NOTES AND CORRESPONDENCE

Incorporation of Stratification Effects on the Oceanic Roughness Length in the Derivation of the Neutral Drag Coefficient*

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ABSTRACT

Based on the assumption that, over the sea, the roughness length of the wind profile scales with the wind stress, a new formulation that describes the drag coefficient as a function of the given neutral drag coefficient and stability is derived. The new formulation is compared to an earlier formulation where roughness changes with stability were ignored. The two are then illustrated with data collected from both the Marine Remote Sensing Project (1979) and the Tower Ocean Wave and Radar Dependence Experiment (1984). It was found that when the surface roughness was allowed to depend on wind stress (and therefore stability), the stratification correction to the neutral drag coefficient was larger than for the case when the roughness length was not allowed to vary.

1. Introduction

In most numerical models of air–sea interaction processes, the wind stress and heat flux are calculated from observed weather data using bulk coefficients. By employing the bulk aerodynamic relations, these fluxes are estimated as follows (Roll, 1965):

$$\tau = \rho C_D(\bar{U}_z - \bar{U}_0{}^2) \quad (1)$$

$$H = \rho C_p C_H(\bar{U}_z - \bar{U}_0)(\bar{T}_0 - \bar{T}_z) \quad (2)$$

$$E = \rho L_v C_E(\bar{U}_z - \bar{U}_0)(\bar{q}_0 - \bar{q}_z) \quad (3)$$

where τ, H and E, are the momentum, sensible heat and latent heat fluxes, respectively; ρ is the air density; c_p the specific heat of air at constant pressure; L_v the latent heat of vaporization; U windspeed; T temperature; and q the specific humidity. Here, C_D, C_H and C_E are, respectively, the drag coefficient, Stanton number and Dalton number. The subscript, z, indicates the measurement height above the surface.

Little is known about the variation of the heat exchange coefficients with wind speed and stratification (Smith et al., 1983); however, it is generally accepted that the drag coefficient depends considerably on these environmental variables (Liu et al., 1979). For unstable conditions, the drag coefficient, C_D, has been observed to increase with an increase in surface heat flux, and a general theory has been developed to explain this behavior (Businger et al., 1971). For stable conditions, there is no theory yet that adequately describes the mechanism of vertical momentum transfer, and only an empirical relationship is used for this regime. Experimental investigations of the behavior of the drag coefficient for different stratifications have been reviewed by Hsu (1974) and Launianen (1979). One of the primary objectives of such investigations has been to calculate the "neutral" drag coefficient, characterized by defining the equivalent neutral profile of wind speed given the observed wind speed as the common reference at height z. The wind speed used for bulk calculations of the wind stress is usually the value representative of the height of 10 m above the surface.

The method of calculating the neutral drag coefficient from a measured drag coefficient is to assume that the surface-layer wind profile depends upon the stratification and surface roughness, where the surface roughness depends upon the wave field. The observed weather data are the input parameters for models. In order to employ Eq. (1) to calculate the wind stress from these data, a value for the drag coefficient must be determined from the known neutral drag coefficient associated with the observed wind speed, and an adjustment must be made if the stratification differs from neutral. The equations relating the stratification-dependent drag coefficient to the neutral drag coefficient were derived by assuming that the roughness length, z_0, does not depend on surface wind stress (Fleagle and Businger, 1980). Over land, this is a good

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assumption, but over the sea, it is not because of the dependence of wave field on the surface stress.

In addition to its application to air and ocean models, the neutral drag coefficient is the standard drag parameter utilized for wind stress comparisons between different investigations. Since stability is normalized to the neutral case, C_{DN} variations additionally serve as a basis for examining the wave influence on wind drag studies.

2. Mathematical framework

By assuming a constant flux layer and horizontal homogeneity of the surface roughness and wind flow, the profile of wind speed may be written in the following form (Fleagle and Businger, 1980):

$$\bar{U}_z - \bar{U}_0 = \frac{u_*}{k}\left(\ln\frac{z}{z_0} - \psi\right) \tag{4}$$

where $\bar{U}_z$ and $\bar{U}_0$ are, respectively, the wind speed at the height above the surface, z, and the wind speed at the surface. The von Karman constant, k, assumes a magnitude of 0.40. The friction velocity, u_*, is related to the wind stress, τ, by the relation:

$$\tau = \rho u_*^2. \tag{5}$$

For land, the surface speed is assumed to be zero, while, for the ocean, U_0 may be assumed to be equivalent to the component of the surface current velocity in the direction of the wind (Geernaert, 1983). If the tidal stream velocity is small relative to the wind drift, U_0 has approximately the same magnitude as u_* (Hicks, 1972). Here, z_0 is the roughness length, and ψ is a stability parameter strongly dependent on the ratio, z/L, where L is the Monin–Obukhov length defined as follows:

$$L = \frac{-T_v u_*^3}{gk\overline{w'T_v'}} \tag{6}$$

where T_v is the virtual temperature, and g is the acceleration due to gravity.

The form of ψ according to Paulson (1970) is

$$\psi = 2\ln\left(\frac{1+x}{2}\right) + \ln\left(\frac{1+x^2}{2}\right) - 2\tan^{-1}(x) + \frac{\pi}{2} \tag{7}$$

where $x = (1 - 16z/L)^{1/4}$, if the atmosphere is unstable, i.e., $z/L < 0$. For stable stratifications, an empirical relation may be applied:

$$\psi = -5z/L. \tag{8}$$

Equations (1), (4) and (5), may be combined into the following form:

$$kC_D^{-1/2} = \ln\frac{z}{z_0} - \psi. \tag{9}$$

If one considers the case of neutral stratifications where $\psi = 0$, the drag coefficient and roughness length may be defined, respectively, to be C_{DN} and z_{0N}, where the subscript N indicates neutral stratification. From Eq. (9), we may then write

$$kC_{DN}^{-1/2} = \ln\frac{z}{z_{0N}}. \tag{10}$$

Combining Eqs. (9) and (10) yields the following general result:

$$C_D = \left(C_{DN}^{-1/2} - \frac{\psi}{k} + \frac{1}{k}\ln\frac{z_{0N}}{z_0}\right)^{-2}. \tag{11}$$

In interpreting (11), for momentum transfer to the sea, it must be remembered that C_D and D_{DN} refer to the same common windspeed at the given reference level. Because wind stress varies according to stability given the same reference windspeed, the sea state and consequently the roughness length, z_0, must also vary.

3. Special cases

The general form exhibited by Eq. (10) may be applied to two special cases: 1) the case of an underlying surface where roughness elements have a structure and appearance independent of the wind stress; and 2) the case of an underlying surface where roughness elements depend on the imposed wind stress. For case 1, the neutral roughness length may safely be assumed to be equal to the roughness length, i.e., $z_{0N} = z_0$, and equation (11) reduces to the following form:

$$C_D = (C_{DN}^{-1/2} - \psi/k)^{-2}. \tag{12}$$

This equation may generally be applied over land. But it has also regularly been applied to models of the atmosphere over water where its application may not be appropriate.

For case 2, we may assume that the magnitude of the roughness length is proportional to the imposed wind stress, as was suggested by Charnock (1955) for the ocean. The Charnock relation is

$$z_0 = \alpha u_*^2/g \tag{13}$$

where α is the Charnock constant. The application of the Charnock relation to the oceans has been supported by the composite analyses of Garratt (1977) and Wu (1980), for wind speeds above 5 m s^{-1}.

Although several authors (Smith, 1980; Large and Pond, 1981) disclaim the wide use of the Charnock formula as a predictive tool for z_0, the data presented in this note have been shown to be fitted better by a Charnock formulation than by a wind speed dependence (Geernaert, 1983).

Equation (13) may be rewritten in terms of C_D and C_{DN} as follows:

$$z_0 = \alpha u_*^2/g = \alpha C_D(U_{10} - U_0)^2/g \tag{14}$$

$$z_{0N} = \alpha u_{*N}^2/g = \alpha C_{DN}(U_{10} - U_0)^2/g. \tag{15}$$

All bulk exchange coefficients are hereafter representative of a 10 m sampling height of wind speed, temperature, and humidity. Substitution of Eqs. (14) and (15) into (11) results in the following:

$$C_D = \left(C_{DN}^{-1/2} - \frac{\psi}{k} + \frac{1}{k}\ln\frac{C_{DN}}{C_D}\right)^{-2}. \qquad (16)$$

This equation may be applied to locations over marine surfaces with 10-m height wind speed greater than 5 m s^{-1}. (Garratt, 1977).

4. Discussion

In the past, for drag measurements in the marine atmospheric surface layer, Eq. (12) has been applied toward correcting the observed drag coefficient for non-neutral conditions to its neutral counterpart, C_{DN}. Over the sea, both the drag coefficient and roughness length are known to show considerable wind speed and stress dependence (Kitaigorodskii, 1973; Geernaert et al., 1986a); therefore, application of Eq. (16) to oceanic data is consistent with observational evidence.

A comparison between Eqs. (12) and (16) was performed over the range of stabilities, $|z/L| < 1.0$, for two values of the neutral drag coefficient. These data are plotted in Fig. 1.

Equations (12) and (16) were also applied to field measurements of drag coefficients and atmospheric stratifications collected during the Marine Atmospheric Remote Sensing (MARSEN) experiment in the autumn of 1979. One of the purposes of the experiment was to determine a neutral drag coefficient parameterization applicable to the North Sea. The wind speed spanned 8 to 18 m s^{-1} with full wave and current information. There were 53 records of data, each of 30 min duration (Geernaert et al., 1986a). The plotted data points in Fig. 2 represent drag coefficient measurements observed over the range of stratifications: $0.00 > z/L > -0.16$. The regression lines are not best fits to these points. They are best fits to the calculated neutral drag coefficients, which were found after the stability correction was applied to the measured drag coefficients. The calculated C_{DN} are determined both by Eq. (12) and by Eq. (16). Consequently, two regression lines are drawn.

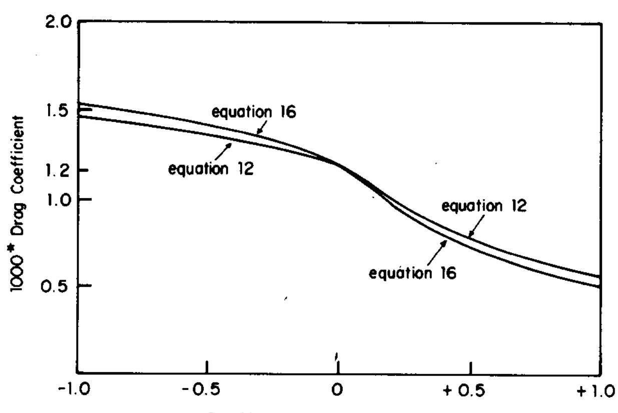

FIG. 1. Comparison of two calculations of the drag coefficient given the stability and a neutral drag coefficient of 0.0012. Equation (12) represents a roughness length that is independent of the wind stress, and Eq. (16) represents a roughness length that is defined according to the Charnock relation.

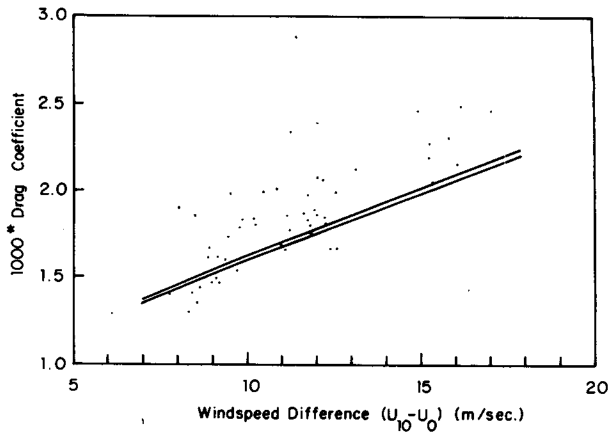

FIG. 2. Distribution of measured drag coefficients with the wind speed difference, $U_{10} - U_0$, obtained over the North Sea. Superimposed over this set are lines of the neutral drag coefficient based on best-fit Charnock coefficients for the two cases: (a) upper line is derived from Eq. (12) corresponding to constant z_0 and (b) lower line is derived from Eq. (16) corresponding to variable z_0.

By combining Eqs. (10) and (15), C_{DN} is related to the Charnock coefficient, α, as follows:

$$C_{DN} = \left[\frac{1}{k}\ln\frac{gz}{\alpha C_{DN}(U_z - U_0)^2}\right]^2. \qquad (17)$$

Equation (17) quickly converges to a solution of C_{DN} for a given Charnock coefficient, height and wind speed.

Because the representation of C_{DN} by Eq. (17) is based on a model with a given Charnock coefficient, C_{DN} using this formulation is called C_{DM} hereafter. Calculations of the best-fit Charnock coefficient for the data presented in Fig. 2 by minimizing the mean square error between C_{DM} and C_{DN} results in the following:

$$\alpha_1\ [\text{Eq. (16)}] = 0.0276 \pm 0.0055 \qquad (18)$$

$$\alpha_2\ [\text{Eq. (12)}] = 0.0292 \pm 0.0060. \qquad (19)$$

These values are larger than the Charnock coefficient suggested by Wu (1980), who found α to be 0.0186; Wu, however, determined the Charnock coefficient by ignoring the surface speed. The MARSEN data of Fig. 2, when U_0 was assumed equal to 0, provided a best-fit Charnock coefficient of 0.0192.

The results from Fig. 2 indicate no statistical difference between α_1 and α_2. Because the data represent generally near-neutral conditions, the difference between α_1 and α_2 corresponds to only a 2% difference in the magnitude of the predicted neutral drag coefficients. The mean square difference (msd) between the measured neutral drag coefficient defined according to Eq. (16) and the modeled drag coefficient according to Eq. (17) is 10^8 msd = 2.88; the best fit C_{DN} dependence on wind speed yields 10^8 msd = 2.89 (Geernaert et al., 1986a); i.e., for this dataset, the Charnock formulation is as good a predictor for the drag coefficient, as is the wind speed formulation.

5. Estimates of C_D from weather observations

In order to illustrate the application of Eqs. (12) and (16) for the purpose of estimating the wind stress, we used routine weather observations from a period when the stratification changed from unstable to stable during the course of one day. These observations were made at the Naval Ocean Systems Center (NOSC) tower during the Tower Ocean Wave and Radar Dependence (TOWARD) Experiment 1984 (Geernaert et al., 1986b) and included air and water temperatures, relative humidity, and wind velocity (Table 1). The atmospheric quantities are representative of a height of 10 m above the surface and the upwind fetch was at least 3 km during the period. The sea surface temperature was measured below the wave troughs. Here we calculate C_{DN} using the more typical version of Eq. (17) where $U_0 = 0$:

$$C_{DN} = \left(\frac{1}{k} \ln \frac{g\ 10\ \mathrm{m}}{\alpha C_{DN} U_{10}^2}\right)^2. \tag{20}$$

Equation (20) in the past has been utilized for flux estimates when the Charnock relation is assumed (Garratt, 1977; Wu, 1980). For comparison, the water column depth at the measurement site during MARSEN was 16 m, and the depth at the NOSC tower was 17 m.

The neutral drag coefficient was calculated for each record of hourly observations during TOWARD 84, using Eq. 20 and using $\alpha = 0.0192$ from the MARSEN data analysis. With the given air and water temperatures and humidity observations, the fluxes of sensible and latent heat were computed from bulk formulations (Smith, 1980; Anderson and Smith, 1981). Because the quantity, z/L, depends on the heat flux and wind stress, the calculation of C_D from D_{DN} and ψ by either (12) or (16) is an iterative process. Consequently, the convergence produces a solution including C_D and z/L when Eq. (16) is applied to each record of data that is different from the solution when Eq. (12) is applied. The estimates of C_D and z/L by both methods are tabulated alongside the observations in Table 1.

When the equation for variable roughness [Eq. (16)] is applied, the results indicate that the atmosphere is slightly more stable, i.e., z/L is larger, than if Eq. (12) is used. For unstable stratifications, the mechanism is that the larger C_D from (16) implies a larger u_* and consequently a stability closer to neutral. On the stable side, the C_D and u_* predicted from (16) are slightly smaller than when (12) is applied, and the surface layer is more stable with (16) than with (12).

Because the drag coefficient is known to vary much

TABLE 1. Hourly observations from the NOSC tower off San Diego during 20–21 October 1984. Drag coefficients and surface layer stabilities are calculated from the neutral drag coefficient [according to Eq. (18)] and weather data from Eqs. (12) and (16). For hours 11 and 17, the asterisk indicates that the drag coefficient asymptotically approaches zero while Z/L approaches infinity.

Hour (PST)	Wind speed (m s^{-1})	Temperature (°C) Air	Water	Relative humidity (%)	$10^3 \times C_{DN}$	$10^3 \times C_D$ (12)	z/L (12)	$10^3 \times C_D$ (16)	z/L (16)
21	2.1	17.9	19.1	66	0.78	0.98	−1.53	1.02	−1.44
22	2.9	17.4	19.1	69	0.87	1.06	−1.01	1.10	−0.95
23	4.0	16.7	19.1	71	0.98	1.16	−0.66	1.20	−0.63
00	3.8	16.4	19.1	73	0.96	1.16	−0.82	1.20	−0.78
01	3.6	16.2	19.1	75	0.94	1.15	−0.99	1.20	−0.93
02	3.4	15.8	19.1	77	0.93	1.16	−1.26	1.21	−1.18
03	3.4	15.7	19.1	75	0.93	1.16	−1.30	1.21	−1.21
04	3.2	15.5	19.0	77	0.91	1.15	−1.52	1.21	−1.42
05	3.0	14.9	19.0	78	0.88	1.16	−2.02	1.22	−1.87
06	3.7	14.2	19.0	78	0.95	1.21	−1.45	1.28	−1.35
07	3.8	15.0	19.0	80	0.96	1.20	−1.16	1.25	−1.09
08	3.2	17.0	19.0	66	0.91	1.10	−0.93	1.14	−0.88
09	1.5	18.4	18.9	64	0.71	0.87	−1.50	0.90	−1.42
10	4.1	20.0	18.9	52	0.99	0.74	0.36	0.66	0.42
11	5.4	21.8	18.9	48	1.10	0.63	0.69	*	*
12	7.2	21.8	18.8	45	1.24	1.04	0.19	0.98	0.21
13	7.0	21.8	18.8	54	1.23	1.01	0.21	0.94	0.23
14	7.0	21.7	18.8	58	1.23	1.01	0.20	0.96	0.22
15	6.5	21.8	18.8	55	1.19	0.92	0.28	0.84	0.32
16	5.7	21.6	18.8	56	1.13	0.77	0.44	0.63	0.60
17	5.2	21.5	18.8	57	1.08	0.42	0.62	*	*

more for stable than for unstable stratifications (Businger et al., 1971), the comparison between (16) and (12) produces larger differences when $z/L > 0$. As z/L becomes larger than 0.5, the differences between the two C_D and the ratio z/L become very dramatic. If variable roughness is included in the derivation of C_D from C_{DN}, the drag coefficient is larger for the unstable case and smaller for the stable case when compared to the method where the roughness length is not allowed to vary.

There are a few limitations to this analysis of the NOSC tower data. For these weather observations, the flux parameterizations produce estimates of $|z/L|$ that occasionally are larger than unity. At such conditions of strongly stable or unstable flow, the flux profile relations are less well defined (Launianen, 1979). In addition to this limitation, the application of Charnock's relation to wind speeds below 3.5 m s^{-1} may not be valid. Below 3.5 m s^{-1}, the characterization of z_0 with wind stress or wave state is not well established. However, the NOSC tower data from the TOWARD experiment illustrate the variations of stability that often occur near the coastline and the influence the variable roughness length has on estimating the drag coefficient.

When employing the exchange coefficients for heat and water vapor in order to obtain an estimate of L, one might consider that z_{0T} and z_{0E}, the roughness lengths in the logarithmic profiles of temperature and water vapor, should also be stratification dependent. The result would be a change in the estimated L and therefore an additional correction to C_D from C_{DN}. We have not taken that step in this analysis since we assume this to be a second-order correction, and very little is known about the effects of sea state on heat and vapor fluxes (for a discussion see Liu et al., 1979).

6. Conclusion

The derivation of the drag coefficient dependence on roughness and stratification examined here provides a more physically consistent approach to determining the value of the drag coefficient over the sea. Because the roughness over land does not usually change due to different imposed wind stresses, no improvement to analyses over land is suggested. The data analyzed here only illustrate that differences do occur if Eq. (12) instead of (16) is applied. For the calculation of the drag coefficient over the sea where atmospheric stabilities are far from neutral, such as cases of warm air advection or cold air outbreaks, the equation for variable roughness should be applied [Eq. (16)] (refer to Table 1); otherwise, a bias in the calculated wind stress may be incurred.

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