ at the same time.

[12] Using the data sets (section 2), $V$ and $VC$ values are used directly from NOGAPS and HYCOM, respectively. $VW$ is calculated using data from the WW3 model. Following Bourassa [2006], $VW$ is expressed as 0.8 $V_{orb}$. The orbital velocity ($V_{orb} = 3.14 H/T$) is based on significant wave height ($H$) and dominant wave period ($T$). We obtain $V$, $VC$ and $VW$ at each $1° \times 1°$ grid point over the global ocean.

[13] $C_D$ is first calculated based solely on $V$ (i.e., without including effects of $VC$ and $VW$). Calculations are performed at each 3 hourly time interval at each grid for a given. As an example, Figure 1 shows how variable $C_D$ can be over the global ocean for a given day, at 00Z on 1 Aug 2005. $C_D$ has generally a value of $<1.0 \times 10^{-3}$ in the eastern tropical Pacific, and $>1.6 \times 10^{-3}$ at high southern latitudes and in the Indian Ocean at this particular time. Magnitude of $V$ values used for calculating $C_D$ are provided in Figure 2, along with means in Aug 2005. Large $C_D$ values (Figure 1) generally correspond to regions having high $V$ over the global ocean.

[14] $VC$ and $VW$ are generally very small ($<1\text{ m s}^{-1}$) in comparison to $V$, and this is evident from both the daily snapshot and monthly mean values (Figures 2a and 2b). $VC$ variability may seem to be noisy, but note that they are from a fine resolution eddy–resolving OGCM (section 2). Relatively large $VC$ ($>1\text{ m s}^{-1}$) are seen in the central tropical Pacific. Currents are also strong in the Kuroshio and Gulf Stream, having speeds of ($>1\text{ m s}^{-1}$). The binning of current speed (from $1/12°$ resolution to $1°$) also resulted in losing the actual strength of some OGCM–based currents at these two regions. As to $VW$ shown in (Figures 2a bottom and 2b bottom), they are generally weak in regions of the western boundary currents. This is because of relatively small wave heights (not shown).

[15] Realizing the large spatial variability in wind, ocean current and wave speed, we now focus on daily and monthly mean changes in $V$ and $C_D$ (Figure 3) when including VC and VW in $V$. The outcome of adding vector averages of VC components themselves to $V$ components is small over the global ocean except the tropical Pacific where VC is relatively large and its components generally have same the direction as $V$. Overall, including VC generally results in a decrease of $\approx 20\%$ in the central equatorial Pacific on 1 Aug 2005 (Figure 3a), but changes in daily $C_D$ are relatively small (Figure 3b). In reference to a given daily $V$ value of $8\text{ m s}^{-1}$ (see Figure 1), the 20% reduction translates to a $V$ value of $\approx 6.5\text{ m s}^{-1}$, entering $C_D$ calculation in this particular region. The impact of including wave effects on daily $V$ (i.e., $V-VW$), and hence $C_D$, is much larger than that of current effects (i.e., $V-VC$) over many parts of the global ocean. The overall influence of daily $VC$ ($VW$) is to reduce $V$ by $1.0\%$ ($5.4\%$) globally. The corresponding decrease in daily $C_D$ is small, 0.3% (1.7%).

[16] The combination result of adding daily VC and VW to $V$ (i.e., $V-VC-VW$) is further to reduce $V$ (e.g., $>20\%$), but relatively less for $C_D$ over a large extent of the global ocean (Figures 3a, bottom; and 3b, bottom). For example, a $C_D$ value of $\approx 1.2 \times 10^{-3}$ in the central equatorial Pacific
(see Figure 1) reduces by $\approx 10\% \ (1.1 \times 10^{-3})$. Because a vector averaging is performed for $V-V\cdot C-V\cdot W$ using horizontal and vertical components of the each term, a consistent increase or decrease in the final result that may be evident the individual $V-V\cdot C$ and $V-V\cdot W$ fields should not be expected. For example, there is almost no change in the daily $V-V\cdot C$ case and $\approx -5\%$ change for the $V-V\cdot W$ case in the northern Indian ocean, but the resulting $V-V\cdot C-V\cdot W$ can even be $\approx -10\%$. This is also reflected in $C_D$ when using $V-V\cdot C$ and $V-V\cdot W$ in the $C_D$ parameterization at the same region.

Insights gained from examining the impact of daily $V$ and $W$ on both $V$ and $C_D$ are extended to monthly time scales, and again this is done during northern (southern) hemisphere summer (winter) over the global ocean (Figures 3c and 3d). In comparison to values on 1 Aug 2005, monthly mean change in $C_D$ can usually be ignored at mid-latitudes, while a reduction of $\approx 5\%$ is noted at other places.

One thing to emphasize is that the data sources used here (NOGAPS, WW3 and HYCOM) have their unique errors as do other similar data sources. Thus, one might argue that there is not nearly enough information to make an estimate as to whether the wind errors from using these data source would not swamp out any signal in $C_D$ changes shown in this paper. On the other hand, they are considered good enough for our purposes. For example, comparisons of 1291 month–long wind speed time series from NOGAPS (used in this paper) and ECMWF with respect to those at mooring buoy locations gives a median wind speed bias of 0.62 m s$^{-1}$ and 0.53 m s$^{-1}$, respectively (not shown).

4. Conclusion

Currents and waves can cause substantial changes in the drag coefficient on daily time scales. This may explain some of the observed scatter in the measured $C_D$ reported in the literature. For the calculation of surface turbulent fluxes via bulk parameterizations, the change in $V$ is much more important; however, the change in $C_D$ is non negligible when integrated over time. Heat and moisture transfer coefficients have little dependence on $V$, except through changes in atmospheric stability; therefore changes in these coefficients are expected to be less than changes in $C_D$. Monthly averaged speed is typically reduced by 5%, resulting in 5% reduction in heat fluxes, and 10% in stress. Such effects can subtly influence the performance of ocean-only, coupled ocean–atmosphere or ocean-wave–atmosphere climate models.

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