A Mechanism for the Increase of Wind Stress (Drag) Coefficient with Wind Speed over Water Surfaces: A Parametric Model

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ABSTRACT

A mechanism is proposed for a physical explanation of the increase in wind stress (drag) coefficient with wind speed over water surfaces. The formula explicitly incorporates the contribution of both winds and waves through the parameterization of an aerodynamic roughness equation. The formula is consistent with measurements from the field and with results obtained by numerical models for storm surges and water level fluctuations.

1. Introduction

Many investigators have demonstrated that the wind stress or aerodynamic drag coefficient, C_D, increases with wind speed over water surfaces, whether they are small lakes or open oceans. A summary of these findings during the last 10 years is given in Fig. 1. Although it can be seen that C_D increases generally with the wind speed, the rates of these increases vary greatly. Several efforts have been made to explain these variations, e.g., by Charnock's (1955) formulation (Garratt, 1977), by Froude number scaling (Wu, 1982), and by state of wave development (Donelan, 1982). However, a mechanism that can explain and estimate these variabilities satisfactorily is still needed.

The wind speed dependence of C_D has been sought based on Charnock's prediction (Garratt, 1977), but dependence upon other parameters such as fetch and wind duration should also be considered. Donelan's (1982) modeling further demonstrated that C_D is also dependent on the wave characteristics. A similar approach was suggested earlier by Hsu (1974), who demonstrated that the roughness length can be parameterized to include explicitly contributions of both winds and waves. He showed that Charnock's well-known relationship is a special case in his general roughness length relationship. Since Hsu's formulation has been further verified (Fig. 2) (see Graf et al., 1984), it is the purpose of this paper to explicitly incorporate wave characteristics into the roughness length parameterization, which can then be applied to obtain C_D. A model is thus proposed to explain the increase in drag coefficient with wind speed over water surfaces. Inputs for the model are the wind speed and the spectral wave height and peak spectral period according to JONSWAP formulation (Hasselmann et al., 1976).

2. The mechanism

In the atmospheric surface boundary layer, the mean wind profile can be described by similarity theory (e.g., Busch, 1973):

$$\frac{\kappa Z}{U_*}\frac{dU}{dZ} = \phi_m\left(\frac{Z}{L}\right) \tag{1}$$

where κ is the von Kármán constant (=0.40 ± 0.01, see Högström, 1985), Z the height above the surface, U_* the shear or friction velocity ($=(\tau/\rho)^{1/2}$, where τ is the wind stress and ρ the air density), U is the mean wind speed, and L the Monin-Obukhov stability length. Parameter Z/L may be derived by obtaining the differences in both temperature and humidity between the sea surface and the overriding air (see Large and Pond, 1982, Eq. 13).

If one adds and subtracts 1 on the right of (1) and integrates from the sea surface, where $Z = Z_0$ and $U = 0$, to an arbitrary height Z [note that in the atmospheric surface boundary layer, variation of U_* with Z is less than 10% (e.g., see Wyngaard, 1973) and thus U_* is treated as constant] (e.g., see Panofsky and Dutton, 1984):

$$U_Z = \frac{U_*}{\kappa}\left[\ln\frac{Z}{Z_0} - \psi_m\left(\frac{Z}{L}\right)\right], \tag{2}$$

$$\psi_m\left(\frac{Z}{L}\right) = \int_0^{Z/L}\left[1 - \phi_m\left(\frac{Z}{L}\right)\right]\frac{d\xi}{\xi} \tag{3}$$

in which

$$\xi = \frac{Z}{L}. \tag{4}$$

The functional relationships between ψ_m and Z/L under various stability conditions are given in Large and Pond (1982).

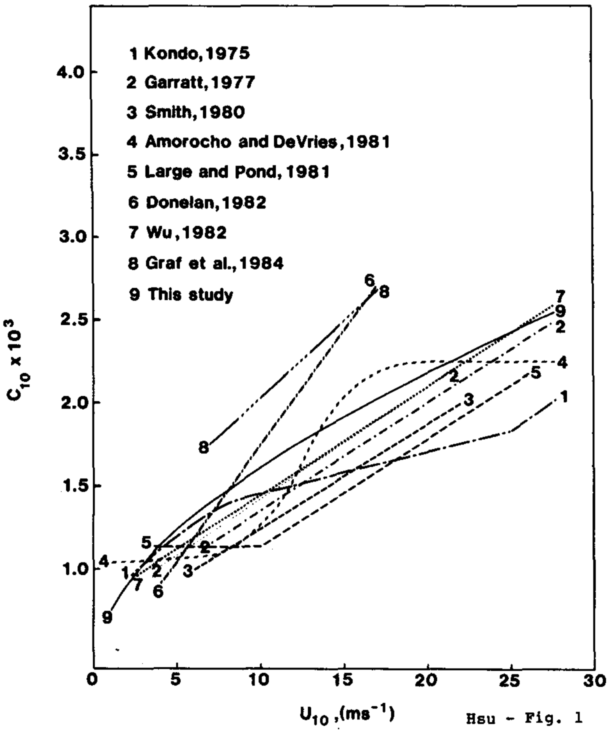

FIG. 1. Variation of the drag coefficient (C_{10}) with wind speed at 10 m above the water surface.

Note that under neutral or adiabatic conditions $L \to \infty$ or $Z/L \to 0$, and from (3)

$$\phi_m\left(\frac{Z}{L}\right) = \phi_m(0) = 1. \tag{5}$$

Thus, (2) can be reduced to the familiar logarithmic wind profile such that

$$U_Z = \frac{U_*}{\kappa} \ln \frac{Z}{Z_0} \tag{6}$$

where Z_0 is defined as the aerodynamic roughness length.

The wind stress or aerodynamic drag coefficient, C_D, is defined as [cf. (2)]

$$C_D = C_Z = \left(\frac{U_*}{U_Z}\right)^2 = \left[\frac{\kappa}{\ln(Z/Z_0) - \psi_m(Z/L)}\right]^2 \tag{7}$$

and under near-neutral conditions [cf. (6)]

$$C_{DN} = \left[\frac{\kappa}{\ln(Z/Z_0)}\right]^2. \tag{8}$$

Thus

$$C_D = C_{DN}[1 - \kappa^{-1} C_{DN}^{1/2} \psi_m(Z/L)]^{-2}. \tag{9}$$

Although the parameter Z_0 is obtained from integration of (1) when $U = 0$ at $Z = Z_0$, this parameter also has physical meaning. On the basis of K theory (e.g., Panofsky and Dutton, 1984)

$$\tau = \rho U_*^2 = \rho K_m \frac{\partial U}{\partial Z} \tag{10}$$

where K_m is the exchange coefficient of momentum, which is also called eddy viscosity.

Under steady state and near neutral conditions, (1) states that

$$\frac{\partial U}{\partial Z} = \frac{U_*}{\kappa Z}. \tag{11}$$

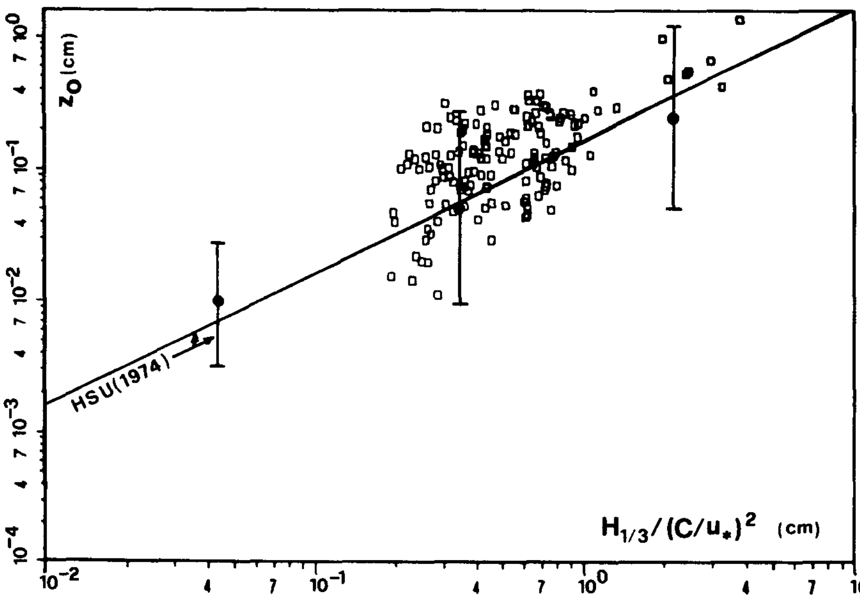

FIG. 2. The roughness length Z_0 in centimeters versus a wave-age parameter $H/(C/u_*)^2$ in centimeters. Plotted are the nearshore data "Buchillon" from Lake Geneva. The data are compared with Hsu's (1974) relation equation (17) and its scatter (after Graf et al., 1984).

According to Businger (1973), this expression indicates an infinite shear at the surface ($Z = 0$), which of course is unrealistic. In order to keep the shear finite, a surface roughness Z_0 is introduced such that

$$\frac{\partial U}{\partial Z} = \frac{U_*}{\kappa(Z + Z_0)}. \quad (12)$$

Integration yields

$$U_z = \frac{U_*}{\kappa} \ln\left(\frac{Z + Z_0}{Z_0}\right). \quad (13)$$

Since $Z \gg Z_0$, (13) becomes (6).

From (10) and (12) we have

$$K_m = \kappa U_*(Z + Z_0)$$

and at the surface

$$K_{m,\,\text{surface}} = \kappa U_* Z_0. \quad (14)$$

Equation (14) states that at the surface, eddy viscosity is taken to represent the product of eddy size and eddy velocity, and we see now that Z_0 represents eddy size at the surface. Clearly, the rougher the ground, the larger these eddies can be. Thus Z_0 is a measure of surface roughness. It is thus called the roughness length.

Parameterization of Z_0 over the water surface was first formulated on dimensional considerations by Charnock (1955):

$$Z_0 = a\frac{U_*^2}{g} \quad (15)$$

where a is the Charnock coefficient, assumed to be a constant.

On the other hand, it is reasonable (e.g., see Kitaigorodskii and Volkov, 1965; DeLeonibus and Simpson, 1972; Hsu, 1974; Coantic, 1978; Donelan, 1982) to assume that Z_0 depends upon the wave age C/U_* as well as upon the wave steepness H/L_w or

$$Z_0 = f\left(\frac{C}{U_*}, \frac{H}{L_W}\right) \quad (16)$$

where C is the wave celerity, H the wave height, and L_w the wave length. A relationship that exists between C/U_* and H/L_w (e.g., Brutsaert, 1982) states that, for large wave age, the wave steepness is small. For waves coming from deep water into shallow water, the wave age decreases and the wave steepness increases (e.g., Graf et al., 1984). Since the effect of atmospheric stability can be explicitly incorporated into the growth rate equation for the surface gravity waves (Janssen and Komen, 1985), the stability parameter is not included in the parameterization of Z_0.

On the basis of these considerations and many experimental results, Hsu (1974) proposed a relationship among these variables that includes both wind and wave contributions, such that

$$Z_0 = \frac{1}{2\pi}\left[\frac{H_{1/3}}{(C/U_*)^2}\right] \quad (17)$$

where $H_{1/3}$ is the significant wave height, defined as the average of the highest one-third waves and $H_{1/3}$ and C depend implicitly on L_W:

$$H_{1/3} = 4\sigma \quad (18)$$

where

$$\sigma^2 = \int_0^\infty E(\omega)d\omega \quad (19)$$

in which $E(\omega)$ is the wave energy spectrum. Note that σ is the standard deviation of the wave record (e.g., see U.S. Army Corps of Engineers, 1984).

Further validation of (17) is provided by Graf et al. (1984) as shown in Fig. 2.

Relationships of wave parameters before shoaling are

$$C = \frac{L_w}{T} \quad (20)$$

and

$$L_w = \frac{g}{2\pi}T^2, \quad (21)$$

so

$$C = \frac{g}{2\pi}T \quad (22)$$

where $T = f_m^{-1}$ is the wave period and f_m is the dominant frequency of the spectral peak.

Since waves are dependent on the duration of the wind, t, and its over-water trajectory, i.e., the fetch, F, relationships among these parameters are needed. In practice, a simple method for making wave estimates is desirable, but is possible only if the geometry of the water body is relatively simple and the wave conditions are either fetch limited or duration limited. Under fetch-limited conditions, winds have blown constantly long enough for wave heights at the end of the fetch to reach equilibrium. Under duration-limited conditions, the wave heights are limited by the length of time the wind has blown. These two conditions represent asymptotic approximations to the general problem of wave growth. In most cases the wave growth pattern at a site is a combination of the two cases. Equations (23) to (28) are obtained from U.S. Army Corps of Engineers (1984) by simplifying the equation used to develop the parametric model based on JONSWAP experiments and others (Hasselmann et al., 1976).

In the fetch-limited case, the parameters required are the fetch, F, and the adjusted wind speed, U_A, as described in Chapter 3, Section IV, of U.S. Army Corps of Engineers (1984), and represent a relatively constant average value over the fetch. The adjusted wind speed represents that the wind speed has been adjusted in height to the standard 10 m above the mean sea surface and in stability to account for the air–sea temperature difference. The spectral wave height H_{m0} and peak spectral period T_m are the parameters predicted.

$$\frac{gH_{m0}}{U_A^2} = 1.6 \times 10^{-3}\left(\frac{gF}{U_A^2}\right)^{1/2} \quad (23)$$

$$\frac{gT_m}{U_A} = 2.857 \times 10^{-1}\left(\frac{gF}{U_A^2}\right)^{1/3} \quad (24)$$

$$\frac{gt}{U_A} = 6.88 \times 10^{1}\left(\frac{gF}{U_A^2}\right)^{2/3}. \quad (25)$$

Note that $T_{1/3}$ is given as $0.95T_m$. The preceding equations are valid up to the fully developed wave conditions given by

$$\frac{gH_{m0}}{U_A^2} = 2.433 \times 10^{-1} \quad (26)$$

$$\frac{gT_m}{U_A} = 8.134 \quad (27)$$

$$\frac{gt}{U_A} = 7.15 \times 10^4. \quad (28)$$

Note that a given calculation for a duration should be checked to ensure that it has not exceeded the maximum wave height or period possible for the given adjusted wind speed and fetch.

Examples of application of JONSWAP formulations are given in Graf et al. (1984), and in addition correction of wind difference between land and sea as well as temperature difference between sea and air are provided in Liu et al. (1984). For our discussion we set $U_{10} = U_A$, $H_{1/3} = H_{m0}$, and $T = T_m$.

As a further check, since

$$T_{1/3} = \frac{U_{10}}{0.13g} \quad (29)$$

for fully developed sea, according to Pierson and Moskowitz (1964) (see Komen et al., 1984, Eq. 3.3)

$$T_m = T_{1/3}/0.95 \quad \text{or}$$

$$= U_{10}/0.95 \times 0.13g = 8.10U_{10}/g$$

which is nearly equal to $8.13U_{10}/g$, as given in (27).

From the Charnock equation as shown in (15) we see that Z_0 is related to the value of U_*, which by definition of (7) is related to the drag coefficient, wind speed, and atmospheric stability explicitly. However, the generalized roughness equation as proposed by Hsu (1974) as shown in (17) explicitly incorporates the wave parameters in addition to the wind and stability because $H_{1/3}$ and C are related to the fetch, the duration, and the wind speed through (22) through (28).

From (15), (17), and (22) it can be shown that the Charnock coefficient, a, is

$$a = \frac{2\pi}{g}\frac{H_{1/3}}{T^2}. \quad (30)$$

This equation shows that the Charnock coefficient is a constant when $H_{1/3}$ and T are given. For example, for fully developed seas, from (26) and (27) we have

$$a = 0.023, \quad (31)$$

which is reasonable compared to the most recent value, 0.0185, obtained by Wu (1982). Note, however, that values of a reported in the literature vary widely, ranging from 0.0144 (Garratt, 1977) to 0.046 (Schwab, 1978). Variations in a are most likely due to field measurements made under conditions of partially developed seas. Furthermore, under fetch- or duration-limited conditions (29) and (30) should be applied with (25) as a constraint.

On the basis of (7), (17), and (22) through (28), a mechanism for the increase of drag coefficient with wind speed is proposed. The mechanism states that C_D is related explicitly to the wind speed, U_Z, the wave height, $H_{1/3}$, and phase velocity, C, or wave period, T, and the atmospheric stability, $\xi = Z/L$, such that

$$C_D = \kappa^2/[\ln Z - \ln Z_0 - \psi_m(\xi)]^2 \quad (7)$$

where Z_0 is given in (17).

Note that for a given $H_{1/3}$ and stability condition, the older the wave, i.e., large C/U_*, the smaller the C_D value. This is explained by the following equation, derived by substituting (17) into (7):

$$C_D = \kappa^2/[\ln Z + \ln(2\pi) - \ln H_{1/3} + 2\ln(C/U_*) - \psi_m(\xi)]^2. \quad (32)$$

This reasoning may explain the difference in the variation of C_D between lake environments such as Lake Ontario (Donelan, 1982) and Lake Geneva (Graf et al., 1984) and oceanic environments (Smith, 1980; Large and Pond, 1981) (Fig. 1).

Since wave age is rarely measured, Z_0 may be reduced by incorporating (22), (23) and (24) so that

$$Z_0 = AC_{10}F^{-1/6}U_{10}^{7/3} \quad (33)$$

where $A = 0.00859$.

Substituting (33) into (7) and simplifying:

$$C_D = \kappa^2/[\ln Z + 4.7571 - \ln C_{10} + 0.167\ln F - 2.333\ln U_z - \psi_m(\xi)]^2. \quad (34)$$

Note that C_D depends more on the effect of wind speed than on fetch.

It is now clear that for a given wind speed and stability, when the fetch is large, C_D is small, as discussed before. On the other hand, for a given fetch and stability, the value of C_D will increase as the wind speed increases. Equation (34) is thus the mechanism proposed here to explain the increase of C_D with U_z. Note that its derivation is based on JONSWAP formulation [(23) through (25)]. Therefore, to apply (34), (23) through (25) should be validated. A quick way to do this is to use the nomogram of deepwater significant wave prediction curves as functions of wind speed, fetch length, and wind duration, as provided in U.S. Army Corps of Engineers (1984, Fig. 3-23 or 3-24).

3. Results

We now apply the results from the previous discussion to a simple case. Under fully developed sea and adiabatic or near-neutral conditions, the stability parameter may be neglected. Thus

$$Z_0 = \frac{1}{2\pi}\frac{H_{1/3}}{(C/U_*)^2} = \frac{H_{1/3}g}{(2\pi)C^2}\frac{U_*^2}{g}.$$

From (22), (26), and (27), we have

$$Z_0 = 0.026\frac{U_*^2}{g}.$$

Now

$$C_{10} = \left(\frac{U_*}{U_{10}}\right)^2 = \left(\frac{\kappa}{\ln(Z_{10}/Z_0)}\right)^2 = \left\{\frac{\kappa}{\ln[z_{10}/(0.026U_*^2/g)]}\right\}^2$$

$$= \left\{\frac{\kappa}{\ln[gZ_{10}(C/U_*)^2/(0.026C^2)]}\right\}^2$$

$$= \left\{\frac{\kappa}{\ln\{gZ_{10}[(C/U_*)^2/0.026][U_{10}{}^2/(0.13\times 2\pi)]^2\}}\right\}^2$$

or

$$C_{10} = \left\{\frac{\kappa}{\ln A - 2\ln U_{10}}\right\}^2$$

where

$$\ln A = \ln\left[\frac{gZ_{10}(C/U_*)^2(0.13\times 2\pi)^2}{0.026}\right\}.$$

Setting $Z_{10} = 10$ m and $C/U_* = 29$ for fully developed seas (e.g., see Graf, 1984), we have

$$\ln A = 14.56.$$

Therefore

$$C_{10} = \left\{\frac{\kappa}{14.56 - 2\ln U_{10}}\right\}^2 = \left\{\frac{0.4}{14.56 - 2\ln U_{10}}\right\}^2. \quad (35)$$

Equation (35) is plotted in Fig. 3. This figure is based on Garratt (1977), who provided a fit of data by Charnock's formulation [see (15), where $a = 0.0144$]. Note that the difference between (35), the solid line in the figure, and the dashed line, from Garratt, is not significant. We concluded that (35) is a useful formula based on fully developed sea conditions. Of course, when the sea is not fully developed, both fetch and duration of the wind should be taken into account and (23) through (28) should be employed. Because (34) is transcendental, an iteration scheme may be applied; C_{10} converges after 5 to 10 iterations, typically.

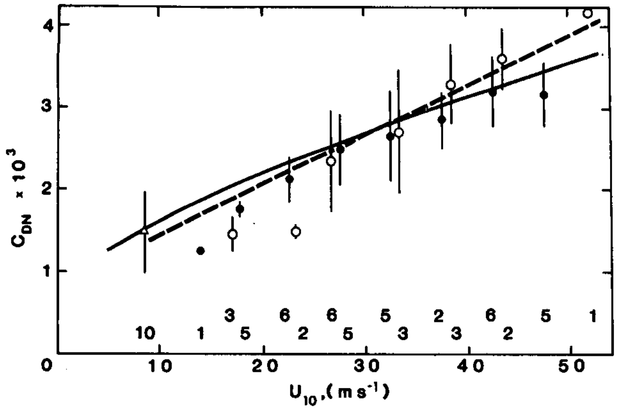

FIG. 3. From Garratt (1977), except the solid line, which is based on (35).

Note that since waves incorporate explicitly the effect of atmospheric stability (Janssen and Komen, 1985), (17) has been applied successfully under nonadiabatic conditions for estimating fluxes of momentum (Hsu, 1976) and heat (Hsu, 1983) over the ocean.

In order to compare (35) with other studies, it is plotted as line number 9 in Fig. 1. It can be seen that C_{10} increases with wind speed and that the value of C_{10} calculated by the present method is approximately the mean value found by the other methods. As a cross check, under the condition of fully developed seas, i.e., in a region of nearly unlimited fetch and duration such as the tropics, wind speed usually ranges from 5 to 10 m s^{-1}. Equation (35) predicts that C_{10} ranges from 1.2 to 1.6×10^{-3}, which is in good agreement with the constant value of 1.5×10^{-3} from BOMEX (Pond et al., 1971) and $1.4 (\pm 0.4) \times 10^{-3}$ from GATE (Businger and Seguin, 1977).

As a further test of (35) against other sources, Fig. 4 is provided. Except for the curve labeled "HSU," the figure is the same as in Donelan (1982, Fig. 2). In the figure G′ is obtained from Garratt (1977), S from Smith (1980), L and P from Large and Pond (1981), H from Heaps (1969), T from Timmerman (1977), and P from Platzman (1963). The lines indicated by G′, S, and L and P are regression lines from eddy correlation estimates; the lines representing T, H, and P are the formulae adopted by three storm surge modelers; the solid circles are derived from water level fluctuation over several months (Schwab, 1981); and the open circles are derived from the peak storm surges for two storms (Simons, 1974, 1975). From Fig. 4 it is concluded that our (35) may be considered as an average line, which is certainly consistent with the sources discussed above. Therefore, (35) should be a useful formula for practical applications in the marine environment.

4. Concluding remarks

Two remarks are in order. First, the results are interesting in that they are capable of explaining the dif-

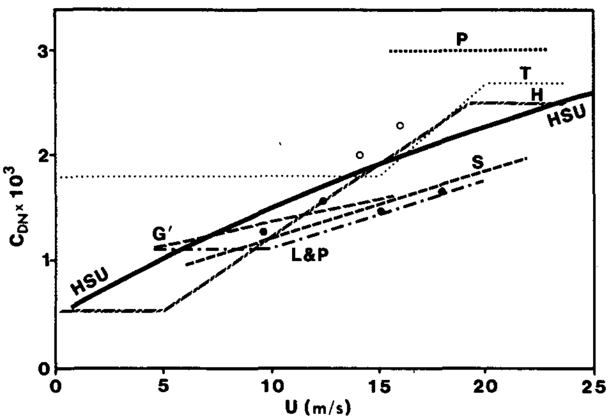

FIG. 4. The neutral drag coefficient C_{DN} versus wind speed U from other sources. The solid lines are regression lines from eddy correlation estimates of $\overline{u'w'}$; the dashed lines are the formulae adopted by three storm surge modelers; the solid circles are derived from the water level fluctuations over several months (Schwab, 1981); the open circles are derived from the peak storm surge for two storms (Simons, 1974, 1975). (After Donelan, 1982.) The curve labeled HSU is based on (35). See text for explanation.

ference in the results obtained by, among others, Donelan (1982, Fig. 2) and Smith (1980). Note, however, that both Smith (1980) and Large and Pond (1981) felt that the Charnock postulate does not apply to the open ocean as far as C_D predictions are concerned. They additionally indicate that even if it could be applied, there would be limitations to where and over which windspeed range it could be considered.

Second, questions were raised by one of the reviewers concerning the effect of swell on the choice of the phase speed, C, and how the modeler handles such cases if routine ocean data is to be applied to the proposed model. Readers are directed to a computer program that computes waves which include seas and swells (see Freeman, 1985). Note that when a group of waves of various heights and periods leaves the fetch area, the waves gradually become swell. The faster low-frequency waves move ahead of the group, while the slower high frequency waves drop behind. The significant swell period increases with decay distance, whereas significant wave height decreases. This change in swell depends on the significant wave spectrum at the end of the fetch and the decay distance (Kinsman, 1965). The formulas for wave decay and travel time are also in the program. Note that Freeman's formula on p. 990 (Eq. 35.2) was in error. There should be a plus sign instead of equal sign inside the square root (Freeman, personal communication, 1985).

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