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PROFILE-BULK METHOD FORMULAS FOR CALCULATING FLUX AND  
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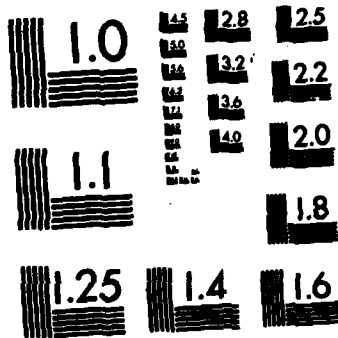
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**Profile-Bulk Method Formulas  
for Calculating Flux and Stability  
in the Marine Atmospheric Surface Layer  
and a Survey of Field Experiments**

THEODORE V. BLANC

*Atmospheric Physics Branch  
Environmental Sciences Division*

December 30, 1982



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**CONTENTS**

**ABSTRACT** ..... 1

**INTRODUCTION** ..... 1

**SURVEY OF MARINE SURFACE LAYER PROFILE FLUX DATA** ..... 1

**PROFILE FLUX AND STABILITY CALCULATIONS** ..... 1

**BULK FLUX AND STABILITY CALCULATIONS** ..... 8

**ACKNOWLEDGMENTS** ..... 11

**BIBLIOGRAPHY** ..... 11



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# **PROFILE-BULK METHOD FORMULAS FOR CALCULATING FLUX AND STABILITY IN THE MARINE ATMOSPHERIC SURFACE LAYER AND A SURVEY OF FIELD EXPERIMENTS**

## **Abstract**

A comprehensive survey of profile flux experiments conducted in the marine atmospheric surface layer over the last 25 years is presented with an extensive bibliography and the formulas for computing fluxes and stability.

## **INTRODUCTION**

A survey of marine profile flux experiments has not appeared in the literature since Roll (1965). An updated catalog has been prepared with an extensive bibliography of field experiments, methods, and instrumentation. The constituent equations used to compute the profile and bulk fluxes are distributed over a variety of sources, which contain different notations and approximations, and several have typographical errors. A complete and accurate set of formulas is presented for the first-time user.

## **SURVEY OF MARINE SURFACE LAYER PROFILE FLUX DATA**

One of the earliest experiments to be found in the literature is that of Wüst (1920), which reported wind profile measurements taken from a ship in the Baltic Sea. Perhaps the earliest observations, as noted by Barenblatt et al. (1975), were those conducted by G. I. Taylor in 1913 using instruments sent aloft with a kite from the stern of a whaling ship under sail off the coast of Newfoundland. A list of marine surface layer profile measurements reported up to 1962 may be found in Tables X and XXIV of Roll (1965). A current survey of marine surface layer wind profile observations accompanied by temperature and/or humidity profile measurements is presented in Table 1.

As indicated in the table, there are two basic observational strategies for acquiring such profile data. The first is to employ a number of identical sensors permanently situated at various altitudes, which acquire data on a continuous basis. The second is to employ a single sensor, which is sequentially moved from one altitude to another, acquiring data while stationary at the desired altitude.

Our literature searches revealed the existence of only about 2,100 h of offshore ocean profile measurements in which all three of the primary fluxes had been measured; 85% of the data had been obtained in the equatorial region of the Atlantic Ocean. It was further determined that only 1% of that data base had been acquired under stable atmospheric conditions and that little data existed for wind speeds in excess of 12 m/s.

## **PROFILE FLUX AND STABILITY CALCULATIONS**

Encompassed within the profile method are several semiempirically derived profile-flux relationships which have been summarized by Yaglom (1977). The two principal competing relationship schemes are those proposed by Dyer and Hicks (1970) and by Businger et al. (1971). Dyer (1974) has reviewed the differences the two schemes. Lo and McBean (1978) found that the two schemes could

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Table 1 — Marine Surface Layer Wind Speed Profile Observations Accompanied by Temperature and/or Humidity Profile Measurements Reported Over the Last 25 Years

Source	Experiment (name or location)	Type and Number of Profile Measurement Levels			Altitude Cross-Section Above Mean Sea Level (m)	Observational Strategy (M = Multiple sensors fixed at measurement levels; S = Single sensor sequentially moved to measurement levels)	Environmental Regime	Measurement Platform	Averaging Period (min)	Amount of Data (hours)	Approximate Range of Wind Speeds at 10 m (m/s)
		Wind Speed	Temperature	Humidity							
Deacon et al. (1956)	Port Phillip Bay	6	2	—	1.5–13	M	Coastal & Inland Water	Ship	30	49	1–14
Takahashi (1958)	Kagoshima Bay	5	5	5	0.3–4	M&S	Inland Water	Boat & Offshore Pole	10	67	1–7
Fleagle et al. (1958)	East Sound	8	4	4	0.4–4	M&S	Inland Water	Raft	60	30	3–9
Bruce et al. (1961)	Lake Erie	6	6	—	0.6–13	S	Coastal	Ship	30	20	3–6
Deacon (1962)	Port Phillip Bay	3	2	—	3.0–13	M	Coastal & Inland Water	Ship	30	93(?)	4–15
Bogorodskiy (1964)	?	2	2	—	1.0–10	M	?	Boat	10	13	2–7
Hoerber (1969)	Equatorial Atlantic	4	4	4	1.6–9	M	Open Ocean	Buoy	10	288	4–10
Miyake et al. (1970a)	Spanish Banks	5	5	5	0.6–5	S	Coastal	Offshore Mast	45	6	4–9
Badgley et al. (1972)	Indian Ocean	6	6	6	1.6–8	S	Open Ocean & Coastal	Buoy	40	79	2–8
Garratt (1972)	Lough Neagh	5	5	5	1.0–16	M	Inland Water	Offshore Tower	10	<1	7
Paulson et al. (1972)	BOMEX <sup>a</sup>	4	4	4	2.0–11	S	Open Ocean	Stabilized Ship	48	113	2–8
Hsu (1973)	Fort Walton Beach	6	3	—	0.2–6	M	Beach	Onshore Towers	15	13	3–6
Donelan et al. (1974)	Lake Ontario	5	5	3&5	2.4–12	M&S	Open Lake	Offshore Tower	30 <sup>d</sup>	132	1–17
Dunckel et al. (1974)	ATEX <sup>b</sup>	7	4	4	1.2–8	M	Open Ocean	Buoy	10	124	4–11
Hasse et al. (1975)	GATE <sup>c</sup>	9	5	5	0.5–8	M	Open Ocean	Buoy	10	6	2–7
Peterson (1975)	Risø	7	8	—	7.0–123	M	Beach & Inland Coastal	Onshore Tower	?	?	6–9
Hupfer et al. (1976)	Baltic Sea	5	5	—	1.5–7	S	Beach	Onshore Pole	30	4	—7
Krügermeyer (1976)	ATEX <sup>b</sup>	7	4	4	1.2–8	M	Open Ocean	Buoy	10	124	4–11
Vuys & Businger (1977)	Schiermonnikoog Island	5	5	—	0.5–13	M	Beach	Onshore Towers	60	45	4–8
Hasse et al. (1978b)	GATE <sup>c</sup>	7	5	5	1.1–8	M	Open Ocean	Buoy	10	1,419	1–12
Thomson (1979)	Risø	4–8	2–3	—	1.0–27	M	Beach & Inland Coastal	On- & Off-shore Towers	10	32	2–7
Blanc (1981)	San Nicolas Island	2	2	2	9.2–18	M	Coastal	Onshore Tower	30	136	2–17

<sup>a</sup>BOMEX = Barbados Oceanographic and Meteorological Experiment, 1969.<sup>b</sup>ATEX = Atlantic Trade Wind Experiment, 1969.<sup>c</sup>GATE = Global Atmospheric Research Program Atlantic Tropical Experiment, 1974.<sup>d</sup>30-min running average updated ever 10 min.

yield differences as large as 40% in the estimated fluxes; they suggested that the discrepancies could be resolved by setting the von Kármán constant equal to 0.40, instead of the recommended 0.35, when using the Businger et al. scheme. This, in effect, renders the Businger et al. scheme equivalent to that of Dyer and Hicks. A justification for this approach was subsequently provided by Wieringa (1980). For additional information see Wyngaard et al. (1982), Wieringa (1982), and Gill (1982). An international turbulence measurement comparison conducted over land and reported by Francey and Garratt (1981) proposed a von Kármán constant of 0.38 ( $\pm 0.04$ ). An alternative analysis of the same comparison presented by Dyer and Bradley (1982) suggested a constant of 0.40. A rationale for the minimum height of the lowest measurement level and the separation between adjacent levels is presented in Blanc (1983a).

At altitudes  $z = 6, 12,$  and  $24$  m, for example, 30-min-averaged measurements are made of the air temperature ( $\bar{T}_n$ ) in  $^{\circ}\text{C}$ , the dew-point temperature ( $\bar{T}_{dn}$ ) in  $^{\circ}\text{C}$ , and the wind speed ( $\bar{u}_n$ ) in m/s. The 30-min-averaged barometric pressure ( $\bar{P}$ ) in pascals is measured at the middle altitude and is used for all levels. For simplicity, the overbar notation used to indicate time-averaged measurements will be implied but not indicated. Thus, at  $z = 24$  m ( $z_{24}$ ):  $T_{24}, T_{d24}, u_{24}$ ; at  $z = 12$  m ( $z_{12}$ ):  $T_{12}, T_{d12}, u_{12}, P$ , and at  $z = 6$  m ( $z_6$ ):  $T_6, T_{d6}, u_6$ . If the dew-point temperature is not measured at each altitude, then the wet-bulb temperature or relative humidity should be measured instead.

The saturated vapor pressure ( $e_{sn}$ ) in pascals at each altitude ( $z_n$ ) is calculated by

$$e_{sn}(P, T) = Pa_1^{b_3} \times 10^{a_2 b_4 + a_4 b_5 + a_5 b_6},$$

where

$$a_1 = \frac{373.16}{T_n + 273.16},$$

$$a_2 = a_1 - 1,$$

$$a_3 = 1 - \frac{1}{a_1},$$

$$a_4 = (10^{a_2 b_1}) - 1,$$

$$a_5 = (10^{a_3 b_2}) - 1,$$

the Goff-Gratch humidity formulation constants are

$$b_1 = -3.49149, \quad b_4 = -7.90298,$$

$$b_2 = 11.344, \quad b_5 = 8.1328 \times 10^{-3},$$

$$b_3 = 5.02808, \quad b_6 = -1.3816 \times 10^{-7};$$

and  $n = 6, 12,$  and  $24$ .

The water vapor pressure ( $e_n$ ) in pascals at each altitude ( $z_n$ ) is calculated by

$$e_n(P, T_d) = Pa_1^{b_3} \times 10^{a_2 b_4 + a_4 b_5 + a_5 b_6},$$

where

$$a_1 = \frac{373.16}{T_{dn} + 273.16},$$



$a_2$  through  $a_5$  are calculated in the same manner as above, and  $n = 6, 12, \text{ and } 24$ . In the event that the wet-bulb temperature ( $T_{wb}$ ) in  $^{\circ}\text{C}$  or the relative humidity ( $\text{RH}_n$ ) in % is measured at each altitude instead of the dew-point temperature ( $T_{dn}$ ), then the water vapor pressure can be calculated by

$$e_n(P, T, e_s, T_{wb}) = e_{sn} - 6.6 \times 10^{-4} [1 + (1.15 \times 10^{-3} T_{wb})] P (T_n - T_{wb})$$

or

$$e_n(e_s, \text{RH}) = \frac{e_{sn} \text{RH}_n}{100}$$

The specific humidity ( $q_n$ ) in kg/kg at each altitude ( $z_n$ ) is calculated by

$$q_n(P, e) = \frac{0.622 e_n}{P - (0.378 e_n)}$$

The absolute humidity or water-vapor density ( $\rho_{v,n}$ ) in  $\text{kg/m}^3$  at each altitude ( $z_n$ ) is calculated by

$$\rho_{v,n}(T, e) = e_n \left[ \frac{1}{R_v (T_n + 273.16)} \right]$$

where  $R_v$  is the water-vapor gas constant,  $4.615 \times 10^2 \text{ J/(kg K)}$ . The mixing ratio ( $r_n$ ) in kg/kg at each altitude ( $z_n$ ) is calculated by

$$r_n = \frac{0.622 e_n}{P - e_n}$$

The relative humidity ( $\text{RH}_n$ ) in percent at each altitude ( $z_n$ ) is calculated by

$$\text{RH}_n(e, e_s) = \left( \frac{e_n}{e_{sn}} \right) \times 100$$

In each case  $n = 6, 12, \text{ and } 24$ .

The virtual temperature ( $T_{v,n}$ ) in  $^{\circ}\text{C}$  at each altitude ( $z_n$ ) is calculated by

$$T_{v,n}(T, q) = \{(T_n + 273.16) \times [1 + (0.608 q_n)]\} - 273.16$$

The potential temperature ( $\theta_n$ ) in  $^{\circ}\text{C}$  at each altitude ( $z_n$ ) is calculated by

$$\theta_n(T, z) = T_n + (0.0098 z_n)$$

The virtual potential temperature ( $\theta_{v,n}$ ) in  $^{\circ}\text{C}$  at each altitude ( $z_n$ ) is calculated by

$$\theta_{v,n}(T_v, z) = T_{v,n} + (0.0098 z_n)$$

Again,  $n = 6, 12, \text{ and } 24$ .

The geometric mean height (gmh) is the geometric middle altitude when vertical distance is represented on a log scale. Because the three measurement altitudes ( $z_6, z_{12}, z_{24}$ ) were selected such that  $(\ln z_{24} - \ln z_{12}) = (\ln z_{12} - \ln z_6)$ ,  $z_{12}$  is, also, the gmh:

$$\text{gmh} = (z_6 z_{24})^{1/2} = z_{12}$$

Since profiles tend to vary approximately logarithmically, the gmh of the profile measurement levels is the altitude for which the profile measurements are typically most valid.

The Richardson number stability ( $\text{Ri}_{\text{gmh}}$ ), a dimensionless quantity, valid for the geometric mean height of the component measurement altitudes is calculated by

$$Ri_{gmh} = Ri_{12} = \frac{g \frac{\partial \theta_v}{\partial z}}{(T_{v12} + 273.16) \times \left( \frac{\partial u}{\partial z} \right)^2},$$

where

$$\frac{\partial \theta_v}{\partial z} \approx \frac{\theta_{v24} - \theta_{v6}}{gmh \ln(z_{24}/z_6)},$$

$$\frac{\partial u}{\partial z} \approx \frac{u_{24} - u_6}{gmh \ln(z_{24}/z_6)},$$

$$gmh = z_{12},$$

and  $g$  is the acceleration constant due to gravity,  $9.8 \text{ m/s}^2$ . The approximations for the partial derivatives are accurate to about  $\pm 3\%$ .

If  $Ri_{12} < 0$ , the atmosphere is unstable and the Monin-Obukhov stability ( $\zeta_{12}$ ) at  $z = 12 \text{ m}$  ( $z_{12}$ ) is calculated by

$$\zeta_{12}(-0.01 < Ri < 0) \approx 1.3 Ri_{12},$$

$$\zeta_{12}(-1.5 \leq Ri \leq -0.01) \approx -10^{c_1 + c_2 + c_3},$$

where

$$c_1 = 2.8443 \times 10^{-2},$$

$$c_2 = 0.96125[\log(-Ri_{12})],$$

$$c_3 = 1.3655 \times 10^{-3}[\log(-Ri_{12})]^2,$$

or

$$\zeta_{12}(-\infty < Ri < -1.5) \approx 1.05 Ri_{12}.$$

The middle-range  $\zeta_{12}$  approximation is a least-squares polynomial fit to the reverse solution of the Businger et al. (1971) equation for  $Ri$  ( $\zeta < 0$ ). All three of the approximations are accurate to about  $\pm 2\%$ . The quantity  $\zeta$  is of questionable validity for  $Ri < -2$ .

The Monin-Obukhov length ( $L$ ) in meters is then calculated by

$$L = \frac{z_{12}}{\zeta_{12}}.$$

Once  $L$  is calculated for one altitude, the Monin-Obukhov stability ( $\zeta_n$ ) for any altitude ( $z_n$ ) can be computed by

$$\zeta_n(z) \equiv \frac{z_n}{L}.$$

When  $Ri_{12} < 0$ , then  $\zeta_{12} < 0$ . Under near-neutral conditions ( $\zeta_{12} \approx 0$ ) the various profiles exhibit a linear form when plotted on a semilogarithmic plot. Under non-neutral conditions the profiles need to be corrected for stability-dependent curvature (see Fig. 1) to properly compute the fluxes. The stability- and altitude-dependent profile corrections for potential temperature ( $\psi_\theta$ ), for specific humidity ( $\psi_q$ ), and for wind speed ( $\psi_u$ ) are calculated by

$$\psi_{\theta_n}(\zeta < 0) = \psi_{\theta_n}(\zeta < 0) - \ln \left[ \frac{1 + y_n}{2} \right]$$

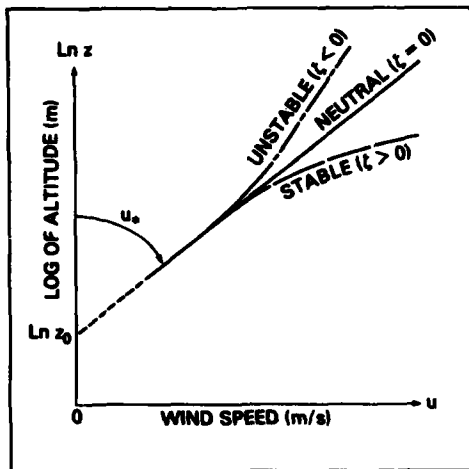


Fig. 1 — Typical stability-dependent curvatures in a semilogarithmic profile representation produced by non-neutral conditions. The scaling parameter, in this case the friction velocity ( $u_w$ ), is a convenient shorthand for expressing the profile and is inversely proportional to the stability-corrected profile slope. The roughness length ( $z_0$ ) is defined as the altitude at which the wind speed would be zero if the stability-corrected wind profile were projected downward.

and

$$\psi_{un}(\zeta < 0) = 2 \ln \left[ \frac{1 + x_n}{2} \right] + \ln \left[ \frac{1 + (x_n)^2}{2} \right] - [2 \arctan x_n] + \frac{\pi}{2},$$

where

$$x_n = (1 - 15\zeta_n)^{1/4},$$

$$y_n = (1 - 9\zeta_n)^{1/2},$$

$\arctan x_n$  is expressed in radians, and

$$n = 6, 12, \text{ and } 24.$$

If  $Ri_{12} = 0$ , the stability is said to be neutral,  $\zeta_{12} = 0$ , and

$$\psi_{\theta n}(\zeta = 0) = \psi_{qn}(\zeta = 0) = \psi_{un}(\zeta = 0) = 0.$$

If  $Ri_{12} > 0$ , the atmosphere is stable and the Monin-Obukhov stability ( $\zeta_{12}$ ) at  $z = 12$  m ( $z_{12}$ ) is calculated by

$$\zeta_{12}(0.2 \geq Ri \geq 0) = \frac{-d_2 - (d_2^2 - 4d_1d_3)^{1/2}}{2d_1},$$

where  $d_1$  through  $d_3$  are the components of a quadratic reverse solution of the Businger et al. (1971) equation for  $Ri(\zeta > 0)$ , in which

$$d_1 = (22.09 Ri_{12}) - 4.7,$$

$$d_2 = (9.4 Ri_{12}) - 0.74,$$

and

$$d_3 = Ri_{12}.$$

The quantity  $\zeta$  is undefined for  $Ri > \sim 0.2$ . The Monin-Obukhov length ( $L$ ) in meters is calculated, as before, by

$$L = \frac{z_{12}}{\zeta_{12}}$$

Again, once  $L$  is calculated for one altitude, the Monin-Obukhov stability ( $\zeta_n$ ) for any altitude ( $z_n$ ) can be computed by

$$\zeta_n(z) \equiv \frac{z_n}{L}$$

When  $Ri_{12} > 0$ , then  $\zeta_{12} > 0$ . The stability- and altitude-dependent profile correction for potential temperature ( $\psi_\theta$ ), for specific humidity ( $\psi_q$ ), and for wind speed ( $\psi_u$ ) are calculated by

$$\psi_{\theta n}(\zeta > 0) = \psi_{qn}(\zeta > 0) = \frac{-4.7\zeta_n}{0.74}$$

and

$$\psi_{un}(\zeta > 0) = -4.7\zeta_n,$$

where

$$n = 6, 12, \text{ and } 24.$$

The mean stability-corrected slope of the potential temperature profile  $\langle S_\theta \rangle$  is calculated by

$$\langle S_\theta \rangle = \frac{S_{\theta(6,12)} + S_{\theta(12,24)} + 2 S_{\theta(6,24)}}{4},$$

where

$$S_{\theta(n,m)} = \frac{(\ln z_m - \psi_{\theta m}) - (\ln z_n - \psi_{\theta n})}{\theta_m - \theta_n}$$

and  $n, m = 6, 12; 12, 24; \text{ and } 6, 24$ . The mean stability-corrected slope of the specific-humidity profile  $\langle S_q \rangle$  and the mean stability-corrected slope of the wind-speed profile  $\langle S_u \rangle$  are calculated by inserting, respectively,  $q$  for all  $\theta$ 's and  $u$  for all  $\theta$ 's in the above two equations. The profile slopes are utilized here, instead of the more conventional profile gradients, to eliminate the confusion that exists in the literature between the contradictory Western and USSR definitions of gradient and lapse rate.

The scaling potential temperature ( $\theta_*$ ) in °C, the scaling specific humidity ( $q_*$ ) in kg/kg, and scaling wind speed ( $u_*$ ) in m/s (see Fig. 1) are calculated by

$$\theta_* = \frac{k}{0.74 \langle S_\theta \rangle},$$

$$q_* = \frac{k}{0.74 \langle S_q \rangle},$$

and

$$u_* = \frac{k}{\langle S_u \rangle},$$

where  $k$  is the von Kármán constant of 0.40. The scaling wind speed parameter ( $u_*$ ) is more commonly called the friction velocity. The roughness length ( $z_0$ ) in meters is computed from the wind-speed profile by

$$z_0 = e^{\frac{(\ln z_{12} - \psi_{\theta 12}) - k z_{12} / u_*}{k}}$$

The density of moist air ( $\rho$ ) in  $\text{kg/m}^3$  for all altitudes is computed by

$$\rho(P, T_v) = \frac{(3.4838 \times 10^{-3})P}{T_{v12} + 273.16}$$

The specific heat of moist air at constant pressure ( $C_p$ ) in  $\text{J}/(\text{kg K})$  for all altitudes is computed by

$$C_p(q) \approx 1.004[1 + 0.9(q_{12})] \times 10^3.$$

The latent heat of water vaporization ( $L_v$ ) in  $\text{J}/\text{kg}$  for all altitudes is computed by

$$L_v(T) \approx 4.1868(597.31 - 0.56525T_{12}) \times 10^3.$$

Employing a negative flux value to indicate a downward direction and a positive flux value to indicate an upward direction, the sensible heat flux ( $H_s$ ) in  $\text{W}/\text{m}^2$  is computed by

$$H_s = -\rho C_p \theta_* u_*.$$

The water-vapor flux ( $E$ ) in  $\text{kg}/(\text{s m}^2)$  is computed by

$$E = -\rho q_* u_*.$$

Frequently the moisture flux is presented in terms of the latent heat flux ( $H_L$ ) in  $\text{W}/\text{m}^2$ ,

$$H_L = E L_v.$$

The momentum flux ( $M$ ) in  $\text{N}/\text{m}^2$  is computed by

$$M = -\rho u_*^2.$$

Frequently the mechanical flux is represented by the stress ( $\tau$ ) in  $\text{N}/\text{m}^2$ ,

$$\tau \equiv -M.$$

## BULK FLUX AND STABILITY CALCULATIONS

Encompassed within the bulk method are more than 20 different semiempirical bulk-flux relationship schemes. Partial summaries may be found in Pond et al. (1974), in Friehe and Schmitt (1976), and in Kondo (1977). The various types of proposed schemes run the complete gamut in terms of complexity and sophistication. For example, where Friehe and Schmitt use single statistical conglomerate coefficients, Krügermeyer (1976) employs stability-dependent coefficients, and Liu et al. (1979) utilize a computational iteration technique requiring five input parameters for twelve dependent variables in five equations which reduce to three simultaneous equations of three unknowns. At the present time there is no single universally accepted bulk-flux scheme. Depending upon which scheme is used, the same input data can yield estimated fluxes differing by as much as 100%.

Admittedly, because of the wide variety from which to choose, the selection of a bulk scheme to accompany the profile method is of necessity somewhat arbitrary. Here I employ the sensible heat flux and latent heat flux schemes proposed by Friehe and Schmitt (1976), as used in Friehe and Pazan (1978), and the momentum flux scheme proposed by Smith (1980a). I have selected these schemes

because of their simplicity and statistically broad data bases. Other types and forms of profile-bulk combinations have been proposed by Okamoto et al. (1968), Fujita (1978), Itier (1980), and Wang (1981).

In addition to those measurements made for the profile calculations, the bulk calculations require a water-temperature measurement ( $T_w$ ) in °C, made at or just below the sea surface. The potential temperature ( $\theta_{10}$ ) in °C, the absolute humidity or water-vapor density ( $\rho_{v10}$ ) in kg/m<sup>3</sup>, and the wind speed ( $u_{10}$ ) in m/s at the standard bulk measurement altitude of  $z = 10$  m ( $z_{10}$ ) can be computed from the profile measurements at  $z = 6$  m ( $z_6$ ) and  $z = 12$  m ( $z_{12}$ ) by

$$\begin{aligned}\theta_{10} &\approx \theta_{12} - 0.263(\theta_{12} - \theta_6), \\ \rho_{v10} &\approx \rho_{v12} - 0.263(\rho_{v12} - \rho_{v6}),\end{aligned}$$

and

$$u_{10} \approx u_{12} - 0.263(u_{12} - u_6),$$

where

$$\frac{\ln z_{12} - \ln z_{10}}{\ln z_{12} - \ln z_6} = 0.263.$$

The mean vertical velocity ( $\omega$ )-temperature ( $\theta$ ) covariance  $\langle \omega'\theta' \rangle$  in (m K)/s is computed by

$$\langle \omega'\theta' \rangle \approx 0.002 + [C_H u_{10} (T_w - \theta_{10})],$$

where  $C_H$  is the bulk sensible heat transfer coefficient at the 10-m altitude,  $0.92 \times 10^{-3}$ .

The sensible heat flux ( $H_s$ ) in W/m<sup>2</sup> is computed by

$$H_s = \rho C_p \langle \omega'\theta' \rangle,$$

where the density of moist air ( $\rho$ ), the specific heat of moist air at constant pressure ( $C_p$ ), and the flux direction sign convention are the same as used in the profile calculations.

The water-vapor pressure near the water surface ( $e_w$ ) in pascals is calculated by

$$e_w(P, T_d) = P a_1^{b_3} \times 10^{a_2 b_4 + a_4 b_5 + a_5 b_6},$$

where

$$a_1 = \frac{373.15}{T_{dw} + 273.16},$$

the barometric pressure ( $P$ ) is in pascals,  $a_2$  through  $a_5$  and  $b_1$  through  $b_6$  are the same as for the profile calculations of vapor pressure, and the dew point near the water surface ( $T_{dw}$ ) in °C is estimated by

$$T_{dw} \approx T_w,$$

assuming the relative humidity at or near the sea surface is 100%. The absolute humidity, or water-vapor density, at the sea surface ( $\rho_{vw}$ ) in kg/m<sup>3</sup> is then calculated by

$$\rho_{vw}(T, e) = e_w \left[ \frac{1}{R_v (T_w + 273.16)} \right],$$

where  $R_v$  is the water-vapor gas constant,  $4.615 \times 10^2$  J/(kg K).

The mean vertical velocity ( $\omega$ )-absolute humidity ( $\rho_v$ ) covariance  $\langle \omega' \rho_v' \rangle$  in  $(\text{m kg})/(\text{s m}^3)$  is calculated by

$$\langle \omega' \rho_v' \rangle \approx C_E u_{10} (\rho_{vw} - \rho_{v10}),$$

where  $C_E$  is the bulk water-vapor transfer coefficient at the 10 m-altitude,  $1.32 \times 10^{-3}$ . The water-vapor flux ( $E$ ) in  $\text{kg}/(\text{s m}^2)$  is computed by

$$E = \langle \omega' \rho_v' \rangle.$$

As before, the latent heat flux ( $H_L$ ) in  $\text{W}/\text{m}^2$  is computed by

$$H_L = E L_v,$$

where  $L_v$  is the same latent heat of water vaporization employed in the profile calculations.

The momentum flux ( $M$ ) in  $\text{N}/\text{m}^2$  is computed by

$$M = -\rho C_D u_{10}^2,$$

where  $C_D$  is the bulk aerodynamic drag coefficient at the 10-m altitude,

$$C_D = (0.61 + 0.063 u_{10}) \times 10^{-3}.$$

As before, the stress ( $\tau$ ) in  $\text{N}/\text{m}^2$  is computed by

$$\tau \equiv -M.$$

The scaling potential temperature ( $\theta_*$ ) in  $^\circ\text{C}$ , the scaling specific humidity ( $q_*$ ) in  $\text{kg}/\text{kg}$ , and the scaling wind speed or friction velocity ( $u_*$ ) in  $\text{m}/\text{s}$  are calculated by

$$\theta_* = \frac{-H_s}{\rho u_* C_p},$$

$$q_* = \frac{-E}{\rho u_*},$$

and

$$u_* = \left( \frac{\tau}{\rho} \right)^{1/2}.$$

The Monin-Obukhov length ( $L$ ) in meters is calculated by

$$L = \frac{-(\theta_{10} + 273.16) u_*^3}{gk \langle \omega' \theta' \rangle},$$

where the acceleration due to gravity ( $g$ ),  $9.8 \text{ m}/\text{s}^2$ , and the von Kármán constant ( $k$ ),  $0.40$ , are the same as used in the profile calculations.

The Monin-Obukhov stability ( $\zeta_{12}$ ) for the geometric mean height of the profile measurements at  $z = 12 \text{ m}$  ( $gmh = z_{12}$ ) is computed by

$$\zeta_{12}(z) = \frac{z_{12}}{L}.$$

If  $\zeta_{12} < 0$ , the unstable condition, the Richardson number stability ( $Ri_{12}$ ) is computed by

$$Ri_{12}(\zeta < 0) = 0.74 \zeta_{12} \frac{(1 - 15\zeta_{12})^{1/2}}{(1 - 9\zeta_{12})^{1/2}}.$$

If  $\zeta_{12} = 0$ , the neutral condition, the Richardson number stability ( $Ri_{12}$ ) is also zero.

If  $\zeta_{12} > 0$ , the stable condition, the Richardson number stability ( $Ri_{12}$ ) is computed by

$$Ri_{12}(\zeta > 0) = \frac{(0.74 + 4.7\zeta_{12})\zeta_{12}}{(1 + 4.7\zeta_{12})^2}.$$

The roughness length ( $z_0$ ) in meters is computed from  $\psi_{u12}$  (via  $\zeta_{12}$ ) and  $u_*$  in the same manner as for the profile method.

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