Bulk Parameterization of Air-Sea Exchanges of Heat and Water Vapor Including the Molecular Constraints at the Interface[1]

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ABSTRACT

A model is developed for the marine atmospheric surface layer including the interfacial sublayers on both sides of the air-sea interface where molecular constraints on transports are important. Flux-profile relations which are based on the postulation of intermittent renewal of the surface fluid are matched to the logarithmic profiles and compared with both field and laboratory measurements. These relations enable numerical determination of air-sea exchanges of momentum, heat and water vapor (or bulk transfer coefficients) employing the bulk parameters of mean wind speed, temperature and humidity at a certain height in the atmospheric surface layer, and the water temperature.

With increasing wind speed, the flow goes from smooth to rough and the bulk transfer coefficient for momentum also increases. The increase in roughness is associated with increasing wave height which in the present model results in sheltering at the wave troughs. Due to the decrease in turbulent transports, the transfer coefficients of heat and water vapor decrease slightly with wind speed after the wind speed exceeds a certain value. The bulk transfer coefficients are also found to decrease with increasing stability. If the "bucket temperature" which typically gives the water temperature a few centimeters below the surface is used, rather than the interfacial temperature, erroneous results may be obtained when the air-sea temperature difference is small.

By including the effects of stability and interfacial conditions in bulk parameterization, the model provides a way to account for physical conditions which are known to affect air-sea exchanges.

1. Introduction

In the bulk parameterization, the fluxes are determined with transfer coefficients which relate the fluxes to the variables measured, i.e.,

$$H = c\rho C_H(U - U_s)(T_s - T), \quad (1a)$$

$$E = \rho C_E(U - U_s)(Q_s - Q), \quad (1b)$$

$$\tau = \rho C_D(U - U_s)^2, \quad (1c)$$

where ρ and c are the density and isobaric specific heat of air; τ, H and E are the stress, heat and moisture fluxes; U, T and Q are the wind speed, potential temperature and specific humidity at a reference height in the atmospheric surface layer; and U_s, T_s and Q_s are the wind speed, temperature and specific humidity at the surface. The transfer coefficient for momentum, C_D, also known as the drag coefficient, has been studied extensively. For deep water with large fetch, it has been expressed as a function of wind speed or assumed to be constant over a range of moderate wind speeds. Summaries of these studies have been given by Businger (1975), Garratt (1977) and others. The transfer coefficient for sensible heat, C_H, and water vapor, C_E, are less well known and generally assumed to be equal to the drag coefficient. In the past few years, a number of investigators (e.g., Pond *et al.*, 1974; Friehe and Schmitt, 1976) haye examined the results of flux measurements over the ocean to determine the values of these two coefficients. However, these coefficients are usually treated as constants and their variations with wind speed and stability are neglected.

The bulk transfer coefficients can be determined by integrating the velocity, temperature and humidity profiles. The turbulent processes in the atmospheric surface layer have been extensively studied and satisfactory similarity descriptions have been given. Close to the surface but beyond the region where molecular effects are important, the distribution of velocity, temperature and humidity are governed by the diabatic profiles (e.g., Businger, 1973a):

$$(T - T_s)/T_* = [\ln(z/z_T) - \psi_T]/\alpha_H k, \quad (2a)$$

$$(Q - Q_s)/Q_* = [\ln(z/z_Q) - \psi_Q]/\alpha_E k, \quad (2b)$$

$$(U - U_s)/U_* = [\ln(z/z_0) - \psi_U]/k, \quad (2c)$$

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where $U_* = (\tau/\rho)^{1/2}$, $T_* = -H/(c\rho U_*)$, $Q_* = -E/(\rho U_*)$, $\alpha_H = K_H/K_M$ and $\alpha_E = K_E/K_M$ at neutral stability; K_M, K_H and K_E are the turbulent diffusivities of momentum, heat and moisture respectively and k is the von Kármán constant. Under unstable condition, the Businger-Dyer model gives

$$\psi_T = 2\ln[(1+Y)/2], \tag{3a}$$

$$\psi_Q = 2\ln[(1+Y')2], \tag{3b}$$

$$\psi_U = 2\ln[(X+1)/2] + \ln[(X^2+1)/2] - 2\tan^{-1}(X) + \pi/2, \tag{3c}$$

where $X = (1 + a_U\zeta)^{1/4}$, $Y = (1 + a_T\zeta)^{1/2}$, $Y' = (1 + a_Q\zeta)^{1/2}$ and $\zeta = z/L$. Under stable conditions

$$\psi_T = -b_T\zeta, \tag{4a}$$

$$\psi_Q = -b_Q\zeta, \tag{4b}$$

$$\psi_U = -b_U\zeta. \tag{4c}$$

Including the effect of moisture fluctuations on buoyancy, the Obukhov length L can be defined as (Busch, 1973) $L = (T_v U_*)^2/(gkT_{v*})$, where $T_v = T(1 + 0.61Q)$ and $T_{v*} = T_*(1 - 0.61Q) + 0.61TQ_*$. In this paper, $k^{-1} = 2.5$, $(\alpha_H k)^{-1} = (\alpha_E k)^{-1} = 2.2$, $a_U = a_T = a_Q = 16$, and $b_U = b_T = b_Q = 7$ are used (Paulson, 1970; Monin and Yaglom, 1971). Experiments designed specifically to measure the values of these parameters over water are needed. For small values of z, ψ_T, ψ_Q and ψ_U approach zero and the profiles (2) approach the logarithmic forms. Methods to include stability effects on the transfer coefficients have been discussed by Deardorff (1968), Leovy (1969) and Paulson (1969).

The profiles (2), however, are not valid very close to the interface. The air-sea interface is a material surface. While transport in air and water is facilitated by turbulent motion, the motion is suppressed near the interface and one expects molecular diffusion to dominate. Since diffusion is a much slower process, the regions on both sides of the interface form a bottleneck for exchanges between the ocean and the atmosphere. Most of the variations in velocity, temperature and concentration in the lower part of the atmosphere and upper part of the ocean are found in these regions. In this paper, the region will be referred to as the interfacial sublayer. The lower boundary values z_0, z_T and z_Q of the logarithmic profiles (2) depends on the distributions in the sublayer in the air.

Although the sea surface temperature can be determined remotely by a radiometer, only the so-called "bucket temperature" T_w is usually available. The bucket temperature can be the temperature of the water from a few centimeters to a meter from the surface and differs from the sea surface temperature due to the aqueous interfacial sublayer. Primed symbols will be used to distinguish the transfer coefficients obtained by using the bucket temperature:

$$C'_H = H/[c\rho(U - U_s)(T_w - T)], \tag{5a}$$

$$C'_E = E/[\rho(U - U_s)(Q_w - Q_s)], \tag{5b}$$

where Q_w is the saturation humidity at T_w.

A number of investigators, particularly those who studied evaporation (e.g., Sverdrup, 1951; Sheppard, 1958; Kitaigorodskii and Volkov, 1965), have included a laminar layer with a rather arbitrary thickness in their evaluation of the coefficients. Kondo (1975) incorporated the empirical relation suggested by Owen and Thomson (1963) for the sublayer Stanton number in rough flow in addition to the theory of laminar sublayer in smooth flow in his evaluation of the coefficients. Deacon (1977) has also applied the results of some viscous sublayer studies to gas transfer. In the evaluation of C_H, Hicks (1972) applies an overall correction to the bucket temperatures to account for the temperature drop across the sublayer. A survey of the studies of the sublayer has been given by Liu (1978). However, further systemic studies on how the sublayers on both sides of the interface affect the bulk transfer coefficients are required.

In this paper, we will develop a surface layer model which includes the interfacial sublayers and allows us to formulate the fluxes (or transfer coefficients) in relation to the sea surface temperature as well as the bucket temperature.

2. The interfacial sublayer

By extending the so-called "surface renewal model", Liu and Businger (1975) suggested a flux-profile relation for the temperature distribution in the interfacial sublayer:

$$(T - T_s)/(T_b - T_s) = 1 - \exp(-z/\delta_T), \tag{6a}$$

$$\delta_T = \rho c(T_s - T_b)/H = (\kappa t_*)^{1/2}, \tag{7a}$$

where κ is the diffusivity of heat. In deriving these equations, the fluid in contact with the interface is assumed to become unstable and to be replaced intermittently by the fluid from the bulk. The temperature of the bulk is T_b and is used as the initial condition in solving for the temperature distribution due to heat diffusion. The average duration of fluid contact with the interface is t_* and δ_T is a scaling depth equivalent to the thickness of a laminar or stagnant layer where the same heat flux and temperature difference are sustained solely by diffusion. The relations (6a) and (7a) agree with measurements in the Black Sea (Khundzhua and Andreyev, 1974) and laboratory measurements on both sides of an air-water interface of a tank of water (Katsaros *et al.*, 1977; Liu, 1974, 1978). Similar relations can be derived for moisture transport, i.e.,

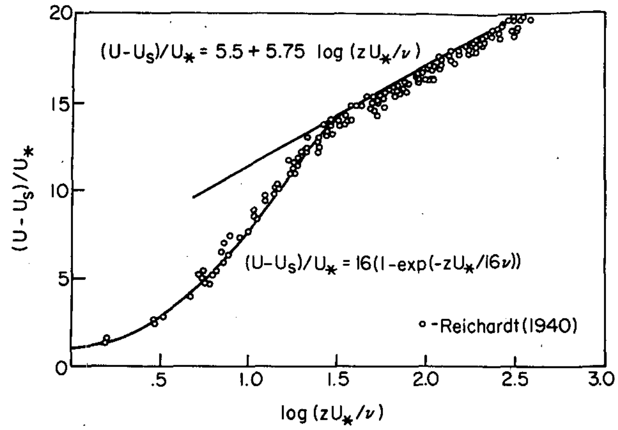

FIG. 1. Velocity profile in the interfacial sublayer evaluated from the present model compared with the laboratory measurements by Reichardt (1940).

$$(Q - Q_s)/(Q_b - Q_s) = 1 - \exp(-z/\delta_Q), \quad (6b)$$

$$\delta_Q = \rho(Q_s - Q_b)/E = (\epsilon t_*)^{1/2}, \quad (7b)$$

where ϵ is the diffusivity of water vapor. In the vicinity of a smooth surface, similar relations for momentum transport would be valid:

$$(U - U_s)/(U_b - U_s) = 1 - \exp(-z/\delta_U), \quad (6c)$$

$$\delta_U = \rho(U_b - U_s)/\tau = (\nu t_*)^{1/2}, \quad (7c)$$

where ν is the kinematic viscosity. Eq. (6) can be written as

$$(T - T_s)/T_* = S[1 - \exp(-zU_*/S\kappa)], \quad (8a)$$

$$(Q - Q_s)/Q_* = D[1 - \exp(-zU_*/D\epsilon)], \quad (8b)$$

$$(U - U_s)/U_* = C[1 - \exp(-zU_*/C\nu)], \quad (8c)$$

where the sublayer parameters are defined as

$$S = (T_b - T_s)/T_* = \delta_T U_*/\kappa, \quad (9a)$$

$$D = (Q_b - Q_s)/Q_* = \delta_Q U_*/\epsilon, \quad (9b)$$

$$C = (U_b - U_s)/U_* = \delta_U U_*/\nu. \quad (9c)$$

The quantities T_b, Q_b and U_b will be redefined as equivalent bulk values which when used as initial conditions in solving the diffusion equations will give sublayer profiles (8) as solutions which match the logarithmic profiles (2) both in magnitude and in slope at certain distances from the interface. This requirement of smooth transition enables one to evaluate not only the thickness of the sublayers but also the lower boundary values of the logarithmic profiles (z_0, z_T, z_Q) when the sublayer parameters S, D, C are known. Brutsaert (1975) suggested that the renewal time scale t_*, is equivalent to the time scale of the Kolmogorov eddies, i.e.,

$$t_* \propto (\nu z_0/U_*^3)^{1/2}. \quad (10)$$

Combination of this hypothesis with (7) and (9) yields

$$S = G\,\mathrm{Rr}^{1/4}\,\mathrm{Pr}^{1/2}, \quad (11a)$$

$$D = G\,\mathrm{Rr}^{1/4}\,\mathrm{Sc}^{1/2}, \quad (11b)$$

$$C = G\,\mathrm{Rr}^{1/4}, \quad (11c)$$

where $\mathrm{Rr} = z_0 U_*/\nu$ is the roughness Reynolds number, $\mathrm{Pr} = \nu/\kappa$ is the Prandtl number and $\mathrm{Sc} = \nu/\epsilon$ is the Schmidt number. The proportionality constant G depends on interfacial roughness characteristics. In rough flow, (11c) is no longer relevant. Lamont and Scott (1970) arrived at an equivalent result for a free surface through a different approach.

In smooth flow, extensive experimental results show that $z_0 = 0.11\nu/U_*$ (e.g., Kondo, 1975) and this value gives a smooth transition at $z = 47\nu/U_*$ and $C = 16$. The values of z_T and z_Q, on the other hand, depend on the type of fluid used. Figs. 1 and 2 show the interfacial velocity and temperature pro-

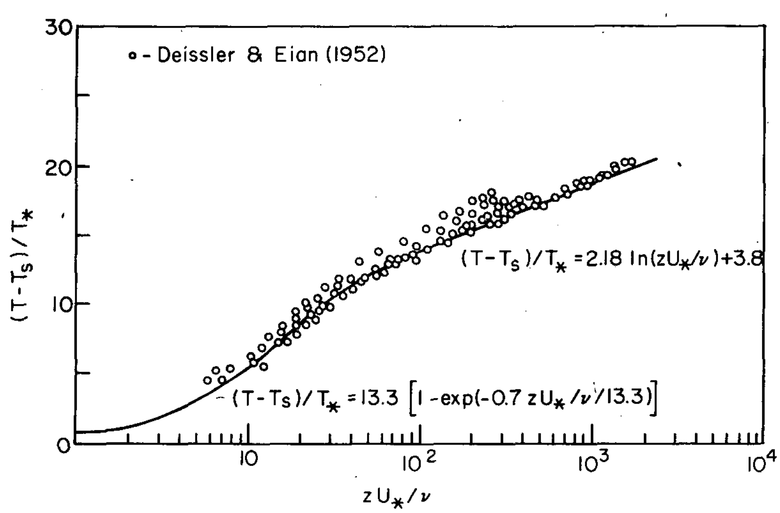

FIG. 2. Temperature profile in the interfacial sublayer evaluated from the present model compared with the laboratory measurements by Deissler and Eian (1952).

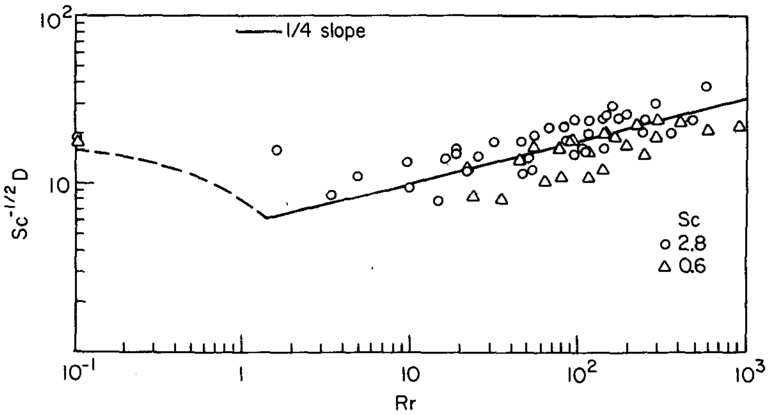

FIG. 3. $Sc^{-1/2}D$ versus Rr evaluated from the present model compared with the measurements by Chamberlain (1968).

files matching smoothly to the logarithmic profiles and in good agreement with measurements in smooth channel flows given by Reichardt (1940) and Deissler and Eian (1952).

In the case when there is little mean wind, convection still causes local horizontal flow near the surface with a characteristic friction velocity of (see Businger, 1973b)

$$U_* = \chi(gd\overline{W'T'}/T)^{1/3}, \quad (12)$$

where g is the acceleration due to gravity, d the convective length scale, $\overline{W'T'}$ the covariance of the fluctuations of vertical wind speed and potential temperature, and χ decreases with increasing d/z_0. With the assumption that $\chi \propto (d/z_0)^{-1/3}$ and replacing $(1/T)$ with λ, the thermal expansivity, combination of (7a), (10) and (12) yields $Nu \propto Ra^{1/3}$, where $Nu = Hd/c\rho\kappa(T_s - T_b)$ and $Ra = \lambda g(T_s - T_b)d^3/\kappa\nu$ are the Nusselt and Rayleigh numbers. This is in agreement with the laboratory results of Katsaros *et al.* (1977).

Chamberlain (1968) obtained mass transfer data of water vapor ($Sc = 0.6$) and radioactive thorium ($Sc = 2.8$) in air, to and from surfaces with spherical, cylindrical and wavelike roughness elements in a wind tunnel. From these data z_Q can be evaluated from (2) and D can be determined by requiring a smooth transition between profiles (2) and (8). The values of $Sc^{-1/2}D$ for the two sets of data were plotted as a function of Rr in Fig. 3. In rough flow they scatter around a line with ¼ slope.

Mangarella *et al.* (1973) measured heat ($Pr = 0.7$) and water vapor ($Sc = 0.6$) transfer in the air over artificially generated waves in a wind tunnel. The results are shown in Fig. 4. The values of z_T and z_Q are evaluated from data at the fourth measurement levels which have different distances from the surface but, according to the authors, all lie within the logarithmic region. In rough flow, almost all the values of $Pr^{-1/2}S$ and $Sc^{-1/2}D$ lie on a line with a slope close to ¼ as predicted.

Another three sets of data from Mangarella *et al.* (1973) processed in a similar fashion are shown in Fig. 5. They are for heat transfer from heated water, and water vapor transfer from heated and unheated water over a wavy surface generated by wind. In the fully rough region, the three sets of data fall around a single line with a slope very close to ¼ and the value of G that agrees best with the data is 9.3. Since the interfacial conditions under which these three sets of data were determined resemble most closely the natural conditions at an air-sea interface, this value of G will be used in the rest of this paper.

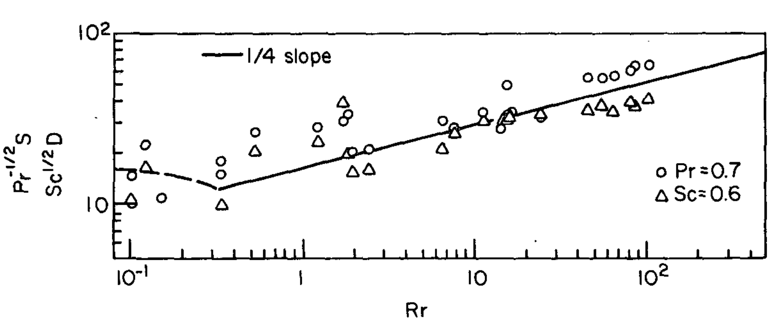

FIG. 4. As in Fig. 3 except for comparison with the measurements over mechanically generated waves by Mangarella *et al.* (1973).

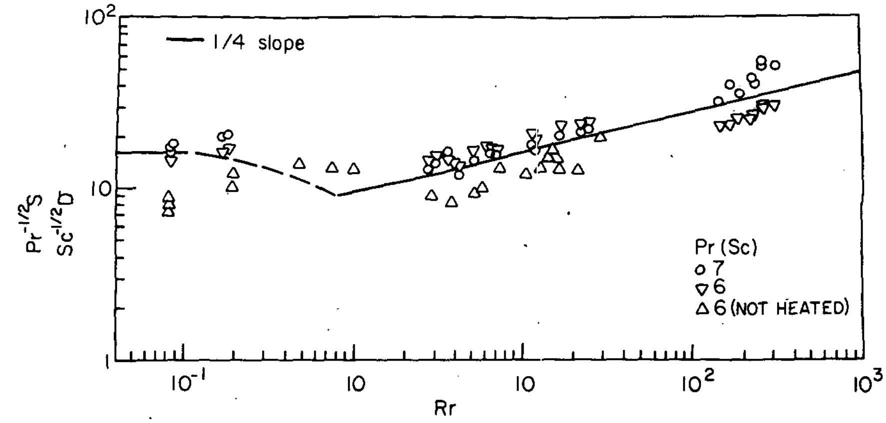

FIG. 5. As in Fig. 3 except for comparison with measurements over wind generated waves by Mangarella *et al.* (1973).

Both the measurements of Nikuradse (1933) on pipe flow and those of Kondo (1975) at an air-sea interface indicate

$$z_0 = \begin{cases} G_1\nu/U_*, & \xi U_*/\nu \leq r_1, \quad (13a) \\ f(\xi U_*/\nu), & r_1 \leq \xi U_*/\nu \leq r_2, \quad (13b) \\ G_2\xi, & r_2 \leq \xi U_*/\nu. \quad (13c) \end{cases}$$

Nikuradse (1933) defined ξ as the actual mean diameter of the sand grains used as roughness elements and found $r_1 = 5$, $r_2 = 70$, $G_1 = 0.11$ and $G_2 = 0.03$. Kondo (1975) defined ξ as the root-mean square wave height for high-frequency components and found $r_1 = 5.7, r_2 = 67, G_1 = 0.11$ and $G_2 = 0.07$. These results separate the flow roughly into smooth, transition and rough regimes. In the smooth regime, momentum is transported by viscosity and (8c) is valid. As the roughness increases, momentum can also be transported by pressure forces on the roughness elements, ξ gradually replaces ν/U_* as the dominating length scale of the sublayer or the Kolmogorov eddies, and (8c) loses its validity. The broken lines in Figs. 3–5 are extrapolations of the hypothesis for smooth flow

$$S = C\ \mathrm{Pr}^{1/2}, \quad (14a)$$

$$D = C\ \mathrm{Sc}^{1/2}, \quad (14b)$$

into the transition regions with C determined from the requirement of a smooth transition between (2c) and (8c). For $\mathrm{Rr} \geq 1.2$, there can no longer be a smooth transition between the two profiles, in agreement with the above physical reasoning. By extending the line with ¼ slope to meet the dashed line, sharp depressions in the value of $\mathrm{Sc}^{-1/2}D$ were created in the figures. In reality, we expect a smoother transition.

With the values of S and D evaluated from (11) and (14), the sublayer profiles are fixed and the values of z_T and z_Q can be determined. They are shown in Figs. 6 and 7. The values can be approximated by relations of the form

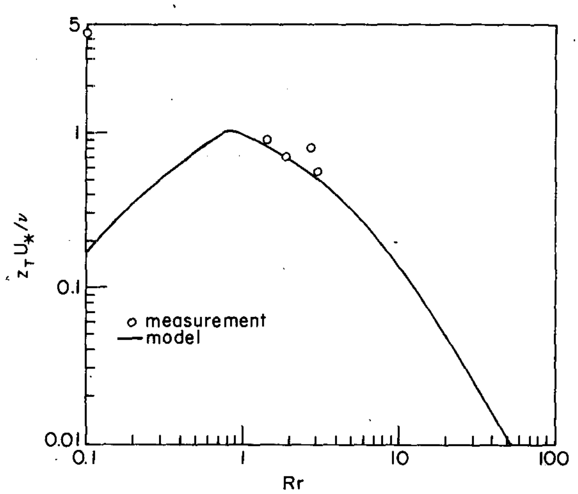

FIG. 6. $z_T U_*/\nu$ versus Rr evaluated from the present model compared with field measurements.

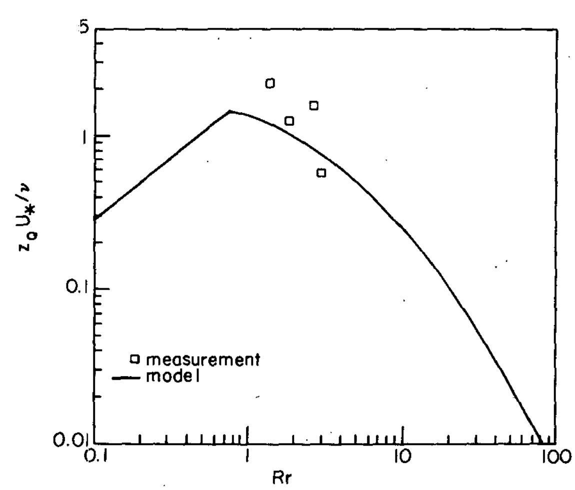

FIG. 7. $z_Q U_*/\nu$ versus Rr evaluated from the present model compared with field measurements.

$$z_T U_*/\nu = a_1 \, \mathrm{Rr}^{b_1}, \quad (15a)$$

$$z_Q U_*/\nu = a_2 \, \mathrm{Rr}^{b_2}, \quad (15b)$$

where the values of a_1, b_1, a_2, b_2 for different ranges of Rr are shown in Table 1. Also shown in Figs. 6 and 7 are a few measurements obtained with instruments mounted on a mast in Lake Washington which is open to wind with a fetch of 4 km over water. The instruments are a propeller-anemometer system, a thermocouple psychrometer mounted about 4 m above water and a Barnes PRT-5 radiometer looking down at the water surface. The values of Rr, z_T and z_Q were evaluated from τ, H, E, U, T, Q at 4 m and T_s according to (2) with $U_s = 0.55U_*$ (Wu, 1975). Most of the data have wind speeds between 4 and 9 m s^{-1} and fall at the margin between the transition and rough regimes. They agree with the prediction in a very general way. Details of the experiment and evaluation are presented in Liu (1978).

The quantities S, D, z_T and z_Q depend on Rr as shown in (11) and (15). Recently, Smith and Banke (1975), Kondo (1975), Garratt (1977) and others have obtained empirical relations between C_D and the wind speed at 10 m, U_{10}. These relations can be used in conjunction with (2c) to determine the values of z_0 and Rr. Smith and Banke (1975) found the relationship between C_D and U_{10} for 111 data runs, *viz*,

$$10^3 C_D = 0.63 + 0.066 U_{10}, \quad (16a)$$

for U_{10} between 3 and 21 m s^{-1}. Garratt (1977) obtained from 18 sets of data including those of Smith and Banke (1975) the relation

$$10^3 C_D = 0.75 + 0.067 U_{10}, \quad (16b)$$

for U_{10} between 4 and 21 m s^{-1}. Both (16a) and (16b) agree with Charnock's (1955) postulation that z_0 is proportional to U_*^2/g. Kondo (1975), by using wind-wave data from Kondo *et al.* (1973), obtained approximate relations of the form

$$10^3 C_D = p + qU_{10}^r \quad (16c)$$

that agree with (13). His values of p, q, r for different ranges of wind speed are shown in Table 2. The differences between C_D given by (16a, b, c) are small for moderate wind speeds. Garratt (1977) did not use the result of Kondo (1975) to evaluate (16b) but indicated that (16c) agrees well with the data he collected. Since (16c) covers wind speeds as low as 0.3 m s^{-1}, it will be used in this paper to determine z_0 at neutral stability. The drag coefficients discussed here are all evaluated with the assumption that $U_s = 0$. The roughness parameter z_0' determined under this assumption is related to the actual z_0 by $z_0' = z_0 \exp(-kU_s/U_*)$.

TABLE 1. The lower boundary values of the logarithmic profiles $Z_T U_*/\nu = a_1 \, \mathrm{Rr}^{b_1}$ and $Z_Q U_*/\nu = a_2 \, \mathrm{Rr}^{b_2}$

Rr	a_1	b_1	a_2	b_2
0 – 0.11	0.177	0	0.292	0
0.11 – 0.825	1.376	0.929	1.808	0.826
0.925– 3.0	1.026	−0.599	1.393	−0.528
3.0 – 10.0	1.625	−1.018	1.956	−0.870
10.0 – 30.0	4.661	−1.475	4.994	−1.297
30.0 –100.0	34.904	−2.067	30.790	−1.845

TABLE 2. Drag coefficient in neutral conditions suggested by Kondo (1975).

U (m s^{-1})	$10^3 C_D$
0.3– 2.2	$1.08U^{-0.15}$
2.2– 5.0	$0.771 + 0.0858U$
5.0– 8.0	$0.867 + 0.0667U$
8.0–25.0	$1.2 + 0.025U$
25.0–50.0	$0.073U$

Near a smooth boundary, Kader and Yaglom (1972) suggested that $-\ln(z_T U_*/\nu)/\alpha_H k = 12.5 \, \mathrm{Pr}^{2/3} - 2.12 \ln(\mathrm{Pr}) - 5.3$ and for a rough rigid boundary Yaglom and Kader (1975) suggested $-\ln(z_T U_*/\nu)/\alpha_H k = 0.55(\xi U_* \nu)(\mathrm{Pr}^{2/3} - 0.2) - 2.12 \ln(\xi U_*/\nu) + 9.5$. The dependence on Pr is partly based on the assumption that the eddy diffusivities are proportional to $(zU_*/\nu)^3$ near the surface. This assumption has been frequently used by investigators who study heat and mass transport in pipe flow. For a smooth and for a rigid wall, the first nonzero term in the Taylor expansion of the eddy diffusivities around $zU_*/\nu = 0$ is of the third order (Monin and Yaglom, 1971). At a free surface, the no-slip constraint is relaxed, the fluid has more horizontal freedom, normal velocity fluctuations are allowed closer to the surface, and the diffusivities may have a dependence on zU_*/ν different from that at a rigid surface. The data in Figs. 3–5 do not cover a large enough range to resolve for the Prandtl number dependence. For air and water under natural conditions, the Prandtl number is about the same order of magnitude and a small difference in the formulations of the Prandtl number dependence should not change the result significantly.

Another approach by which the lower boundary conditions for the logarithmic profiles may be expressed is by means of the so-called inverse sublayer Stanton number $\mathrm{S}_h = (T_h - T_s)/T_*$ and the inverse sublayer Dalton number $\mathrm{D}_h = (Q_h - Q_s)/Q_*$, where T_h and Q_h are the temperature and humidity at $z = h$ and h is the boundary between the sublayer and the region where the logarithmic profiles are valid. Assuming that the velocity at $z = h$ is U_s, i.e., $h = z_0$, Owen and Thomson (1963) found by fitting experiment measurements that $\mathrm{S}_h = 0.25 \, \mathrm{Pr}^{0.8} \times (30 \, \mathrm{Rr})^{0.45}$ for fully rough flow. The experiment of Chamberlain (1968) and the analysis of Garratt and Hicks (1973) were aimed at testing this relation. Brutsaert (1975) assumed that both heat and mass transports near a smooth surface are controlled by renewal of the surface fluid by Kolmogorov eddies. He also assumed that the velocity changes linearly with height and obtains $S_h = 13.6\ \mathrm{Pr}^{2/3}$. The approach shares some similarity with that of Ruckenstein (1958) and the results agree with the measurements of Merlivat (1978). In rough flow, Brutsaert assumed $h = 7.4z_0$ and found that $S_h = 7.3\ \mathrm{Pr}^{1/2}\ \mathrm{Rr}^{1/4}$. These relations depend on the definition of h. The results given by our model lie among those given by these approaches.

3. Sea surface temperature

The bucket temperature measured from a ship can be viewed as the bulk temperature in the surface renewal model and can be approximated by a relation equivalent to (9a), i.e.,

$$T_s = T_w - ST_{*w}, \tag{17}$$

where the subscript w indicates water. The scale temperature T_{*w} in water is defined in a similar fashion as that in air: $T_{*w} \equiv -H_w/(c_w\rho_w U_{*w})$ and $H_w \equiv -(H + L_vE + R)$, where R is the net radiation. The nondimensional temperature difference $S\ \mathrm{Pr}^{-1/2}$ across the aqueous sublayer is plotted versus Rr in Fig. 8. The solid line represents prediction from (11a) and (14a). Since measurements are usually made in the air, it is convenient to relate parameters in the water to those in the air. For a fully developed sea, the stress is continuous across the interface with

$$U_{*a}/U_{*w} = (\rho_w/\rho_a)^{1/2}, \tag{18}$$

where the subscript a indicates air. If the capillary and short gravity waves which constitute the roughness elements at an air-sea interface are essentially the same on both sides, then

$$\mathrm{Rr}_a/\mathrm{Rr}_w = (\nu_w/\nu_a)(\rho_w/\rho_a)^{1/2}. \tag{19}$$

Saunders (1967) and Paulson and Parker (1972) suggested that $S_1 = c_w\rho_w\kappa_w U_{*w}(T_s - T_w)/(H_w\nu_w)$ is constant while Hasse (1971) suggested $S_2 = (T_s - T_w)U_{10}/H_w$ is constant. S can be expressed in terms of S_1 or S_2 through (17), (18) and (1c): $S = S_1\ \mathrm{Pr} = S_2 C_D^{1/2} c_w(\rho_w\rho_a)^{1/2}$. The value of S_1 was evaluated by Paulson and Parker (1972) from laboratory measurements at about 20°C over a smooth surface. An equivalent value of $S\ \mathrm{Pr}^{-1/2}$ is estimated and included in Fig. 8. It is higher than the prediction of our model and other measurements. Saunders (1967) estimated S_1 from field measurements. Hasse (1971) determined S_2 from his model and checked this value with field data. The water temperature and surface roughness were not given. Using Pr of water at 15°C and assuming a constant C_D as suggested by Hasse (1971), an equivalent value of $S\ \mathrm{Pr}^{-1/2}$ is found which is shown on the right margin of Fig. 8 for comparison. They fall within the range for a rough sea surface. The dependence of S_1 and S_2 on Pr and Rr are neglected by these investigators and their empirical relations on $(T_s - T_w)$ must therefore be viewed as approximations or averages over a range of Pr and Rr.

Hill (1972) and Street and Miller (1977) measured S in wind tunnel experiments over a range of wind speeds and wave conditions. Their average values of $S\ \mathrm{Pr}^{-1/2}$ at high wind speed (when waves start to form) are shown on the right margin of Fig. 8. The average value for a smooth surface from Hill's experiment is also shown. Both Hill (1972) and Street and Miller (1977) found that S decreases as waves start to build (transition from smooth to rough flow) which is in qualitative agreement with the present model prediction. However, they did not indicate any increase in S with further increase in roughness. The average values of S from these two sets of data do approximately correspond to the value at the transition regime of our model.

Also shown in Fig. 8 are the values of $S\ \mathrm{Pr}^{-1/2}$ evaluated from BOMEX data. The values of $(T_s - T_w)$ was obtained by Leavitt (1976, private

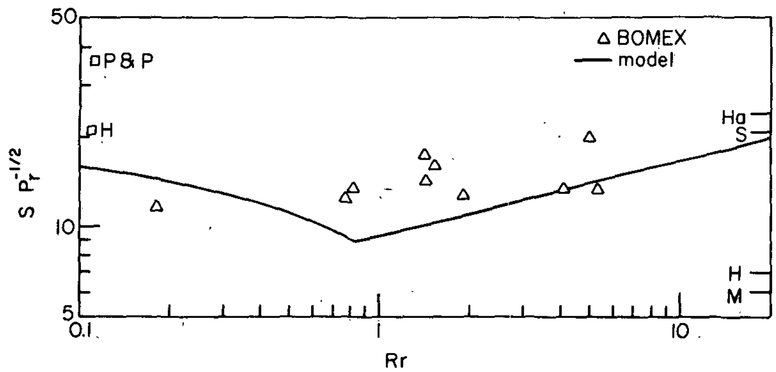

FIG. 8. $\mathrm{Pr}^{-1/2}S$ versus Rr in the aqueous interfacial sublayer evaluated from the present model compared with other hypothesis and measurements: H, Hill (1972); Ha, Hasse (1971); M, Street and Miller (1977); P&P, Paulson and Parker (1972); S, Saunders (1967).

communication) from radiometric measurements in the BOMEX experiment. The average of a 50 min data run is subtracted from the peak value of the surface temperature to give a value of $(T_s - T_w)$. The peak value was believed to be the bulk temperature revealed by mixing. The momentum, sensible and latent heat fluxes were evaluated by Paulson *et al.* (1972) by the profile technique. Radiative fluxes were not measured but had been estimated by us from the bulk parameters by an empirical formula. Though scattered, the values of $S\ \mathrm{Pr}^{-1/2}$ do show general agreement with our model.

Wu (1971) and Schooley (1971) simplified the problem drastically by suggesting a laminar layer on top of the ocean with thickness equal to the intersection between a linear and a logarithmic velocity profile. Street and Miller (1977) and Kondo (1976) have also advanced hypotheses on the thickness of the aqueous sublayer which were based on the studies of Yaglom and Kader (1974) and Owen and Thomson (1963), respectively. These studies have been discussed in the previous section. The effect of surface tension on the sublayer has been discussed by Omholt (1973) and Katsaros (1977). O'Brien and Omholt (1969), Witting (1971, 1972) and others have proposed theories on the effects of surface distortion due to waves and the laboratory experiments of Chang and Wagner (1975) and Miller and Street (1978) did show a correlation between surface temperature fluctuations and waves. However, such effects are not included in our model.

The net radiation R is generally not measured and must be estimated from the bulk parameters. In this paper we used Angstrom's formula (Geiger, 1965)

$$R = \sigma_s T_s^4 - \sigma_s T_a^4 (0.82 - 0.250 \times 10^{-0.094 e_a}), \quad (20)$$

where σ_s is the Stefan-Boltzmann constant and e_a the vapor pressure (mb). Brunt (1932), Swinbank (1963) and Idso and Jackson (1969) have suggested other empirical relations but they do not give significantly different values of $(T_s - T_b)$ obtained with the present model. These empirical relations are for the ideal situation of clear night sky when solar radiation can be ignored. Although only a small portion of the solar radiation is absorbed by the aqueous sublayer, it may become important in daytime and when the sum of sensible and latent heat fluxes is small.

In summary the results and data shown in Fig. 8 agree with the model, but only in a very general way. The method described here provides only a simple approximation of the sea surface temperature from bulk parameters. Further experiments, especially in the field, are needed to confirm or improve the method.

4. Bulk transfer coefficients

If U, T, Q, z and T_w are independently assigned, there are 12 dependent variables in (2): T_s, Q_s, U_s, ψ_T, ψ_Q, ψ_U, z_T, z_Q, z_0, T_*, Q_* and U_*. For $U_s = 0$ and Q_s assumed to be saturation humidity at T_s, Eqs. (3), (4), (15), (16) and (17) together reduce (2) into a set of three equations and three unknowns: T_*, Q_* and U_*. This set can be solved by iterations starting from the neutral values. The fluxes and the transfer coefficients thus can be determined from the bulk parameters.

The variations of the transfer coefficients with Rr and U_{10} at neutral stability determined in this manner are shown in Fig. 9. For neutral stability, (1) and (2) together give

$$C_D/C_H = \alpha_H^{-1} + (\alpha_H k)^{-1} C_D^{1/2} \ln(z_0/z_T). \quad (21)$$

C_D is equal to C_H only if the eddy diffusivities are equal ($\alpha_H = 1$) and the molecular effects are the same ($z_0 = z_T$). Similar conditions apply to the ratio

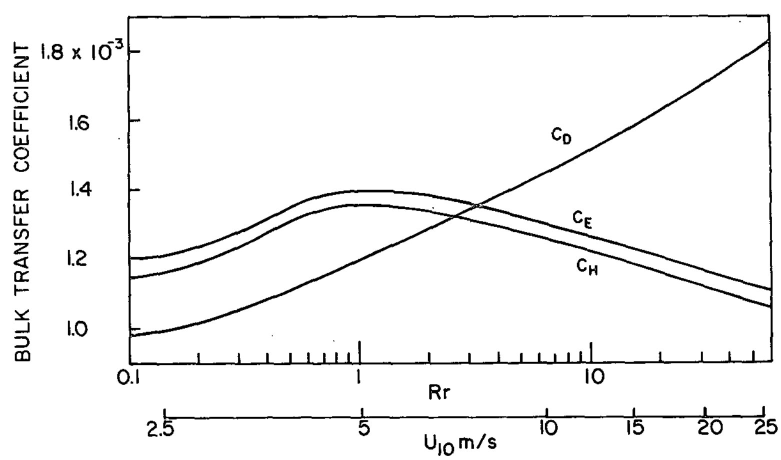

FIG. 9. Bulk transfer coefficients at neutral condition versus Rr and U_{10} evaluated from the present model.

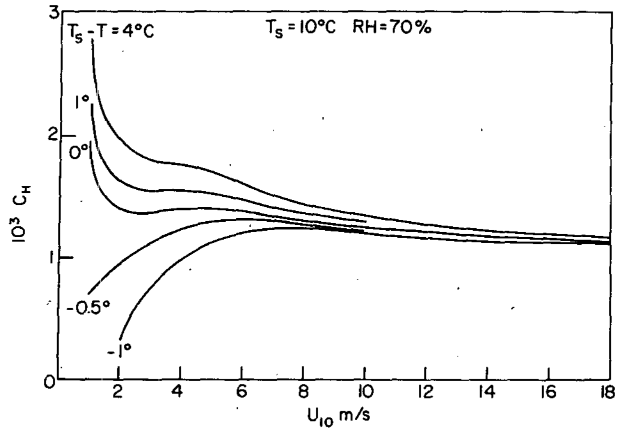

FIG. 10. C_H versus U_{10} under different diabatic conditions evaluated from the present model.

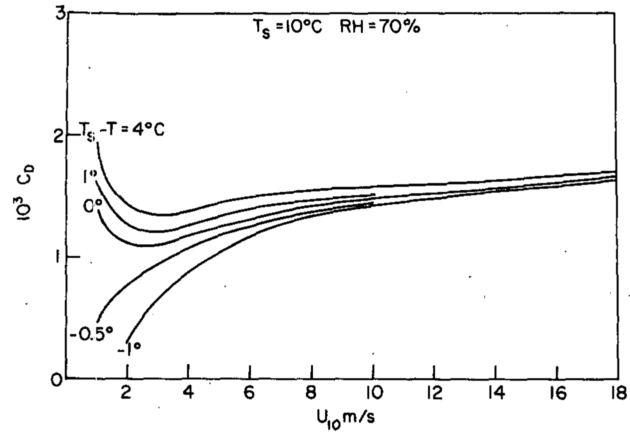

FIG. 12. C_D versus U_{10} under different diabatic conditions evaluated from the present model.

C_D/C_E. When the interface is smooth, momentum, heat and water vapor are all transported by molecular processes near the interface and the variations of C_D, C_H and C_E should share the same characteristics. When the roughness of the surface increases, turbulent transport is facilitated and the transfer coefficients thus increase with wind speed. While momentum can be transported by pressure forces on the roughness elements, molecular diffusion is the only process which transports heat and mass at the interface. According to (10) and (13c), the fluid elements would tend to have a longer contact time with the interface when the roughness height increases. This can be interpreted as an increase in sheltering effect at the troughs between the roughness elements. Therefore, heat and mass transports are subjected to greater constraints as the wind increases further. The two opposing effects due to increase in roughness balance each other at a wind speed of 4–5 m s^{-1}. This phenomenon reflects the interfacial characteristics discussed in Section 2. Since both heat and water vapor are transported by the same process near the interface, C_H and C_E have similar variations with roughness. C_E is always higher than C_H due to the difference in κ and ϵ as pointed out by Friehe and Schmitt (1976).

Assuming that T_s is known and the relative humidity is 70%, C_H, C_E and C_D are evaluated with our model for different air-sea temperature differences and plotted versus U_{10} in Figs. 10–12. With only T_w available, the transfer coefficients C_H', C_E' and C_D' evaluated from our model are similarly plotted in Figs. 13, 14 and 15. The values of the transfer coefficients evaluated depend not only on the air-sea temperature differences but also on the absolute values of the air and water temperature. However, the variation due to temperature is small. The variation increases slightly if only T_w is available because the aqueous sublayer whose parameters are more temperature-dependent has to be taken into account.

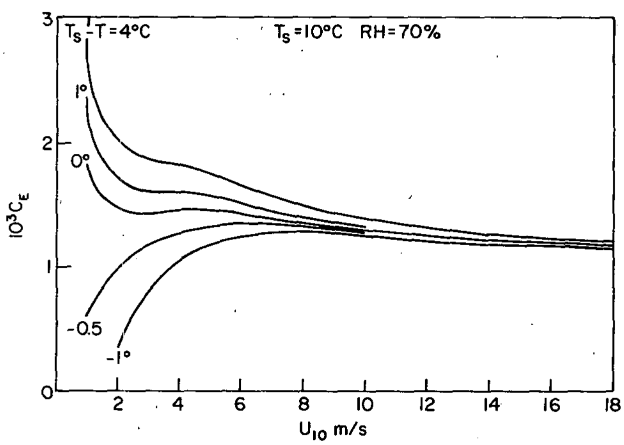

FIG. 11. C_E versus U_{10} under different diabatic conditions evaluated from the present model.

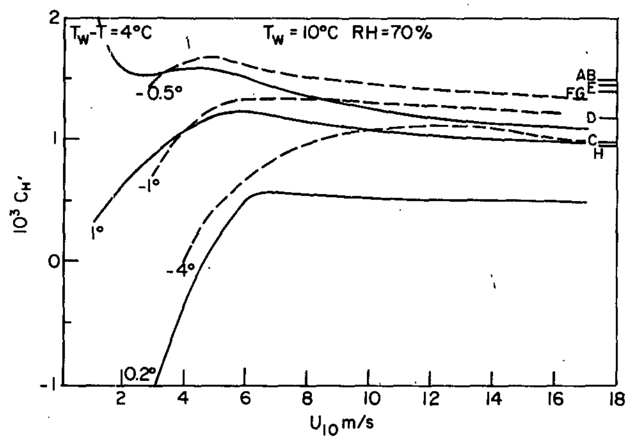

FIG. 13. C_H' versus U_{10} evaluated from the present model compared with the results of other studies (symbols are listed in Table 3).

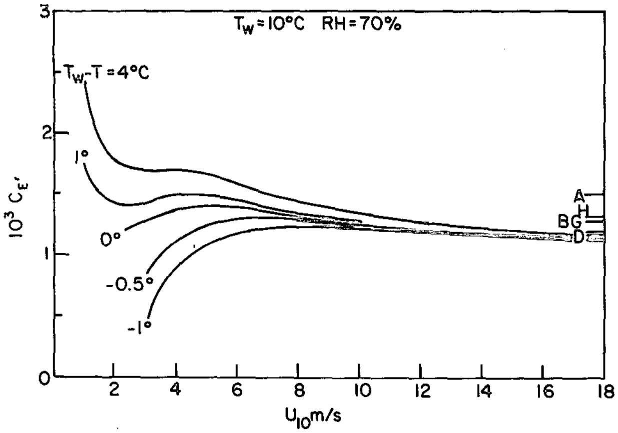

FIG. 14. C_E' versus U_{10} evaluated from the present model compared with the results of other studies (symbols are listed in Table 3).

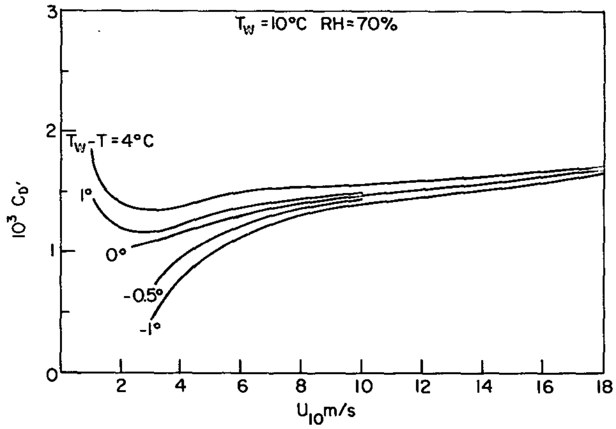

FIG. 15. C_D' versus U_{10} evaluated from the present model.

The diabatic values of the transfer coefficients decrease with increasing stability in agreement with Deardorff (1968). The magnitude of ζ for the same air-sea temperature difference is larger at low wind speed which is part of the reason why the transfer coefficients change more rapidly with air-sea temperature difference at low wind speed. At high wind speeds, they approach their respective neutral values. Under stable condition, assuming $K_H = K_E$ and $b_T = b_U = b$, the criterion for finding a solution to (2) and (4) is $\mathrm{Ri}_v < (\alpha_T b)^{-1}$, where $\mathrm{Ri}_v = g(\partial T_v/\partial z)/[T_v(\partial U/\partial z)^2]$ is the Richardson number for virtual temperature. With the stability functions represented by (4), $(\alpha_T b)^{-1}$ is the asymptotic value of Ri_v for large ζ. The present model fails to give a physically meaningful solution when Ri_v exceeds this value. This occurs when the air is much warmer than the sea or when the wind speed is small. Under these conditions turbulence is suppressed by stable stratification. The values of C_H and C_H' under these conditions are not shown in Figs. 10–15.

For slightly stable and unstable conditions over a moderate range of wind speed, C_D, C_H, C_E, C_D' and C_E' can be approximated by a constant. When T_w is used instead of T_s, there is no large difference between C_D and C_D' or C_E and C_E', but the behavior of C_H' is quite different from that of C_H. When $(T_w - T)$ is small and positive, the latent heat flux and radiative flux may maintain a cool film on the water surface. $(T_s - T)$ may be negative, causing a downward sensible heat flux and consequently C_H' will be negative. Alternatively, when solar radiation is strong, it is possible to have a warm film on the surface causing a finite upward sensible heat flux with $(T_w - T) \leqslant 0$. This phenomenon is noted by Friehe and Schmitt (1976). As $(T_s - T)$ approaches zero, H approaches zero too, and Eq. (1) indicates that C_H varies slowly and becomes undefined. At $(T_s - T) = 0$, C_H is finite and is given by (21) with additional stability terms. However, as $(T_w - T)$ approaches zero, H does not approach zero and Eq. (5) shows that C_H' varies at an increasing rate and finally becomes infinite. Fig. 13 shows that for a small air-sea temperature difference, C_H is very sensitive to small changes in this quantity. Therefore, it is understandable that some investigator (e.g., Pond *et al.*, 1974; Smith, 1974) discarded data with air-sea temperature differences less than 0.5°C in their evaluation of the correlations between H and $U(T_w - T)$. Since in most cases, the atmosphere above the ocean is near neutral, it may not be wise to assume a constant heat transfer coefficient.

Table 3 is a list of empirically determined values of C_H' and C_E' which are shown on the right margins of Figs. 13 and 14 for comparison. Taking into consideration that the values in Table 3 are averages of widely scattered data over a range of wind speed and stability, they do agree with our model. Over a range of moderate wind speed and for slightly unstable conditions, the average value of C_H is about 1×10^{-3} and C_E is about $1.2 - 1.3 \times 10^{-3}$, in close agreement with the results of Friehe and Schmitt (1976).

Four sets of data with moderate wind speeds (2–10 m s^{-1}) are used to compare with the models. Since the surface temperature is not measured by radiometer, it is assumed to be the bulk value T_w.

TABLE 3. A list of empirically determined transfer coefficients.

Data source	$10^3C_H'$	$10^3C_E'$
A. Pond *et al.* (1974)	1.5	1.5
B. Dunckel *et al.* (1974)	1.5	1.28
C. Müller-Glewe and Hinzpeter (1974)	1.	—
D. Smith (1974)	1.2	1.2
E. Smith and Banke (1975)	1.46	—
F. Sahashi and Maitani (1975)	1.4	—
G. Tsukamoto *et al.* (1975)	1.4	1.28
H. Friehe and Smith (1976)	0.97	1.32

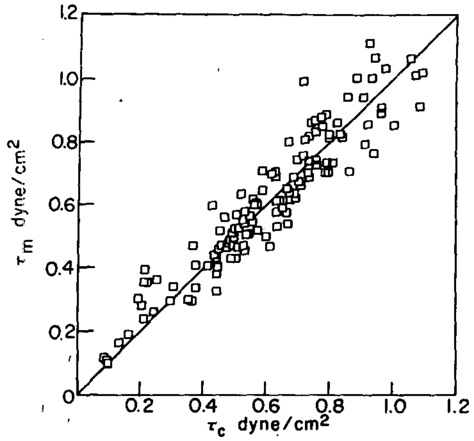

FIG. 16. Stress (τ_m) measured by Paulson *et al.* (1972) compared with that (τ_c) evaluated with the present model.

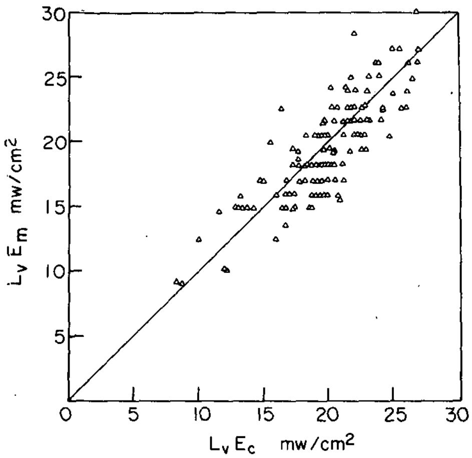

FIG. 18. Latent heat flux (L_vE_m) measured by Paulson *et al.* (1972) compared with that (L_vE_c) evaluated with the present model.

Except for the data by Paulson *et al.* (1972), the air-water temperature difference is given rather than the temperatures of air and water. In such cases, T_w is assumed to 10°C. Since the humidity is not given by Hasse (1970) or Müller-Glewe and Hinzpeter (1974), the relative humidity at instrument height is assumed to be 70%.

In Figs. 16–18, the stress, sensible and latent heat fluxes calculated from the bulk parameters with our model (τ_c,H_c,L_vE_c) are compared with those obtained from profiles measured by Paulson *et al.* (1972) in the BOMEX experiment (τ_m,H_m,L_vE_m). When the profile technique of flux determination is used, only the gradients in the turbulent part of the surface layer are required with no assumption necessary on the sublayer distribution. Fig. 19 compares sensible heat fluxes calculated from our model with those obtained with the eddy correlation method by Hasse (1970), Müller-Glewe and Hinzpeter (1974) and Smith (1974). The values of measured stress and latent heat flux agree reasonably well with model predictions. The measured values of sensible heat flux, although scattered, do agree in general with these calculated ones. However, at low heat fluxes, the data of Paulson *et al.* (1972) and Müller-Glewe and Hinzpeter (1974) are in general higher than the predictions. At low heat fluxes, the air-sea temperature differences are smaller, and any systematic error or instrumental error (Schmitt *et al.*, 1978) will cause a large percentage error in the results. Furthermore, in estimating net radiation from bulk parameters with our model, solar radiation is neglected and T_s and H may be consequently underestimated. This is particularly significant at low heat fluxes.

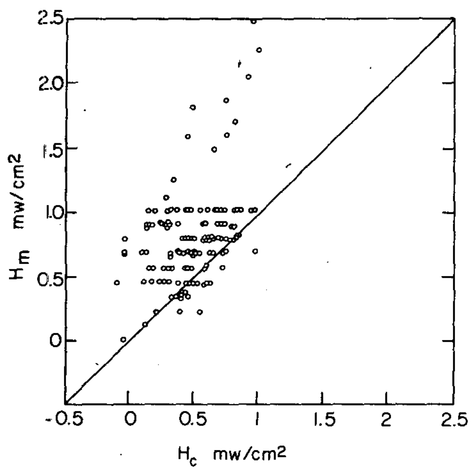

FIG. 17. Sensible heat flux (H_m) measured by Paulson *et al.* (1972) compared with that (H_c) evaluated with the present model.

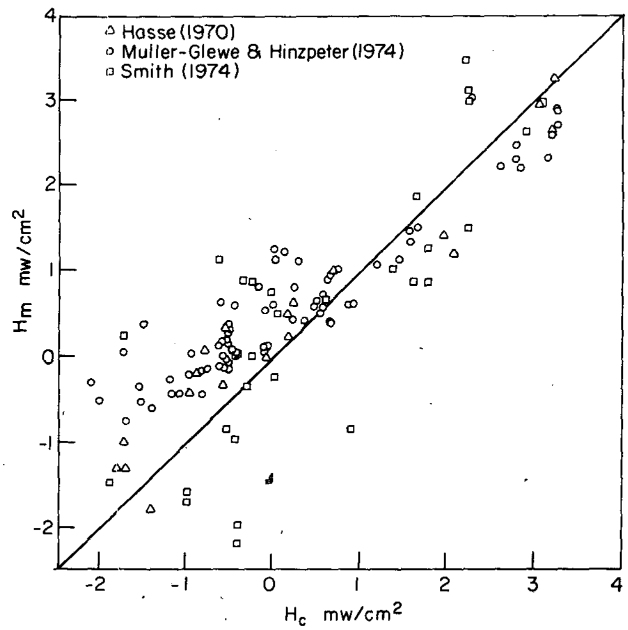

FIG. 19. Sensible heat flux (H_m) obtained by eddy correlation method compared with that (H_c) evaluated with the present model.

TABLE 4. Root-mean-square differences (mW cm^{-2}) between heat flux measured and model prediction.

Data source	rms difference		
	Our model	$C'_H = 1.5 \times 10^{-3}$	$C'_H = 0.97 \times 10^{-3}$
Hasse (1970)	0.54	0.99	0.55
Paulson *et al.* (1972)	0.45	0.35	0.46
Müller-Glewe and Hinzepeter (1974)	0.67	0.81	0.48
Smith (1974)	0.85	0.96	0.92

The rms differences between the measured heat fluxes and those evaluated with our model and two different values of C'_H are shown in Table 4. For two sets of data, our model gives the least rms differences. For the other two sets, our model gives the second least differences and each of the two values of C'_H gives the least difference for one set of data. The value of $C_H = 0.97 \times 10^{-3}$ also gives good agreement with these data. Among the four experiments, only Paulson *et al.* (1972) have evaluated enough moisture fluxes for meaningful comparison. Table 5 shows that our model gives the best representation. In comparison with constant transfer coefficients, our model gives satisfactory representation of the data. The evaluation of heat and moisture fluxes are related to the evaluation of stress through stability. Table 6 shows that our model generally gives better representation than $C_D = 1.5 \times 10^{-3}$. Furthermore, by adding the stability effects to (16a, b, c), our model improves the determination of stress. In evaluating the rms differences in Tables 4–6, only data with absolute values of $T_w - T$ larger than 0.5°C are used.

Our model is based on matching profiles (2) in the outer part of the atmospheric surface layer where the molecular effects can be neglected to profiles (8) in the inner part where molecular constraints are important. It enables one to determine air-sea exchanges of momentum, heat and moisture from bulk parameters. Unlike models with constant transfer coefficients, it includes the effects of stability and interfacial conditions. It should be applicable in approximately stationary and horizontally homogeneous conditions over an open fetch of water and with moderate wind. At high wind speeds, breaking waves disturb the sublayer, making the application of this model questionable. Furthermore, water droplets produced by sea spray may alter the temperature and humidity fields in the surface layer (e.g., Ling and Kao, 1976) and such effects are not included in this model. Our model includes only a simple way to approximate surface water temperature from bulk values. Improvement in the accuracy of measured sea surface temperature will improve the bulk parameterization.

TABLE 5. Root-mean-square differences (mW cm^{-2}) between latent flux predicted by different models and measured by Paulson *et al.* (1972).

Present model	C_E		
	1.5×10^3	1.3×10^3	1.0×10^3
2.23	2.62	2.80	6.25

TABLE 6. Root-mean-square differences between stress (dyn cm^{-2}) predicted by models and measured by Hasse (1970)—I, Paulson *et al.* (1972)—II and Smith (1974)—III.

Data	Present model with z_0 from			$C_D = 1.5 \times 10^{-3}$	C_D from		
	(16a)	(16b)	(16c)		(16a)	(16b)	(16c)
I	0.23	0.29	0.33	0.40	0.26	0.31	0.35
II	0.14	0.10	0.08	0.09	0.18	0.13	0.10
III	0.09	0.12	0.16	0.22	0.10	0.12	0.16

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