Variation of the Heat Transfer Coefficient with Environmental Parameters

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ABSTRACT

Experimentally determined coefficients of the sensible heat flux across the air–sea interface are shown to vary with both wind velocity and difference in temperatures between the sea surface and the 10-m elevation. A simple formula is proposed to associate the heat transfer coefficient with the product between the wind velocity and the temperature difference. The formula appears to be physically sound and also represents well all available data.

1. Introduction

Transfers of momentum and heat across the air–sea interface drive, and therefore link, atmospheric and oceanic systems. Much more, however, has been studied on the momentum than on the heat flux. Earlier results of the heat transfer obtained under various wind and atmospheric stability conditions were compiled and reviewed by Friehe and Schmitt (1976). An empirical formula of the heat transfer coefficient was proposed, by considering that the coefficient was largely independent of the wind velocity; this formula has been widely used. It was supplemented later by results of Smith (1980, 1988), who also adopted the wind-independent concept in analyzing his data.

Following a firmly established increase of the wind stress coefficient with wind velocity (Garratt 1977; Smith 1980; Wu 1980), the constancy of the heat transfer coefficient may require reexamination. In addition, behavior inconsistent with data is found to be implicitly contained in the proposed formulas. Furthermore, the idea of having a constant roughness length of the thermal boundary layer over the sea surface was advanced by Large and Pond (1982). Consequently, data on the sensible heat flux compiled by Friehe and Schmitt (1976) and collected by Smith are reevaluated. The heat transfer coefficient is now seen to vary systematically with not only the difference between temperatures at the mean sea surface and at 10 m above, but also the wind velocity. An empirical formula is proposed, with the transfer coefficient varying continuously with the product between the wind velocity and temperature difference. The formula appears to be physically sound and represents these data well. It also appears to be consistent with more recent data (Large and Pond 1982; Smith and Anderson 1984; Geernaert et al. 1987; Smith 1988). The results of Large and Pond on the constant roughness length of the thermal boundary layer are also discussed.

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2. Previous results on heat-transfer coefficients

a. General definitions

The heat flux across air–sea interface can be expressed as

$$H = -(\lambda + \rho c_p K_t)\partial T/\partial z = H_0 = -\rho c_p u_* t_*,$$
$$K_t = -\overline{t'w'}/\partial T/\partial z, \quad (1)$$

where H and T are the sensible heat flux and potential air temperature at the elevation z above the mean sea surface, and the subscript 0 indicates the surface value; c_p is the specific heat of air at a constant pressure; ρ and λ are the density and thermal conductivity of air, respectively; t' and w' are temperature and vertical velocity fluctuations; u_* is the wind friction velocity; t_* is the scaling temperature; and K_t is the turbulent diffusivity; the overbar indicates the temporal average.

In the atmospheric surface layer, the molecular term in Eq. (1) is negligible in comparison with the turbulent term. The heat flux across the air–sea interface is commonly represented by the following bulk formula:

$$C_t = \mathrm{St} = H_0/\rho c_p U_z(T_0 - T_z) = \overline{t'w'}/U_z(T_0 - T_z), \quad (2)$$

where C_t is the coefficient of sensible heat transfer and is also called the Stanton number (St); the subscript z is attached to U and T to indicate the elevation of their measurements.

b. Adopted models

Direct measurements of $\overline{w't'}$ from field studies of Hasse (1970), Müller-Glewe and Hinzpeter (1974), Dunckel et al. (1974), Mitsuta and Fujitani (1974), and Smith and Banke (1975) were compiled by Friehe and Schmitt (1976) along with their own data. "Cold spikes," consisting of brief periods of colder-than-ambient air measured by the temperature probe, were detected in some investigations. The data known to be contaminated by cold spikes were excluded by Friehe and Schmitt. Finally, a formula was proposed by them to represent the data obtained under different wind and atmospheric stability conditions:

$$\overline{w't'} = A + C_t' U_{10}\Delta T, \tag{3}$$

in which C_t' is used to differentiate it from C_t in Eq. (2), and $\Delta T = T_0 - T_{10}$ with the subscript of T_{10} as well as that of U_{10} indicating the elevation of their measurements at 10 m above the mean sea surface.

Since variations of $\overline{w't'}$ with $U_{10}\Delta T$ may differ for stable and unstable cases, Eq. (3) was fitted by Friehe and Schmitt not only to the entire dataset but also separately to data obtained under stable and unstable conditions. The bulk of the data compiled by Friehe and Schmitt was over the range of $U_{10}\Delta T$ from -14 to 23 m s^{-1} °C. The coefficients determined by fitting Eq. (3) to this main portion of the data, consisting of 116 points, are discussed here. Smith and Banke's (1975) data, consisting of only 14 points, provided the high-wind extension of Friehe and Schmitt's review. This portion is not included, as it was questioned subsequently by Smith (1980). Moreover, a much more comprehensive set of data for high winds, consisting of 87 points, was reported by Smith with $U_{10}\Delta T$ varying from -163 to 131 m s^{-1} °C. This group of data is discussed here, as they were analyzed in a similar fashion to those in Friehe and Schmitt. Values of coefficients A and C_t' reported by Friehe and Schmitt and by Smith are compiled in Table 1.

TABLE 1. Previous results on the sensible heat transfer. The dimension of coefficient A is m s^{-1} °C.

Investigation	Coefficients (10^{-3})	Conditions	
		Stable	Unstable
Friehe and Schmitt (1976)	A	2.60	1.80
	C_t'	0.86	0.97
Smith (1980)	A	−0.10	3.20
	C_t'	0.83	1.10

3. Variation of heat transfer coefficient with environmental parameters

a. Internal inconsistency hidden in current models

The formulas proposed by Friehe and Schmitt (1976) have been commonly used, while the experiment of Smith (1980) has provided the most comprehensive coverage of environmental conditions. Equation (3), adopted in both studies to represent their data, can be rewritten as

$$C_t = C_t' + A/(U_{10}\Delta T). \tag{4}$$

With the coefficients shown in Table 1, such formulas proposed by Friehe and Schmitt and by Smith are diagrammed in Fig. 1, where the lines extend over only the range of $U_{10}\Delta T$ covered by their data. A logarithmic scale is used in the figure to make a clear presentation of two groups covering rather different ranges of $U_{10}\Delta T$. At very small magnitudes of $U_{10}\Delta T$, the transfer coefficient proposed by Friehe and Schmitt shifts abruptly from $-\infty$ to ∞ from stable to unstable conditions, while those by Smith approach ∞ for both conditions. [These infinite values are also noted by Liu et al. (1979) but not seen in Fig. 1, as its horizontal axis is in the logarithmic scale.] At large magnitudes of $U_{10}\Delta T$, the transfer coefficient proposed by Smith has nearly constant values, one for stable and the other for unstable conditions. Those proposed by Friehe and Schmitt, on the other hand, are still in the regions having rapid variations. The above trends contained in these formulas are probably not appreciated by many of their users.

Friehe and Schmitt considered C_t' as C_t, but stated "although it only applied when $A = 0$." Such a difficulty, of course, is greatly exemplified with the abrupt shifting of the C_t value from $-\infty$ to ∞ at very small magnitudes of $U_{10}\Delta T$. Two regions of nearly constant values identified with Smith's (1980) formula are the results of assuming a wind-independent transfer coefficient. A constant value was subsequently assigned by Smith (1988) to the coefficient under neutral atmospheric conditions with $U_{10}\Delta T = 0$ m s^{-1} °C to circumvent this problem; the average of two values at large magnitudes of $U_{10}\Delta T$ shown in Fig. 1 was assigned. Transfer coefficients under various wind velocities and air–sea temperature differences were tabulated by him; see Fig. 2, where the same symbol is used for a given wind velocity. Smith's (1980) curves shown in Fig. 1 are also drawn in Fig. 2; the range of $U_{10}\Delta T$ covered by his experimental conditions is seen to occupy less than the middle one third of that shown in Fig. 2. Surprisingly, the tabulated results, which were calculated largely on the basis of these curves, deviate notably from them. Some other interesting trends shown in Fig. 2 will be discussed in a later section.

Note again that we discuss first here the data and formulas reported by Friehe and Schmitt (1976) and Smith (1980), as their data were analyzed in the same fashion and their formulas have been widely used. The work of Large and Pond (1982) featuring the constant roughness length of the thermal boundary layer over the sea surface will be discussed separately.

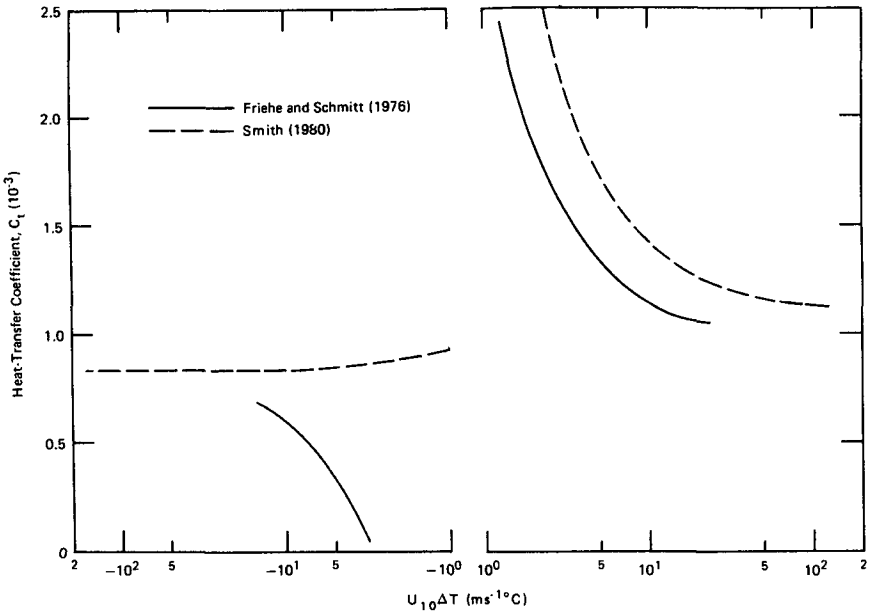

FIG. 1. Formulas of the heat transfer coefficient proposed by Friehe and Schmitt (1976) and Smith (1980).

b. Wind dependence expected for transfer coefficient

The data compiled in Friehe and Schmitt's (1976) review were processed on a general concept: under nearly neutral atmospheric conditions the heat transfer coefficient should behave similarly to the wind-stress coefficient, which was considered then to be invariant with the wind velocity. This group of data was greatly supplemented by Smith (1980), who widened by about eightfold the range of $U_{10}\Delta T$. Similar procedures were

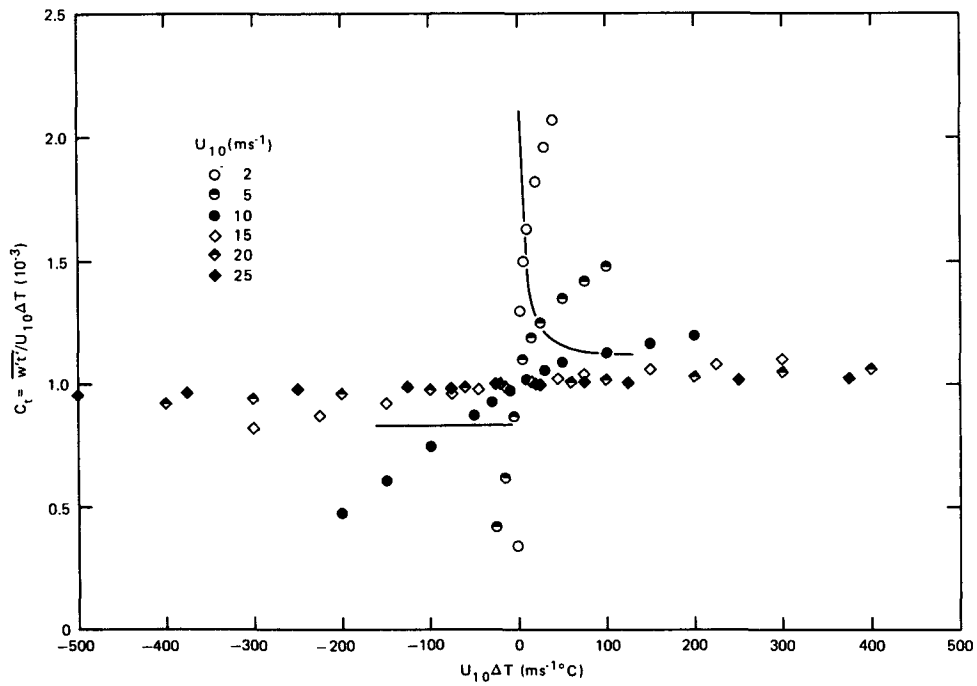

FIG. 2. Heat transfer coefficients tabulated by Smith (1988). The solid line represents the formula proposed by Smith (1980).

adopted by Smith in processing his data, as well as in more recent investigations (Smith and Anderson 1984; Geernaert et al. 1987), with the dependency of transfer coefficient on the wind velocity being somewhat suppressed.

A consensus of the wind-stress coefficient increasing with the wind velocity has been reached subsequently (Smith and Banke 1975; Garratt 1977; Smith 1980; Wu 1980). The increase of wind-stress coefficient with the wind velocity is principally due to the increase of form drag associated with the growth of roughness elements with the wind. Although the heat flux right at the sea surface is not associated with the form drag as closely as the momentum flux, the heat transfer from a rough surface has been shown to be governed by the roughness Reynolds number (Owen and Thompson 1963; Yaglom and Kader 1974); the latter is defined as $u_* z_0/\nu$, where z_0 is the roughness length of sea surface and ν the kinematic viscosity of air. The roughness length represents the characteristic height of roughness elements, which provide the form drag (Wu 1968). As for the other parameter in $u_* z_0/\nu$, the wind friction velocity is already involved prominently in the formulation of heat transfer coefficient in Eq. (1). The rate of heat removal from the sea surface is determined by turbulence structures and velocity gradients of airflows near the sea surface; both parameters are scaled by the wind friction velocity. Along the same line, the following was deduced by Roll (1965) and adopted by Large and Pond (1982):

$$C_t = \kappa^2/[\ln(Z/z_0)\ln(Z/z_t)] = \kappa C_{10}^{1/2}/\ln(Z/z_t), \quad Z = 10 \text{ m}, \quad (5)$$

where κ is the von Kármán constant, $C_{10} = (u_*/U_{10})^2$ is the wind-stress coefficient, z_t is the roughness length of thermal boundary layer, and Z is the standard anemometer height at 10 m above the mean sea surface. The functional variation of heat transfer coefficient with wind velocity is clearly seen in the above expression to be influenced by that of the wind-stress coefficient. Such a variation, as illustrated in Eq. (5), is not inconsistent with the concept of adopting a constant "thermal roughness length." The detailed variation of the heat transfer coefficient with wind velocity in this line of thinking will be discussed in a later section.

Note that we have discussed so far only the direct influence of wind velocity, not its indirect influence; the wind velocity is also involved in the parameterization of atmospheric stability effects.

c. *True trend indicated by reported data*

Inasmuch as the data have played a pivotal role in deducing the empirical formula (Friehe and Schmitt 1976; Smith 1980), let us examine now the trend of the data. Values of the transfer coefficient, as defined in Eq. (4), are obtained from the data compiled by Friehe and Schmitt and collected by Smith; see Fig. 3a,b. Quite expectedly, the results at very small magnitudes of $U_{10}\Delta T$ are seen to scatter over a wide range. For the formula shown in Eq. (3) to be applicable to all wind and atmospheric stability conditions, the formula in the rewritten form of Eq. (4) should have the following trends for both stable and unstable conditions: $\overline{w't'}/U_{10}\Delta T$ has a very large value when $U_{10}\Delta T$ is near zero and decreases as the magnitude of $U_{10}\Delta T$ increases to approach asymptomatically the value of C_t' shown in Table 1. The trend near the origin is due to a positive flux at $U_{10}\Delta T = 0$ m s^{-1} °C discussed in Friehe and Schmitt, while the asymptotical nature was illustrated by Smith's formula diagrammed in Fig. 1. The results somewhat away from the origin of $U_{10}\Delta T$ are seen in Fig. 3, however, to follow an overall tilting trend as indicated by the line drawn. In other words, instead of approaching asymptotically the constant value as proposed by Friehe and Schmitt and by Smith, the coefficient C_t varies systematically with $U_{10}\Delta T$. This trend, however slight, is important in discussing the results at small (data scatter) and large (systematic trend) magnitudes of $U_{10}\Delta T$.

In summary, transfer coefficients were defined by

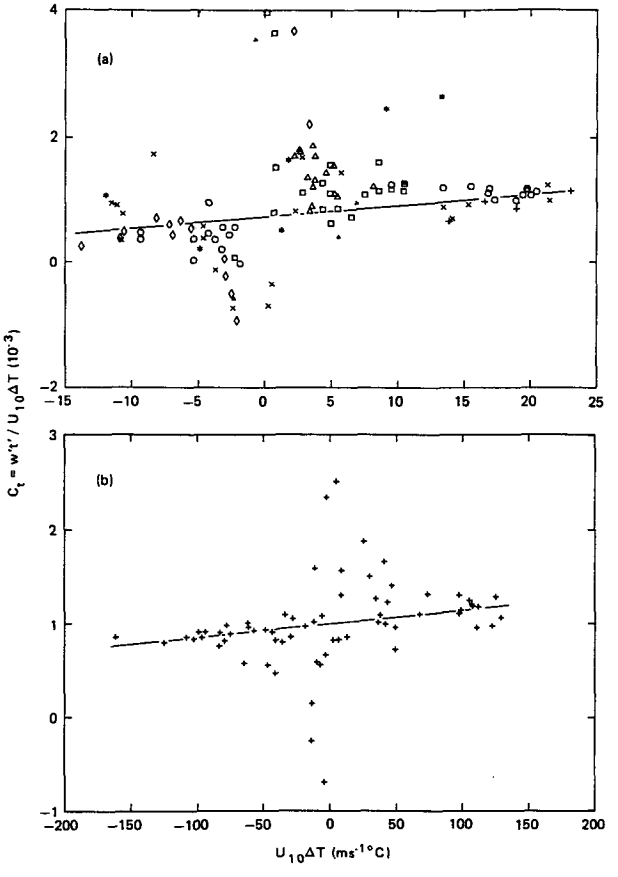

FIG. 3. Heat transfer coefficients (a) compiled by Friehe and Schmitt (1976) and (b) measured by Smith (1980).

Friehe and Schmitt (1976) and by Smith (1980) in a rather special form, but the conventional definition is adopted here. Unlike the trends proposed by them, there are signs of variations of the coefficient with the product $U_{10}\Delta T$. Smith's (1988) results at high winds are also seen in Fig. 2 to asymptotically approach this trend, while his results at light winds approach the origin in a fashion opposite to that shown in Fig. 1. All these trends are inconsistent with the concept of wind-independent transfer coefficient contained in their proposed formulas. The trends demonstrated in Figs. 2 and 3 appear then to indicate a systematic variation of the transfer coefficient with wind velocity. Because of the very nature of its definition, the coefficient at small magnitudes of $U_{10}\Delta T$ obtained by dividing the heat flux with a smaller product of $U_{10}\Delta T$ is prone to have a greater error. In addition, the resolution of temperature measurements was generally poor, as noted by Large and Pond (1982), while the bucket temperature of subsurface water has been measured in almost all experiments, despite that the sea surface temperature is used in the formulation.

d. Approximate formula proposed for transfer coefficient

The zone of data uncertainty is seen in Fig. 3a to fall in the region from approximately -3 to 5 m s^{-1} °C, and in Fig. 3b from -20 to 40 m s^{-1} °C. This discrepancy is due mainly to different ranges of the wind velocity covered by two groups of data; it was extended from about 8 m s^{-1} in Friehe and Schmitt's (1976) collection to 22 m s^{-1} in Smith's (1980) experiment. In other words, error bands of measuring the air–sea temperature difference were probably the same in both sets of data; differences in the data scatter shown in Fig. 3a,b are due to multiplying the error in temperature measurements by different ranges of wind velocities.

The coefficient of sensible heat transfer has always been considered to have finite values for nearly neutral atmospheric conditions. For example, in terms of 10^{-3} they are: 1.0 (Hasse 1970), 1.4 (Hicks 1972), 1.2 (Smith 1974), 1.5 (Pond et al. 1974), 0.71 (Geernaert et al. 1987), and 1.0 (Smith 1988). Two values, one for stable and the other for unstable conditions, were provided by Smith (1980): 0.83 and 1.10; and by Large and Pond (1982): 0.66 and 1.13. All these investigators avoided the infinite value of C_t for neutral conditions. The two-value proposal, of course, provides a closer representation of the data at large magnitudes of $U_{10}\Delta T$ than does the single-value one. It is, however, rather hard to understand that at ΔT near 0°C the coefficient shifts abruptly from one value for stable conditions to another for unstable conditions; this objection was also raised by Large and Pond. The average value of stable and unstable coefficients is adopted in the single-value model; the latter provides a poor representation at large magnitudes of $U_{10}\Delta T$, but avoids the sudden shift.

Before further studies can be conducted to clarify irregularities discussed above at very small magnitudes of $U_{10}\Delta T$, it may be advisable to determine the transfer coefficient from the trend of data detected in Fig. 3. Moreover, if a finite value is chosen for $\Delta T = 0$°C as accepted in most studies, it appears entirely reasonable to consider that the transfer coefficient varies continuously with $U_{10}\Delta T$ over the region of small magnitudes of $U_{10}\Delta T$. In later sections, such a variation will also be shown to be consistent with the data collected in recent studies (Large and Pond 1982; Smith and Anderson 1982; Geernaert et al. 1987). Straight lines are then fitted to the results shown in Fig. 3a,b; these lines over two different wind velocity ranges can be represented by

$$C_t = (0.720 + 0.0175U_{10}\Delta T) \times 10^{-3} \qquad U_{10} < 8 \text{ m s}^{-1} \quad (6)$$

$$C_t = (1.000 + 0.0015U_{10}\Delta T) \times 10^{-3} \qquad U_{10} > 8 \text{ m s}^{-1}, \quad (7)$$

where U_{10} and ΔT are expressed in meters per second and degrees Celsius, respectively. The division at $U_{10} = 8$ m s^{-1} for low and high winds is based on the upper wind velocity limit of Friehe and Schmitt's compilation. The reason for discrepancies between two sets of results is not entirely clear at this stage.

4. Discussion

a. Justification of scaling parameter $U_{10}\Delta T$

We are encountering here a rather complicated situation, with the transfer coefficient varying with several parameters, including those governing wind–wave interaction (the wind velocity, fetch, and wave age), and those governing stability effects (the Monin–Obukhov length encompassing the wind velocity as well as the air–sea temperature difference). We realize that the common approach to evaluate the dependence of transfer coefficient on the wind velocity is to first remove its variation with stability conditions. The supporting data, however, were not collected in most investigations; as for the present analysis, U_{10} and ΔT are also not independently available. Furthermore, although advances have been made in correcting stability effects on the wind-stress coefficient (Dyer 1974), it is less certain about similar corrections of the heat transfer coefficient. Their poorer states of understanding are discussed herewith throughout the article.

Strictly speaking, the nondimensional coefficient C_t should also be regressed against a nondimensional variable instead of $U_{10}\Delta T$. Discussions were presented earlier (Wu 1986) on a nondimensional grouping of U_{10} and ΔT to represent the Monin–Obukhov length in the form of $\Delta T/U_{10}^{2.75}$. The very parameter $U_{10}\Delta T$, adopted here from Friehe and Schmitt (1976) and Smith (1980), is intended, however, to represent effects

of not only the stability length but also the wind velocity. The heat transfer coefficient should increase with the wind velocity in accordance with the roughness growth as discussed in a previous section; it should decrease with the wind velocity in accordance with the stability effect as discussed above. In comparison with the wind-stress coefficient, the former variation with the wind velocity is weaker, while the latter is probably stronger. Their combined effects can apparently be represented by the parameter $U_{10}\Delta T$, as displayed by systematic variations shown in Fig. 3.

Let us now examine further the use of $U_{10}\Delta T$ separately in variations of the heat transfer coefficient with the wind velocity and with the air–sea temperature difference. For a given temperature difference, the heat transfer coefficient is seen in Fig. 3 and Eqs. (6) and (7), as expected, to increase with the wind velocity under unstable conditions. However, it decreases under stable conditions as the wind velocity increases. This is believed to be due to a combination of direct influence of the wind velocity on the roughness growth and its indirect influence included in the stability effect. The air–sea temperature difference, on the other hand, is included only in the stability effect. In this case, the heat transfer coefficient under a given wind velocity should decrease under stable conditions and increase under unstable conditions with the increasing magnitude of ΔT. These reflect exactly stability effects associated with the air–sea temperature difference.

Proposed formulas of the wind-stress coefficient have reached a common form, especially its rate of increase with the wind velocity. This rate was suggested by Smith and Banke (1975), Garratt (1977), and Wu (1980) as $dC_{10}/dU_{10} = 0.066$, 0.067, and 0.065×10^{-3} m^{-1} s, respectively. These values, generally for relatively high winds, are much greater than that indicated in Eq. (7) for any given rational temperature difference; for example, $dC_t/dU_{10} = 0.0075 \times 10^{-3}$ m s^{-1} for $\Delta T = 5$°C, and is of course smaller for smaller magnitudes of $U_{10}\Delta T$. This is reasonable, as we suspected earlier that the wind velocity, causing the increase of form drag, should have a greater influence on the wind stress than the heat transfer coefficient. These trends are consistent with those reported earlier by Kondo (1975) and Liu et al. (1979), confirming that these coefficients have different values in the atmospheric surface layer.

b. *Comparison with other studies on C_t*

As mentioned previously, a comprehensive study on the transfer of sensible heat across the sea surface was conducted by Large and Pond (1982). Data of C_t under open-ocean conditions obtained by them were corrected for stability effects to determine the corresponding value under neutral atmospheric conditions, C_{tn}. The results presented in their Fig. 10 are reproduced in Fig. 4. (One data point for 20–25 m s^{-1} winds was omitted by Large and Pond, as conditions for the averaging process in this case might not have been realized; it is also omitted here.) In accordance with Large and Pond, two solid straight lines are fitted to the data with wind velocities larger than 10 m s^{-1}. The neutral coefficient C_{tn} is seen to increase with U_{10} for a large portion of the data under unstable conditions and decrease as U_{10} increases under stable conditions. These trends are, of course, consistent with those proposed here. Very interestingly, both segments of lines, extended here with dashed lines, intersect at $U_{10} = 0$ m s^{-1}. This point, which was not discussed by the original authors, serves to eliminate the abrupt change between stable and unstable conditions. The neutral transfer coefficient at $U_{10}\Delta T = 0$ m s^{-1} °C is 0.735×10^{-3}, comparing well with 0.720×10^{-3} shown in Eq. (6).

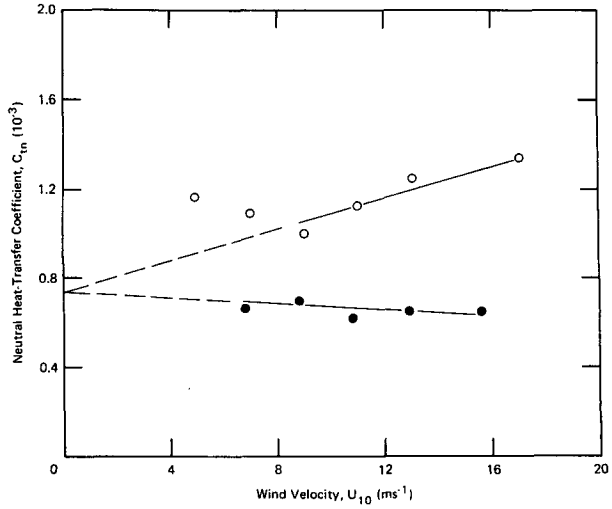

FIG. 4. Variation of the heat transfer coefficient with wind velocity. The data are from Large and Pond (1982); the open circles are for unstable atmospheric conditions and solid circles stable conditions.

To represent their results, Large and Pond (1982) proposed the following:

$$\left.\begin{array}{rl} \text{Stable conditions} & z_t = 2.2 \times 10^{-9} \text{ m} \\ \text{Unstable conditions} & z_t = 4.9 \times 10^{-5} \text{ m} \end{array}\right\}. \quad (8)$$

Earlier, it was suggested that the wind-stress coefficient could be also approximated by (Wu 1969)

$$C_{10} = 0.5 \times 10^{-3} U_{10}^{1/2}, \quad (9)$$

where U_{10} is expressed in m s^{-1}; this kind of variation was also confirmed by Garratt (1977). Substituting Eqs. (8) and (9) into Eq. (5), we have

$$\left.\begin{array}{rl} \text{Stable conditions} & C_t = 0.40 \times 10^{-3} U_{10}^{1/4} \\ \text{Unstable conditions} & C_t = 0.73 \times 10^{-3} U_{10}^{1/4} \end{array}\right\}. \quad (10)$$

The variation of heat transfer coefficient with the wind velocity is seen again to be much weaker than that of the wind-stress coefficient, shown in Eq. (9). As for

the data shown in Fig. 4, the trend for unstable cases is consistent with Large and Pond's proposal shown above and with our Eqs. (6) and (7). On the other hand, the data under stable conditions follow only our proposal; there, the heat transfer coefficient actually decreases as the wind velocity increases. This is not affected by replacing Eq. (9) with the wind-stress coefficient suggested by Large and Pond (1981), as the data are under relatively high wind velocities where the wind-stress coefficient suggested by them also increased with the wind velocity. We believe that this discrepancy is due to an incomplete removal of atmospheric stability effects in determining the neutral coefficient. Such a difficulty was mentioned previously.

The results of Smith and Anderson (1984) and Geernaert et al. (1987) are presented in Fig. 5a,b; both sets are greatly scattered. Lines are drawn in the figure to illustrate their overall trends, showing that C_t increases with $U_{10}\Delta T$. The rates of increase are about 0.022×10^{-3} and 0.015×10^{-3} m^{-1} s °C^{-1} in Fig. 5a,b, respectively. Both rates are quite comparable with the value of 0.0175×10^{-3} m^{-1} s °C^{-1} shown in Eq. (6). In other words, these values determined at small magnitudes of $U_{10}\Delta T$ are actually in rather close agreement.

5. Concluding Remarks

Several implicit trends in the previously proposed formulation on the coefficient of sensible heat transfer are discussed; they do not appear to be physically sound. In the meantime, the increase of heat transfer coefficient with the wind velocity has been herein demonstrated to be real and significant. Although signs of their increases were detected earlier (Francey and Garratt 1979; Large and Pond 1982), these coefficients have been mostly considered to remain constant. The increase of heat transfer coefficient with the wind velocity is also compatible with the trend that the wind-stress coefficient increases with the wind velocity. The variation of the heat transfer coefficient with the air–sea temperature difference is generally known and is of course consistent with atmospheric stability effects on the momentum transfer (Friehe and Schmitt 1976; Smith 1980; Wu 1986). Formulas proposed here appear to be physically sound; a continuous variation of the heat transfer coefficient is also a more logical choice, at this stage anyway with limited data, than large sudden variations at small values of $U_{10}\Delta T$. Finally, presently proposed formulas are substantiated by the data in earlier and recent reports.

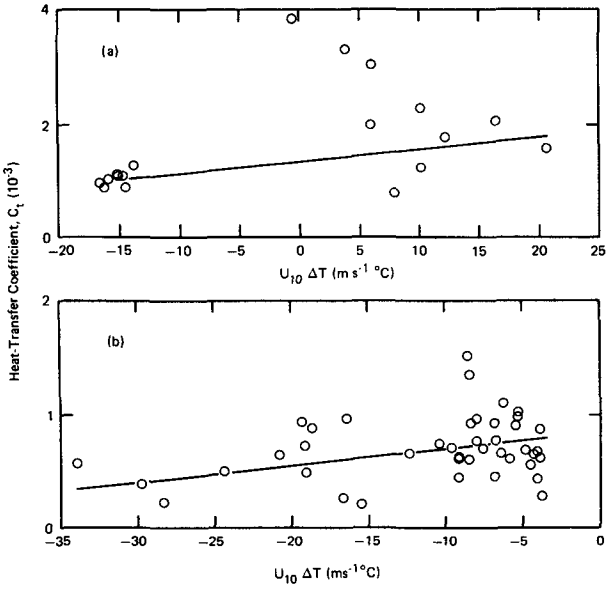

FIG. 5. Variations of other reported Stanton numbers with environmental parameters. The results are from (a) Smith and Anderson (1984), and (b) Geernaert et al. (1987).

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