3.5 PARAMETERIZATION OF WIND GUSTINESS FOR THE COMPUTATION OF OCEAN SURFACE FLUXES AT DIFFERENT SPATIAL SCALES

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1. INTRODUCTION

The atmosphere and ocean interact through the exchange of surface fluxes of heat, fresh water, and momentum. Surface fluxes are computed using bulk methods in numerical models and for data analyses. Various bulk algorithms have been developed in the past several decades, but most of them deal with fair weather conditions. On the other hand, a number of studies have shown the significant modulation of surface fluxes by precipitation convection using observational data over the eastern tropical North Atlantic and over the western Pacific warm pool region.

In order to consider the above mesoscale enhancement in surface flux parameterization, Redelsperger et al. (2000) reported their preliminary parameterization of mesoscale gustiness as a function of convective precipitation rate or cloud mass flux. Williams (2001) also developed a simple theoretical model for moist convective gustiness based on convective rainfall rate and ambient thermodynamic structure for the lower atmosphere. The purpose of this work is to develop a comprehensive parameterization scheme of wind gustiness that is caused by atmospheric boundary layer large eddies (under unstable conditions), convective precipitation, and convective cloudiness at different spatial scales. This scheme can then be used in bulk algorithms (e.g., Zeng et al. 1998) to compute ocean surface fluxes in weather and climate models.

2. RESULTS

a. Cloud-resolving model

Our new scheme is developed primarily by analyzing the Goddard cloud-ensemble (GCE) model output forced by observational data over different climate regimes. GCE is a fine-scale cloud-resolving model (Tao and Simpson 1993; Simpson and Tao 1993). The COARE bulk algorithm (Fairall et al. 1996) is used to compute ocean surface fluxes. GCE is used here to understand and parameterize the mesoscale effect of convective precipitation and clouds on surface fluxes, similar to its use for developing and testing cumulus parameterizations.

We have analyzed data from two-dimensional (2-D) GCE simulations with 512 horizontal grid points with a grid spacing of 1 km for three cases: the fast-moving, slow-moving, and non-organized convection over the eastern tropical North Atlantic (September 1-8, 1974) (denoted as the GATE case) and the squall lines and convective systems prior to and during westerly wind bursts over the western Pacific warm pool region (December 10-17, 1992; December 19-27, 1992) (denoted as COARE1 and COARE2 cases, respectively). The model performance for the COARE1 and COARE2 cases is discussed in Tao et al. (2000). The average vector wind speed at a height of 37 m (i.e., the first model level above surface) is lowest for the GATE case (1.0 m s^{-1}) and highest for the COARE2 case (4.9 m s^{-1}).

b. Parameterization of Wind Gustiness due to Boundary Layer Large Eddies, Convective Precipitation, and Cloudiness

For a model grid box (e.g., approximately $2.8^\circ \times 2.8^\circ$ for a T42 global spectral model), average ocean surface fluxes are computed from (e.g., Zeng et al. 1998):

$$\overline{LH} = \overline{\rho_a} L_e C_h U [\overline{q_s} - \overline{q_a}], \qquad (1)$$

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$$\overline{SH} = \overline{\rho_a} c_p C_h U[\overline{\theta_s} - \overline{\theta_a}], \quad (2)$$

$$\overline{\tau_v} = \overline{\rho_a} C_d U U_v, \text{ and} \quad (3)$$

$$\overline{\tau_s} = \overline{\rho_a} C_d U^2 \quad (4)$$

where the overbars denote averaging over a model grid box. LH and SH are latent and sensible heat fluxes, respectively, τ_s and τ_v are scalar and vector mean wind stresses, respectively, C_h and C_d are the turbulent exchange coefficients for heat (or moisture) and momentum, respectively, ρ_a is air density, L_e is latent heat of vaporization, c_p is specific heat of air, q_a and θ_a are the near-surface air humidity and potential temperature, respectively, and θ_s and q_s are surface potential temperature and saturated specific humidity, respectively. The near-surface vector (U_v) and scalar (U) mean wind speeds are

$$U_v = (\overline{u}^2 + \overline{v}^2)^{1/2}, \text{ and} \quad (5)$$

$$U = \overline{(u^2 + v^2)^{1/2}} = (U_v^2 + U_g^2)^{1/2} \quad (6)$$

where u and v are the two horizontal components of near-surface wind vector, and U_g is the wind gustiness. While an atmospheric model would provide $\overline{u}$ and $\overline{v}$, and hence U_v, it does not provide the scalar wind U so that the wind gustiness U_g needs to be parameterized. Using the GCE domain (of 514 km) to represent a global model grid box, the overbars in (1)−(6) then refer to averages over all GCE grids (at 1 km) except for the scalar mean wind U. Because boundary layer large eddies are largely unresolved by GCE with a grid size of 1 km, GCM includes the gustiness parameterization of Fairall et al. (1996), which is similar to (7) (to be discussed later), in the computation of scalar mean wind at each 1 km grid box. Therefore, the mean scalar wind U in (6) over the GCE domain is computed as the average of scalar wind speed at each GCE grid box.

For most global models, the wind gustiness under unstable conditions is parameterized as

$$U_{gb} = \beta w_* = \beta(\frac{g}{\theta_v}\overline{\theta_v' w'} z_i)^{1/3} \quad (7)$$

where w_* is the convective velocity scale, g is acceleration due to gravity, z_i is the convective boundary layer height, θ_v is virtual potential temperature, and $\overline{\theta_v' w'}$ is the surface buoyancy flux. The value of z_i is taken as 1000 m and β as unity in (7) for easy implementation in global models (e.g., Zeng et al. 1998), and Redelsperger et al. (2000) gave additional discussion on the value of β. This accounts for the contribution of large eddies in the convective boundary layer to surface fluxes and the wind gustiness is denoted as U_{gb}. Note that the buoyancy flux in (7) depends on the scalar mean wind [see (1) and (2)], and (1)−(7) need to be solved iteratively.

When convective precipitation occurs, Redelsperger et al. (2000) parameterized the wind gustiness as

$$U_{gr} = \log(1 + 6.7R - 0.48R^2) \quad (8)$$

to fit their cloud-resolving model output. Equation (8) accounts for the mesoscale contribution of convective precipitation (through the occurrence of downdrafts and updrafts) to surface fluxes, and the wind gustiness is denoted as U_{gr}. R is precipitation rate in cm day^{-1}, and U_{gr} in m s^{-1} in (8). As the precipitation rate approaches zero, the wind gustiness computed from (8) approaches zero rather than a non-zero gustiness from (7) which was used by Redelsperger et al. (2000) under fair weather conditions only. To maintain a smooth transition between rain versus no rain, here we parameterize the total (i.e., boundary layer and mesoscale) wind gustiness as

$$U_{gt} = \sqrt{(U_{gb})^2 + (U_{gm})^2} \quad (9)$$

where U_{gm} represents the mesoscale gustiness due to convective precipitation and cloudiness. Just as convective precipitation, convective cloudiness could be associated with mesoscale gustiness through two possible mechanisms: updrafts associated with non-precipitating convective clouds could generate (albeit weak) local circulations; both convective cloudiness and mesoscale wind variability take a long time to decay after convective precipitation ends. Because the impact of clouds on grid-box averaged variables is already considered in a numerical model, it is reasonable to assume that the impact of cloudiness on U_{gm} is symmetric to cloud fraction f_c of 0.5. For instance, stratus clouds with 100% cover would not affect gustiness here. Therefore we parameterize U_{gm} in (9) as

$$U_{gm} = \min[3, \max(2.4\, R^{1/2},\ 1.8\, f_c^{1/3})]. \quad (10)$$

where precipitation rate R is in mm h^{-1} and $f_c = \min(f_c, 1 - f_c)$ is the cloud fraction that is symmetric to 0.5. Based on the above formulation, the maximum U_{gm} due to cloudiness alone is 1.4 m s^{-1} for $f_c = 0.5$, which is equivalent to the U_{gm} value at R = 0.35 mm h^{-1}. Therefore, when R is smaller than 0.35 mm h^{-1}, the impact of cloudiness could become very important. The coefficients and exponents in (10) are chosen to fit the GCE data for the COARE2 case and are then tested for the COARE1 and GATE cases. Further work is still needed to determine whether (10) [or (8)] is applicable over other regions.

c. Dependence of the Parameterization on Spatial Scales

Our parameterization was developed for the GCE domain of about 500 km. An obvious issue is the scale dependence of our scheme or other schemes (e.g., Williams 2001; Redelsperger et al. 2000; Emanuel and Zivkovic-Rothman 1999; Zulauf and Krueger 1997). The dependence of sea surface fluxes on horizontal scales has been demonstrated before (e.g., Sun et al. 1996). Mahrt and Sun (1996) and Vickers and Esbensen (1998) have also proposed empirical formulations to compute wind gustiness as a function of horizontal scales directly (without explicitly considering precipitation rate or cloudiness). There are two approaches to address this issue: one is to develop different set of coefficients in (10) for each horizontal scale (or model grid size); the other is to multiply our parameterization by a horizontal-scale-dependent function. The latter approach is adopted here for its simplicity. Furthermore, because U_{gb} represents the contribution to wind gustiness of boundary layer large eddies with a horizontal scale of about 2 km, it is reasonable to consider the dependence of U_{gm} alone on horizontal scales. For each case, we use the GCE data to compute U_{gm} averaged over horizontal scales from 10 km to 510 km. Then for each horizontal scale L, U_{gm} is averaged for the whole integration period. Furthermore, these scale-dependent U_{gm} values are normalized by their maximum value (taken as an average of U_{gm} values for L = 410 to 510 km). Our analysis of data from these three cases shows that overall the normalized U_{gm} value increases more rapidly with L when L is relatively small. Therefore, a simple scale-dependent function can be developed:

$$F(L) = (L/400)^{0.37} \qquad (11)$$

with L in km and under the restriction of $0 \leq F(L) \leq 1$. Then (10) multiplied by (11) gives our parameterization of U_m that depends on convective precipitation rate, convective cloudiness, and horizontal scales.

In summary, our scheme for wind gustiness can be written as

$$U_g = \sqrt{(U_{gb})^2 + [F(L)U_{gm}]^2} \qquad (12)$$

where U_{gm} and F(L) are given in (10) and (11) respectively, and U_{gb} is given in (7) under unstable stratification conditions. For stable conditions, U_{gb} can be replaced by a small value (e.g., 0.1 m s^{-1}). Equation (12) can be easily implemented into weather and climate models with various horizontal resolutions.

Results have been published in Zeng et al. (2002), and will be presented in our presentation.

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