Parameterization of Air-Sea Interface Fluxes of Sensible Heat and Moisture by the Bulk Aerodynamic Formulas

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ABSTRACT

The parameterizations of the sensible heat and moisture fluxes by the bulk aerodynamic formulas are determined from a compilation of existing data, together with some new results. The data set comprised 152 determinations of the sensible heat flux and 30 of the moisture flux from experiments in which the fluxes were measured directly over water with suitable turbulence instrumentation. Least-square-error fits were performed on the data. The moisture flux (and therefore the latent heat flux) is adequately described by the bulk formula with a coefficient of 1.32×10^{-3}. The parameterization of the sensible heat flux is complicated, for the data show 1) a small positive heat flux for zero temperature difference between the air and sea surface, 2) the coefficient for stable conditions is smaller than for unstable conditions, and 3) the coefficient appears to increase at high wind speeds, as shown by the data of Smith and Banke (1975). Separate bulk formulas are presented for the sensible heat flux for the different conditions of the temperature field.

1. Introduction

Estimations of the air-sea interface fluxes of heat, moisture and momentum by simple measurements of mean quantities and use of the bulk aerodynamic formulas are attractive for several reasons. There exists a large body of observations from "ships of opportunity" which provide a data base from which the large space scale and time history of the flux fields may be estimated by the bulk aerodynamic method. It is impractical at the present time to directly measure the fluxes at more than a few stations at sea in large-scale oceanographic experiments, except possibly by the use of research aircraft. Surface observations from meteorological/oceanographic buoys provide mean measurements suitable for flux estimation by the bulk aerodynamic method. A central question in the field of air-sea interaction has been the determination of the applicability of the bulk aerodynamic formulas and evaluation of associated constants. Recently, a relatively large number of investigations of direct measurements of the sensible heat and moisture fluxes over the sea have been reported, and the purpose of this work is to assemble the data and obtain the parameterizations of the fluxes by the bulk aerodynamic formulas. Data from a recent experiment in the North Pacific (NORPAX POLE) are also reported.

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The sensible heat and moisture fluxes are given by

$$H_s=\rho C_p\overline{w\theta}, \tag{1}$$

$$E=\overline{wq}, \tag{2}$$

where:

- $\overline{w\theta}$ mean vertical velocity-temperature covariance (m K s^{-1})
- $\overline{wq}$ mean vertical velocity-water vapor density covariance (m s^{-1}) (g m^{-3})
- w vertical component of turbulent velocity vector (m s^{-1})
- θ fluctuating temperature (K)
- q fluctuating water vapor density (g m^{-3})
- ρ air density (assumed to be constant) (g m^{-3})
- C_p heat capacity (cal g^{-1} K^{-1})
- H_s sensible heat flux (cal m^{-2} s^{-1})
- E moisture flux (g m^{-2} s^{-1}).

The overbar indicates a time average, usually of the order of 1 h. In this review we shall consider primarily data where the fluxes were obtained by direct measurements of $\overline{w\theta}$ and $\overline{wq}$ with suitable turbulence instrumentation and data processing techniques. There are other means of estimating the turbulence fluxes, such as the mean profile technique and dissipation rate measurements [see, e.g., Pond *et al.* (1971); see for review, Stegen *et al.* (1973)] but such results are not included in the present analysis except for the water vapor profile data of Dunckel *et al.* (1974). The purpose of the present work is to obtain the parameterizations of the bulk aerodynamic formulas rather than compare different flux measurement techniques.

The bulk aerodynamic formulas for estimates of the sensible heat and moisture fluxes are

$$H=\rho C_p C_H \bar{U}(\bar{T}_s-\bar{T}_a), \quad (3)$$

$$E=C_E \bar{U}(\bar{Q}_s-\bar{Q}_a), \quad (4)$$

where:

- $\bar{U}$ mean velocity (m s^{-1}) at reference height h above the sea surface
- $\bar{T}_s$ mean sea temperature (K), usually that obtained about a meter below the surface
- $\bar{T}_a$ mean potential air temperature (K) at height h
- $\bar{Q}_s$ mean water vapor density (g m^{-3}) near the sea surface, usually obtained by assuming the air to be saturated at dewpoint temperature $\bar{T}_d$
- $\bar{Q}_a$ mean water vapor density (g m^{-3}) at reference height h
- C_H sensible heat transfer coefficient
- C_E moisture transfer coefficient.

The reference height h is usually taken to be 10 m above the sea surface. Roll (1965), Kitaigorodskii (1970), and Kraus (1972) discuss the derivations of these formulas. It is usually assumed that the coefficients C_H and C_E are constants and are approximately equal to the drag coefficient for momentum transfer C_D. Values of C_D of $(1.0\text{–}1.5)\times10^{-3}$ have been reported with a small increase with wind speed indicated (Smith and Banke, 1975).

Previous determinations of the transfer coefficients have generally found that they are roughly equal to the drag coefficient. Pond *et al.* (1971), however, reported for data obtained in Project BOMEX, values of C_H about five times those for the moisture flux (C_E), and also the C_H values obtained near San Diego in a pre-BOMEX experiment. The reason for the large C_H values was attributed partly to the presence of "cold spikes" in the temperature field under the BOMEX conditions. The "cold spikes" consisted of brief periods of colder-than-ambient air measured by the temperature probe under unstable conditions. An example of the phenomenon is shown in Fig. 1 from data obtained in the North Pacific POLE experiment for unstable conditions. Two types of temperature

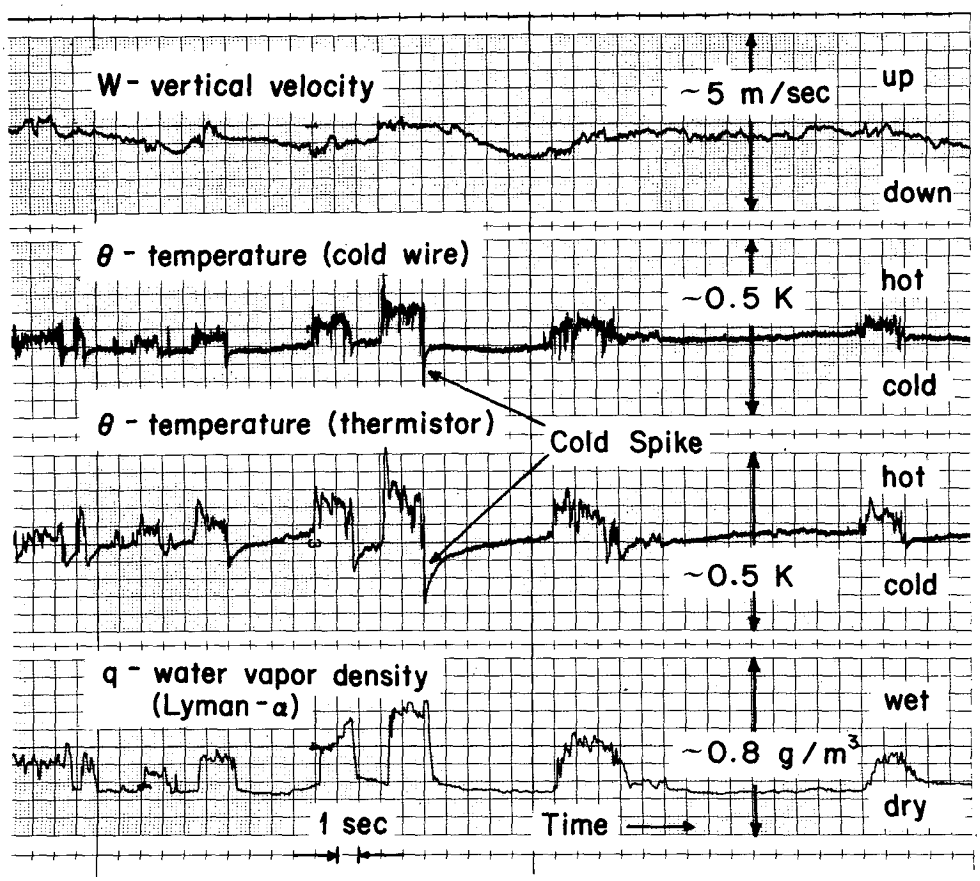

FIG. 1. Time series showing "cold spikes" in the temperature field under unstable conditions. $\bar{U}\approx6$ m s^{-1}, $\Delta T=0.3$ K, $h=29$ m, at 35°N, 155°W, February 1974, on R/V FLIP.

probes—a thermistor and platinum cold wire—sensed "cold spikes." There are no corresponding spikes in the moisture field, also shown in Fig. 1 as measured by a Lyman-alpha humidiometer spaced about 20 cm from the temperature probes. Sometimes the "cold spikes" are not prevalent, as Pond *et al.* (1971) found in the pre-BOMEX experiment near San Diego, nor have they been reported in measurements over land. The reason or reasons for the anomalous behavior of the temperature field have not been determined, although several hypotheses have been advanced. Pond *et al.* (1971) suggested that upper air profiles of temperature and humidity of certain forms could produce the cold spike as the upper level air was mixed by the turbulent motion. They also suggested, as have Coantic and Seguin (1971), that radiative effects may be important in the temperature field, especially in high humidity conditions where longwave radiation is strongly absorbed. Holland (1972) also proposed a model of certain vertical distributions of temperature and humidity and radiative effects to account for differences in energy spectra of temperature and humidity found by Pond *et al.* (1971) during BOMEX. Holland also replotted the results of Pond *et al.* (1971) to show that the sensible heat flux was better correlated with $\bar{U}(\bar{Q}_s-\bar{Q}_a)$ than $U(\bar{T}_s-\bar{T}_a)$ for the BOMEX data. Dreyer (1974) proposed that cold spikes may be due to evaporation of spray droplets in the marine boundary layer. Because of the relatively large heat of vaporization of water, rapid adiabatic evaporation of a drop would produce a significant local cooling of the temperature field, but the amount of water vapor added would be small.

Hicks (private communication) reported observing cold spikes in an ocean boundary layer experiment, but after cleaning the thermistor probe with distilled water, the cold spikes were no longer observed. Hicks concluded that the thermistor element was contaminated by salt nuclei which formed a humidity-sensitive shunt across the thermistor. Hicks also observed that the two temperature signals shown in Fig. 1 imply that the cold spikes are erroneous. The magnitude of the observed spike sensed by the thermistor is approximately equal to the magnitude sensed by the fine platinum cold wire, although the thermistor response is much slower (~5 Hz) than the cold wire (~600 Hz). Low-pass filtering the cold wire signal at ~5 Hz would not give the measured thermistor signal. Therefore, there must exist another mechanism for generating such large temperature signals besides the normal response to the air temperature. However, the humidity-sensitive shunt resistance mechanism would not apply to the results shown in Fig. 1 because the signs of the temperature-resistance sensitivities of the platinum cold wire and the thermistor are opposite: the same shunt resistance applied to each probe would result in an apparent increased temperature in one and an apparent de-

TABLE 1. List of experimental determinations of sensible heat and moisture fluxes.

Investigator	Experiment and/or location	Platform and measurement height	Instruments*	Temperature** signal observations	Remarks	Graph symbol
Müller-Glewe and Hinzpeter (1974)	Baltic	Mast in 12 m water, $z=2$ and 4 m	w: sonic anemometer θ: Pt. wire	Not reported		□ ○ △
Dreyer (1974)	OWAX/near San Diego	FLIP,	w: sonic anemometer θ: Pt. wire q: Lyman-Alpha humidiometer	"cold spikes" observed, unstable	Approximate pitch motion correction to w	×
Friehe and Schmitt	NORPAX/35°N, 155°W	FLIP, $z=29$ m	w: sonic anemometer θ: thermistor q: Lyman-Alpha	"cold spikes" observed in unstable conditions; none in stable conditions	True pitch correction to w	*
Pond *et al.* (1971)	San Diego	FLIP, $z=8.5$ m	w: sonic anemometer θ: Pt. wire q: Lyman-Alpha	Few "cold spikes," unstable conditions		⊕
Pond *et al.* (1971)	BOMEX/Caribbean	FLIP, $z=8.1$ and 8.6 m	Same as above	"cold spikes" observed, unstable conditions	Small sensible heat fluxes; high humidity	⊠
Dunckel *et al.* (1974)	ATEX/Atlantic	Gyro-stabilized buoy, $z=2.4$ m	w: hot-wires θ: Pt. wire q: wet-bulb thermometers (mean only)	Not reported		◇
Hasse (1970)	Baltic	Gyro-stabilized buoy, $z=2.8$ to 5.5 m	w: hot-wires θ: Pt. wire	Not reported	Water depth 14 m	+
Mitsuta and Fujitani (1974)	Pacific	Ship, z not reported; top of foremast	w: sonic anemometer θ: thermocouple q: wet thermocouple	Not reported	Motion correction used	X
Smith and Banke (1975)	Sable Island/Atlantic	Mast on sand bar, $z=10$ m	w: thrust anemometer θ: thermistor	No "cold spikes"	Very high winds	▽

* w indicates vertical turbulent velocity component, θ indicates fluctuating air temperature, q indicates fluctuating water vapor content of air.
** Observations of the time series of the fluctuating temperature signal to determine if it contains anomalous "cold spikes." See reference of Pond *et al.* (1971) and Phelps and Pond (1971) for a discussion of this phenomenon.

creased temperature in the other. Schmitt *et al.* (1976) have proposed that salt spray contamination on the sensor surface can cause cold spikes as moisture evaporates from the salt spray with corresponding latent heat release which is provided by the sensor. This is seen in Fig. 1 where the cold spikes occur when the humidity field suddenly decreases. (There is a small time lag in the humidity signal caused by an aspirator used to protect the Lyman-alpha humidiometer.) Based on the observations of Hicks and the hypothesis of Schmitt *et al.*, it appears that over ocean temperature data which are significantly contaminated by cold spikes are not correct, and therefore we shall not include such sensible heat flux data in the bulk formula parameterization analysis.

Previous reviews of the parameterizations of the sensible heat and moisture fluxes have been made. Pond *et al.* (1974) recommended $C_H = C_E = 1.5\times10^{-3}$ for normal temperature conditions based on the San Diego and Indian Ocean profile data. Hasse (1970) recommended $C_H = 1.0\times10^{-3}$ from measurements in the Baltic Sea for both stable and unstable conditions. Coantic (1974) recommended $C_H = C_E = 1.3\times10^{-3}$ for $\bar{U} < 10$ m s^{-1} and $C_H = C_E = (1.0 + 0.05\,\bar{U})\times10^{-3}$ for $0 < \bar{U} < 20$ m s^{-1}. Smith (1974) found $C_H = C_E = 1.2\times10^{-3}$ in a study over Lake Ontario. Hicks (1972) found $C_H = C_E = 1.4\times10^{-3}$ from three separate experiments over water bodies of various sizes. Sverdrup *et al.* (1942) presented a bulk aerodynamic formula for the yearly average evaporation rate from the oceans, which is equivalent to $C_E = 1.62\times10^{-3}$. Deardorff (1968) predicted the dependence of C_H and C_E on the stability of the boundary layer. The coefficients were predicted to be smaller under stable conditions than under unstable conditions.

2. Data set

A total of nine data subsets were compiled, giving 152 data points for the sensible heat flux and 30 points for the moisture flux. The references, details of instrumentation, experiments and other factors are given in Table 1 for the nine subsets. As shown in Table 1, most of the measurements were not performed at the standard reference height of 10 m, but the results were reported with corrections for the height, except in those cases where the height corrections were considered insignificant (Pond *et al.*, 1974).

3. Analysis

a. Sensible heat flux

The heat flux data that were obtained in light to moderate wind speeds are shown in Fig. 2. In this figure, the directly-measured values of $\overline{w\theta}$ are plotted against $\bar{U}(\bar{T}_s - \bar{T}_a) = \bar{U}\Delta T$. Thus the slope of a line on this plot is C_H. Although there is scatter in the data, the following trends may be observed:

1) There is a positive value of $\overline{w\theta}$ for $\bar{U}\Delta T = 0$. This observation has also been made by Hasse (1970),

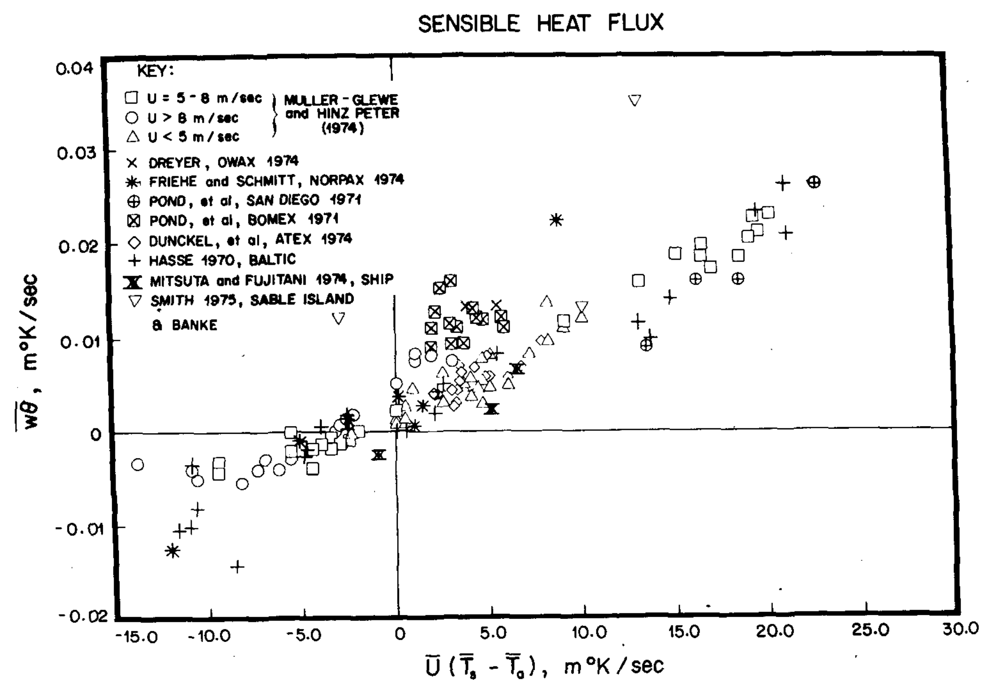

FIG. 2. Sensible heat flux. Directly-measured $\overline{w\theta}$ covariance versus $\bar{U}(\bar{T}_s - \bar{T}_a)$. Slope of a line on the graph is C_H.

Dunckel *et al.* (1974) and Müller-Glewe and Hinzpeter (1974).

2) The slope of a line through the stable data ($U\Delta T<0$) appears to be less than that for the unstable data (Müller-Glewe and Hinzpeter, 1974).

3) For unstable conditions at low heat flux levels, there are some data points which are anomalously high: Pond *et al.* (1971) BOMEX; Dreyer (1974) and some of the present data. For these data anomalous "cold spikes" were observed. The 22 "cold spike" data points are not included in the following analysis.

Least-square fits were performed on the data with the formula

$$\overline{w\theta}=A+C_H\bar{U}\Delta T. \tag{5}$$

Here we have retained the symbol C_H for the coefficient of $\bar{U}\Delta T$, although according to the bulk formula (3) it only applies when $A=0$.

The data were separated into various subclasses—stable, unstable, and with and without the inclusion of the high wind speed data of Smith and Banke (1975). The results of the least-squares analyses on each subset and the entire data set are given in Table 2. From the results it appears that, in general, the observations from Fig. 2 are supported to within about one standard deviation of the least-squares fits. In particular:

(i) For all of the data a small positive value of $\overline{w\theta}$ at $\bar{U}\Delta T=0$ is indicated, about $+0.0012$ m K s^{-1}. The data of Smith and Banke for low flux levels ($\bar{U}\Delta T<20$) also indicate a small positive intercept of about the same value. The intercept value for the data for stable conditions is approximately the same as the value for the data under unstable conditions without the large $\bar{U}\Delta T$ data of Smith and Banke, i.e., the parameterization is continuous through $\bar{U}\Delta T=0$.

(ii) The coefficient C_H for the stable data (0.86×10^{-3}) is marginally smaller than the value for the comparable unstable moderate wind speed data, 0.97 $\times10^{-3}$. A comparison to the wide-range unstable data of Smith and Banke (1975) indicates that the value for C_H is significantly smaller for stable conditions (0.86×10^{-3} compared to 1.46×10^{-3}). Here the distinction between stable and unstable conditions is made on the sign of ΔT. In many of the experiments, only temperature was measured and the effect of moisture on the state of stability was not known.

For large flux levels, the fit of the Smith and Banke (1975) data alone suggests that the intercept is negligible and that $C_H=1.46\times10^{-3}$. This is shown graphically in Fig. 3, where the entire data set of Smith and Banke is plotted together with the data of Fig. 2. Most of the observations made on the detailed examination of the data in Fig. 2 appear insignificant compared to the wide-range unstable data of Smith and Banke, as the graph emphasizes the large $\bar{U}\Delta T$ values.

The standard deviation of the linear fit to all of the sensible heat flux data is 6.6×10^{-3} m K s^{-1} or 0.8 mW cm^{-2}, about 10 times the typical error of a direct flux measurement (Müller-Glewe and Hinzpeter, 1974). For the individual data subsets, stable, unstable, etc., the standard deviations from the linear fits are about 3.0×10^{-3} m K s^{-1}, except for the Smith and Banke (1975) data, which have a larger standard deviation of 14.8×10^{-3} m K s^{-1}.

b. Moisture flux

There are significantly fewer observations of directly measured moisture fluxes over the ocean than of the sensible heat flux, primarily due to the difficulty in measuring the fluctuating water vapor content of the air with sufficient spatial and temporal resolution. A total of 30 data points from four experiments were analyzed and are shown in Fig. 4. Similar to the sensible heat flux plots, the moisture flux data are plotted as $\overline{wq}$ versus $\bar{U}(\bar{Q}_s-\bar{Q}_a)=\bar{U}\Delta Q$ and the slope of a line on the plot is C_E. The data presented in Fig. 4 do not exhibit any anomalous features that were peculiar to the sensible heat flux data. In particu-

TABLE 2. Least-squares fit to $\overline{w\theta}=A+C_H\bar{U}\Delta T$.

Data set*	Number of data points	10^3A (m K s^{-1})	10^3C_H	$10^3\sigma$** (m K s^{-1})
All data	130	1.2(±0.62)†	1.41(±0.02)	6.6
All data except high wind speed data of Smith and Banke (1975)	116	2.3(±0.23)	0.91(±0.026)	2.4
Stable ($\bar{U}\Delta T<0$) conditions only	41	2.6(±1)	0.864(±0.14)	3.2
Unstable ($\bar{U}\Delta T>0$) conditions only	89	−1.8(±0.77)	1.462(±0.021)	6.3
Unstable conditions except Smith and Banke (1975) data	76	1.8(±0.38)	0.968(±0.039)	2.2
Smith and Banke (1975)	14	−0.4(±7.4)	1.46(±0.086)	14.8

* Except data for which "cold spikes" were observed.

** σ=standard deviation of fit of $A+C_H\bar{U}\Delta T$ to $\overline{w\theta}$.

† Values in parentheses are error estimates (standard deviations) of $A\times10^3$ and $C_H\times10^3$.

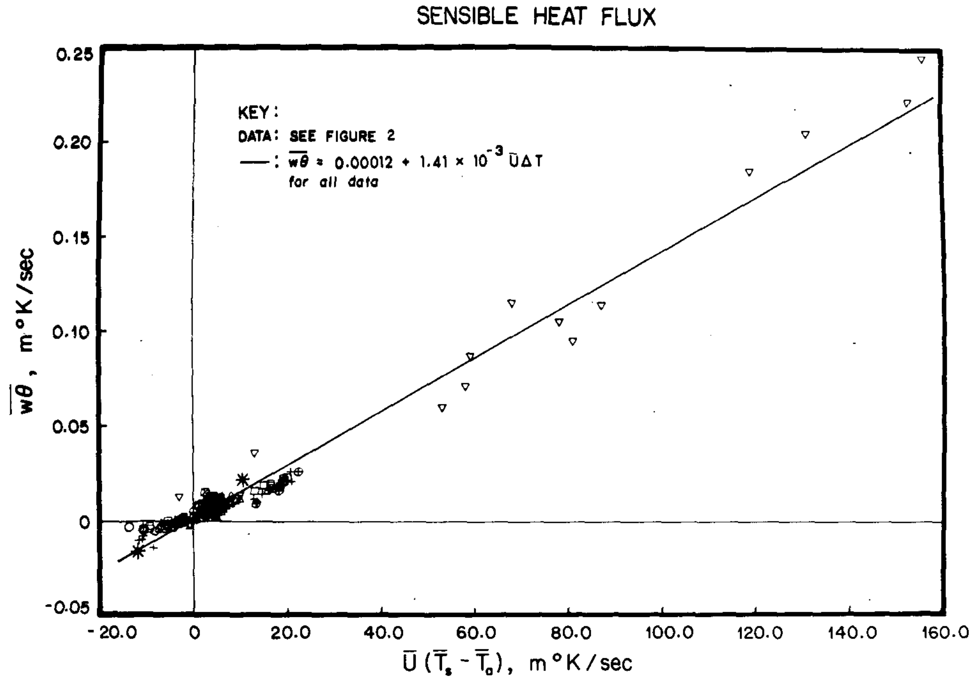

FIG. 3. As in Fig. 2 except additional data of Smith and Banke (1975).

lar the data of Pond *et al.* (1971) obtained off of San Diego appear to obey the same parameterization as the BOMEX data. Therefore, the moisture flux data appear to be well-described by the relation

$$\overline{wq}=B+C_E\bar{U}\Delta Q,$$

where

$$\left.\begin{aligned} B&=-2.6\times10^{-3}\ (\pm2.6\times10^{-3})\ (\text{m s}^{-1})\ (\text{g m}^{-3})\\ C_E&=1.32\times10^{-3}\ (\pm7.2\times10^{-5})\\ \sigma&=6.4\times10^{-3}\ (\text{m s}^{-1})\ (\text{g m}^{-3})\end{aligned}\right\}.$$

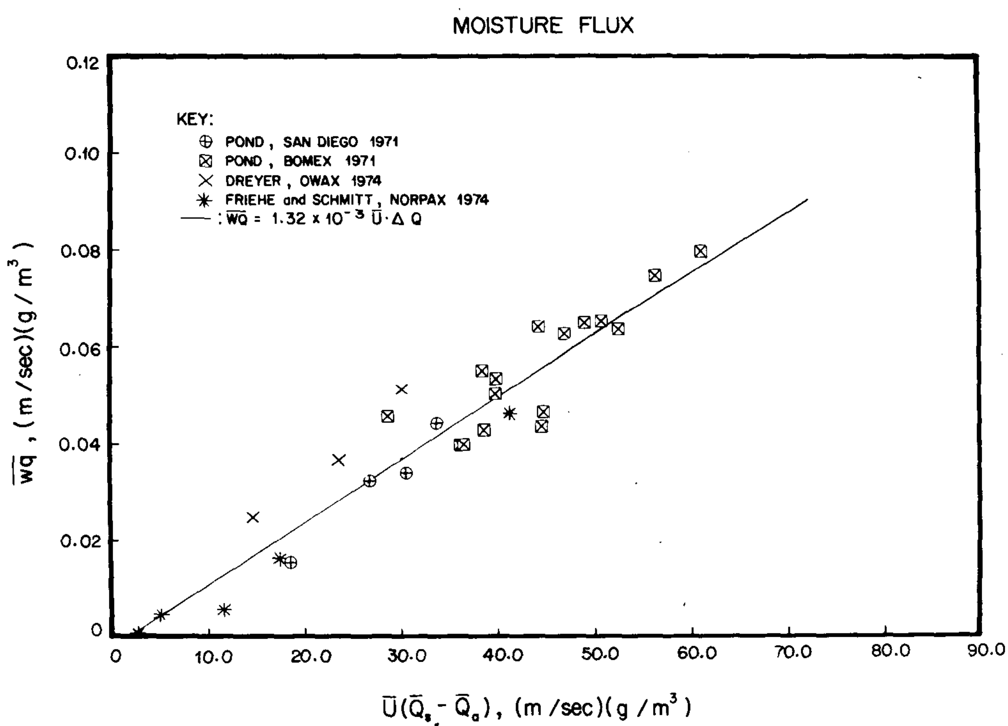

FIG. 4. Moisture flux. Directly-measured $\overline{wq}$ covariance versus $\bar{U}(\bar{Q}_s-\bar{Q}_a)$; slope of the line on the graph is $C_E=1.32\times10^{-3}$.

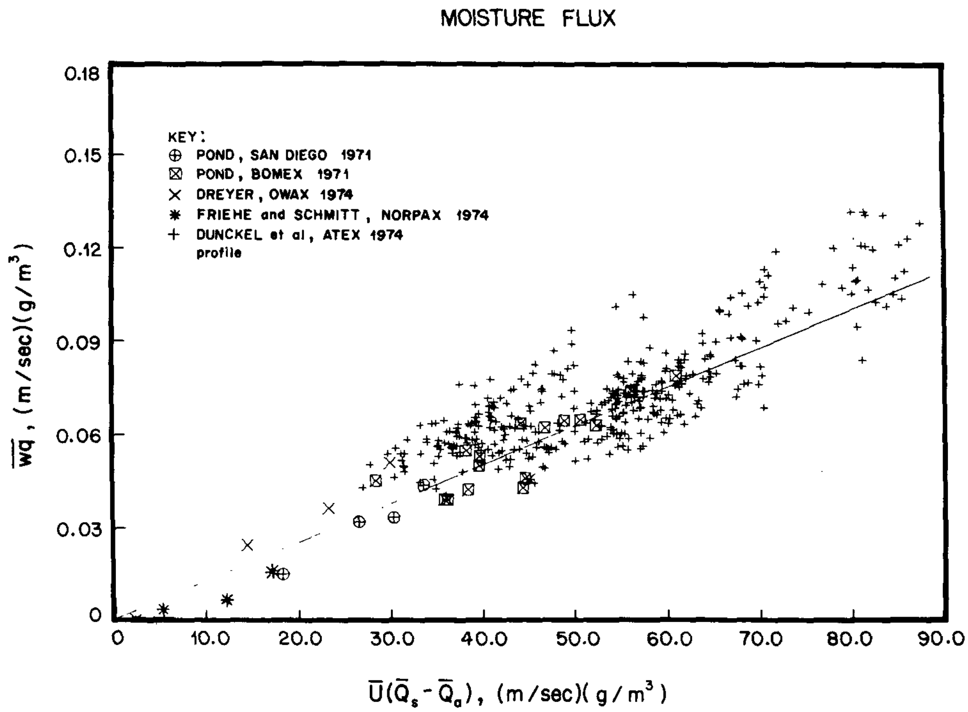

FIG. 5. As in Fig. 4 except with additional profile data of Dunckel *et al.* (1974); slope of the line is $C_E=1.32\times10^{-3}$.

The standard deviation σ of the linear fit to the moisture flux data of 6.4×10^{-3} (m s^{-1}) (g m^{-3}) is equivalent to a standard deviation of the corresponding latent heat flux of 1.5 mW cm^{-1}, about twice the standard deviation found for the sensible heat flux.

Dunckel *et al.* (1974) obtained an extensive set of moisture flux profile data in the Atlantic from a buoy. The moisture flux data are shown in Fig. 5, together with the data of Fig. 4. The least-squares fit of the Dunckel *et al.* data gave

$$\left.\begin{aligned} B&=0.014\ (\mathrm{m\ s^{-1}})\ (\mathrm{g\ m^{-3}})\\ C_E&=1.1\times10^{-3}\\ \sigma&=0.01\ (\mathrm{m\ s^{-1}})\ (\mathrm{g\ m^{-3}})\end{aligned}\right\}.$$

4. Discussion and conclusions

The compilation and analysis of sensible heat and moisture flux data from the various sources presented above shows, in general, good correlations with the mean parameters of wind speed and temperature and water vapor concentration differences. The parameterization of the moisture flux appears to be straightforward, and the error associated with the constant C_E should decrease as more measurements are added to the data set. The parameterization of the sensible heat flux is more complicated due to the observed effect of change of stability [in sign only; the effect of degree of stability (Deardorff, 1968) was not investigated] and indication that C_H increases with $\bar{U}\Delta T$ from the high wind speed data of Smith and Banke (1975). Therefore, for the sensible heat flux it is proposed that the bulk formula parameterization is best represented by the following conditional formulas:

1) For $0<\bar{U}\Delta T<25$

$$\overline{w\theta}=0.002+0.97\times10^{-3}\,\bar{U}\Delta T\ [\mathrm{m\ K\ s^{-1}}].$$

2) For large $\bar{U}\Delta T$ values under unstable conditions, i.e., $\bar{U}\Delta T>25$, the Smith and Banke relation is valid

$$\overline{w\theta}=1.46\times10^{-3}\,\bar{U}\Delta T\ [\mathrm{m\ K\ s^{-1}}].$$

3) For stable conditions, $\bar{U}\Delta T<0$

$$\overline{w\theta}=0.0026+0.86\times10^{-3}\,\bar{U}\Delta T\ [\mathrm{m\ K\ s^{-1}}].$$

If only approximate sensible heat flux values are required, the formula obtained by the fit to all of the available data can be used:

$$\overline{w\theta}=0.0012+1.41\times10^{-3}\,\bar{U}\Delta T\ [\mathrm{m\ K\ s^{-1}}].$$

In order to describe the stable and unstable regimes better, a second-order fit of all of the data was performed, but the decrease in the standard deviation was not significant. The data in Fig. 2 suggest that an appropriately weighted curve fit would describe the entire range $-20<\bar{U}\Delta T<160$ with a smaller standard deviation, but this was not attempted due to the lack of enough high wind speed data.

The agreement between the parameterizations of the directly-measured moisture flux data and the profile data of Dunckel *et al.* (1974) and that reported by Pond *et al.* (1974) is adequate. The formula obtained from the direct flux measurements is recommended:

$$\overline{wq}=1.32\times10^{-3}\,\bar{U}\Delta Q\ \ (\mathrm{m\ s^{-1}})\ (\mathrm{g/m^{-3}}),$$

where the intercept term is negligible. Unfortunately, there are no measurements of $\overline{wq}$ under high wind speed conditions to determine if C_E increases similarly to C_H.

The present analysis suggests that the coefficient C_E for moisture transfer is larger than that for heat transfer C_H under comparable wind speed conditions. As noted by Pond *et al.* (1974) and Coantic (1974), the bulk aerodynamic coefficients are in more general applications termed the Stanton number S for heat transfer and the Dalton number D for mass transfer. For turbulent flow over a smooth flat plate without buoyancy effects, semi-empirical turbulent boundary layer theory (Schlichting, 1960) gives

$$\mathrm{S}=C_H=0.03\ \mathrm{Pr}^{-\frac{2}{3}}\ \mathrm{R}^{-0.2},$$
$$\mathrm{D}=C_E=0.03\ \mathrm{Sc}^{-\frac{2}{3}}\ \mathrm{R}^{-0.2}.$$

Here Pr is Prandtl number, the ratio of the kinematic viscosity to thermal diffusivity=0.72 at 15°C for dry air; Sc the Schmidt number, the ratio of the kinematic viscosity to species diffusivity=0.58 at 15°C for water vapor in air; and R the Reynolds number, the characteristic velocity×characteristic length scale/kinematic viscosity.

For the boundary layer over the open ocean, the constant 0.03 and the dependence on the Reynolds number may not be the same as for the flat plate, but the Prandtl and Schmidt number dependences are likely to be the same. The equations and Prandtl and Schmidt number values give $C_E/C_H=1.16$, compared to $C_E/C_H=1.32/0.97=1.36$ from the data analysis. Therefore, the observation of $C_E>C_H$ may be partially explained by the non-negligible difference between the Prandtl and Schmidt numbers.

Dunckel *et al.* (1974) offered several hypotheses to explain the positive value of $\overline{w\theta}$ for $\Delta T=0$: 1) temperature fluctuations still exist with $\Delta T=0$, although the source and mechanism by which they are positively correlated with the w velocity fluctuations was not specified; 2) horizontal temperature differences can induce upward heat transfer, again by an unknown mechanism; and 3) buoyant motions caused by unstable water vapor will transfer heat upward if a positive correlation of temperature and humidity fluctuations (which is observed for $\overline{wq}>0$ and $\overline{w\theta}>0$) exists. Müller-Glewe and Hinzpeter (1974) indicated that their observed non-zero values of $\overline{w\theta}$ for $\Delta T=0$ were not due to measurement error, and further noted that the values increased with wind speed.

A concern with the use of the bulk formulas is that the actual surface temperature of the sea should be used for $\bar{T}_s$, rather than the average temperature at some shallow depth. Hasse (1971) developed a model to determine the temperature difference between the surface and upper portion of the mixed layer in terms of the net heat flux (sum of sensible, latent and radiative fluxes) across the air-sea interface. Differences of up to ~0.5 K were predicted. To our knowledge the water temperature data in the studies reported in Table 1 were obtained by the bucket or equivalent method below the sea surface, rather than by a radiometric surface thermometer. Therefore, some of the scatter in the data in Figs. 2–5 may be due to deviations of the actual surface temperature from the temperature at the sampled depth. Hopefully, the data are representative of a wide variety of radiative flux conditions, and the deviations are averaged-out in the final formulas presented. The finite positive values of $\overline{w\theta}$ for $\Delta T=0$ do not appear to be explained by the deviation of the actual surface temperature from the bucket temperature. For unstable conditions and assuming a temperature profile which decreases monotonically from the bucket temperature depth, through the interface and into the air, the actual sea surface temperature would be less than the bucket temperature. If actual surface temperatures were then used on a $\overline{w\theta}$ versus $\bar{U}\Delta T$ plot, the data would be shifted to the left and a higher intercept at $\Delta T=0$ would be implied. For the stable case, assuming a monotonic increase of temperature with height, the extrapolated intercept would decrease.

It should be emphasized that in the micrometeorological experiments reported in Table 1, accurate and precise mean measurements were generally made in conjunction with the turbulence measurements. In application of the above bulk formulas to other data sets, the lack of accuracy and precision in the measurements of the wind speed, temperature and moisture differences will be additional sources of error. Also, the bulk aerodynamic formulas presented here are based on fluxes averaged over relatively short periods of time—of the order of an hour. It is not clear that the same formulas apply for significantly larger averaging times, such as used in climatological studies.

More data of surface fluxes at high wind speeds are required under open ocean conditions. For the moisture flux data presented in Fig. 4, the largest wind speed was only 8 m s⁻¹. The heat flux data of Smith and Banke (1975) were obtained at wind speeds of up to 21 m s⁻¹, but the measurements were made from an island and there may be subtle fetch and wave-modification effects.

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