EDDY FLUX MEASUREMENTS OVER THE OCEAN AND RELATED TRANSFER COEFFICIENTS

R. J. FRANCEY and J. R. GARRATT

CSIRO Division of Atmospheric Physics, Aspendale, Victoria, Australia

(Received 22 June, 1977)

Abstract. Eddy correlation measurements of vertical turbulent fluxes made during AMTEX 1975 are used to assess the reliability of flux prediction from established bulk transfer relations, using both surface-layer and planetary boundary-layer formulations. The surface-layer formulae predict momentum and latent heat fluxes to an accuracy comparable to the direct eddy correlation method, using transfer coefficients of C_{DN} (at 10 m and in neutral conditions) increasing with wind speed, and a constant $C_{EN} \simeq 1.5 \times 10^{-3}$. The data suggest C_{HN}, for sensible heat, increases significantly with wind speed and is on average 30% lower than C_{EN}. The boundary-layer drag coefficient, C_{GD}, agrees within about 40% of recently published values using a vertically averaged geostrophic wind to the height of the lowest temperature inversion, corrected for trajectory curvature. Values of $\theta_*/\Delta\theta$ from which C_{GH} is derived, are in excellent agreement if the published values are modified to account for inappropriate surface temperatures used in their derivation. Preliminary values of C_{GE} are also presented.

1. Introduction

Numerical models of the atmosphere often require large-scale estimates of surface fluxes consistent with the grid scale used, typically of the order of 100 km (e.g., Gadd and Keers, 1970) to 500 km (e.g., Delsol *et al.*, 1971). These estimates are generally based upon bulk transfer relations using both surface-layer and planetary boundary-layer formulations, which derive from similarity theories of a steady-state, horizontally homogeneous lower atmosphere. The associated transfer coefficients must be established empirically from direct measurements of the fluxes. The validity of the transfer relations on the grid scale then depends on the accuracy of such field measurements and the degree of horizontal inhomogeneity and time dependency of the velocity (u), temperature (θ) and humidity (q) fields. It has been normal practice to determine transfer coefficients from flux measurements (complex and susceptible to local influence) at a single location in the real atmosphere in which inhomogeneities and time variability are usually present.

During AMTEX 1975 (14 February–1 March) we obtained independent sets of direct measurements of momentum, sensible heat and water vapour fluxes over the ocean. This period was one of intense air mass modification and hence considerable horizontal inhomogeneity and baroclinity. However, the time scale of 'large-scale' variations was generally much greater than a few hours, which is typically the time scale of vertical mixing throughout the whole boundary layer in the unstable conditions of the AMTEX period. In addition, diurnal variations were small, in marked contrast to those experienced in an unstable atmosphere over land. Thus, apart from isolated occasions of rapid change in the large-scale u, θ fields due to frontal passage, conditions approximated to the steady-state requirement upon which the transfer relations are based. Further, the concept of the 'constant flux layer' suggests that the surface layer relations are a good approximation where the spatial inhomogeneities are no more than several m s^{-1}, °C or g · kg^{-1} per 100 km for u, θ and q, respectively. These conditions were generally met during the AMTEX period. In the boundary-layer case, it is normal to use horizontal averaging of data to account for inhomogeneities (and measurement error); vertical averaging has recently been suggested (Arya and Wyngaard, 1975) to account for baroclinity. The scatter in the averaged parameters is then indicative of the suitability of their application in the idealized equations.

In the following sections we make a critical comparison of three sets of eddy correlation flux measurements separated by 1.5 m and 60 km, in order to separate effects of measurement accuracy and field inhomogeneities. The flux data from the two instruments at one station (separation 1.5 m) are compared with surface-layer transfer estimates employing simple hourly observations of mean wind speed, wet- and dry-bulb temperatures (at 10 m) and sea surface temperature, with transfer coefficients selected from recent literature. Such a comparison permits the reassessment of the coefficients.

Application of the boundary-layer relations requires a geostrophic wind obtained from standard synoptic pressure observations, plus temperature and humidity differences across the boundary layer from standard radio-sonde flights. Combination of the direct flux measurements from stations separated by 60 km is used to give fluxes related to a scale comparable to that of the geostrophic wind. In this manner, boundary-layer transfer coefficients are determined and compared to the limited available published values.

2. Site and Synoptic Conditions

Approximately 200 hours of data were extracted from the 15-day observation period of AMTEX 1975. In general these relate to times of a well-defined northerly air flow ('cold-air outbreak') over the sub-tropical East China Sea, when average near-surface wind speeds exceeded 5 m s^{-1}. With air temperatures as much as 10 °C below the sea surface temperature, there were occasions of extreme heat transfer (hourly mean vertical fluxes of sensible heat and latent heat exceeding 300 and 900 Wm^{-2}, respectively). A marked temperature inversion at 1–2 km was prevalent, capping an unstable baroclinic boundary layer.

Average patterns of surface pressure and sea-surface temperature are shown in Figure 1. The grid scale of approximately 500 km connects some eight observation stations used to obtain large-scale wind, temperature and humidity parameters (AMTEX 1975 Data Report, Vol. 1). Within this area, the pressure gradient was predominantly SE–NW with a pronounced curvature of surface isobars of radius $r \simeq 1200$ km, as indicated in the figure. The sea surface temperature gradient is predominantly S–N implying a W–E thermal wind shear.

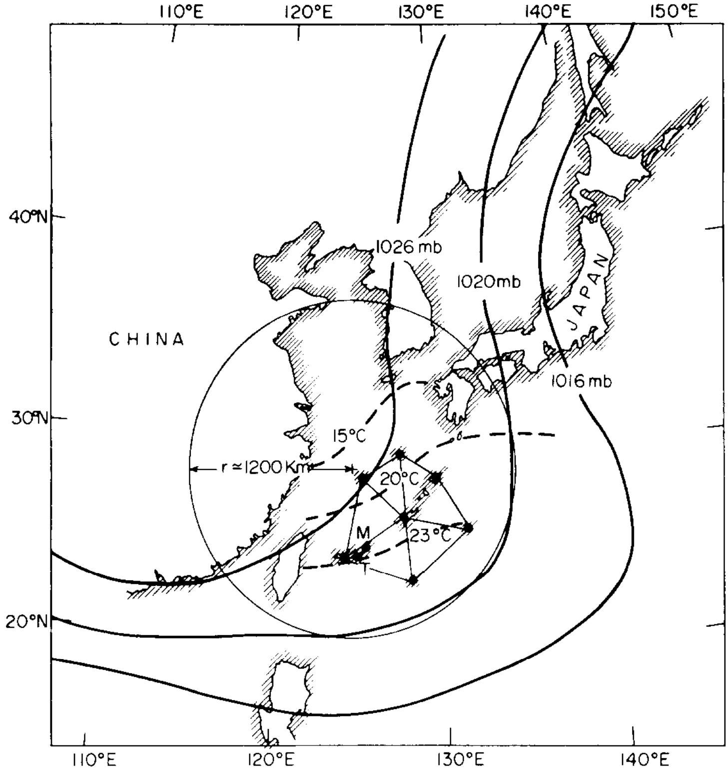

Fig. 1. Schematic of the AMTEX observational area showing stations T (Tarama) and M (Miyako) which provided direct flux and surface-layer measurements. The grid connects stations used in the boundary-layer analysis. Mean isobars (solid line) and sea surface isotherms (dashed line) for the AMTEX period are shown. The circle illustrates a radius of isobar curvature employed in the analysis.

The eddy correlation and mean profile instrumentation were well exposed off the northern coasts of the islands of Tarama and Miyako (marked T and M on Figure 1). Further details of these sites are given in Garratt and Hyson (1975) and Sahashi and Ohtaki (1975).

3. Eddy Correlation Measurements

The eddy correlation sensor array comprised a Gill propellor anemometer for vertical wind component, a six-cup anemometer for the horizontal component, a bead thermistor for temperature and an infrared hygrometer for specific humidity fluctuations and was typically mounted at a height of about 10 m. The output covariances are routinely corrected for sensor frequency response (see e.g., Sahashi *et al.*, 1978), mean-flow distortion (due to proximity of coastal topographic features) and/or sensor misalignment (Hyson *et al.*, 1977). In this Section, errors of instrumental origin are determined by a comparison of sensors separated by 1.5–

2 m in the vertical at Tarama. Differences between mean winds, temperatures, humidities and measured fluxes at the two stations, 60 km apart, with due consideration of statistical sampling and measurement errors, are then used to determine the representativeness of the measurements over a larger area.

Whereas the correlation between simultaneous measurements of a turbulent parameter in a homogeneous surface layer falls off on a scale of metres (see e.g., Koprov and Sokolov, 1973), typical 30–60 min averages of the parameters are expected to agree over a much larger scale. This is supported by observation (e.g., Tsvang *et al.*, 1973; Dyer and Hicks, 1972) for cross-wind separations ranging from 1 to 150 m over a uniform land site. Hourly averages of the Tarama fluxes are compared in Table I, which also includes the results of Dyer and Hicks obtained with similar equipment.

TABLE I

Comparison of eddy fluxes from neighbouring instruments

	AMTEX '75	Dyer and Hicks (1972)
Sensor height	≈ 10 m	4 m
Sensor separation	1.5–2 m (vertical)	1–150 m (horizontal)
Momentum:		
No. of samples	165 (1 hr)	120 ($\frac{1}{2}$ hr)
τ_1/τ_2	1.01	1.05
Scatter* (%)	17	20
Heat:		
No. of samples	111 (1 hr)	135 ($\frac{1}{2}$ hr)
H_1/H_2	0.93	1.06
Scatter (%)	29	12
Latent Heat:		
No. of samples	171 (1 hr)	—
L_wE_1/L_wE_2	1.10	
Scatter (%)	19	

* The r.m.s. of differences corrected for the difference in the means

$$\left(\text{i.e., } \sqrt{\sum \frac{(x-y)^2}{N} - (\bar{x}-\bar{y})^2}\right).$$

The AMTEX long-term means of momentum are in excellent agreement, and in accord with laboratory calibration accuracy of 1 or 2%. The 17% scatter in hourly averages represents local interference; for example, the vertical velocity sensor is extremely sensitive to mounting structure and wind direction. The Dyer–Hicks' results are similar, with a 5% difference in long-term means (a suggestion of decreasing agreement in the data in their paper, with increasing separation from 1 to 150 m, implies that site inhomogeneity also contributed to the scatter). The AMTEX heat fluxes differ by 7% in the long-term average, and have a significantly larger scatter than the momentum results, unlike the Dyer–Hicks' data. The scatter is possibly a consequence of sensor performance in a marine atmosphere in which the microbead thermistor was observed to be susceptible to salt encrustation (see also Friehe and Schmitt, 1976). This generally resulted in an increased thermistor resistance and sensitivity, requiring continuous monitoring of the resistance to indicate the relatively rapid onset (on a time scale of hours) of severe contamination effects. This necessitated thermistor replacement every 1–2 days, with subsequent data rejection. Any remaining data affected by sensitivity would give anomalously high temperature fluctuations, and hence *over*-estimates in the heat flux (see next Section). The latent heat fluxes agree to within 10%, which is the laboratory calibration accuracy of the hygrometers. Scatter is similar to that for momentum.

The mean Tarama fluxes are compared to the Miyako fluxes, 60 km away, in Table II. The separation is now significantly larger than the scale of low-frequency eddies contributing to the measured turbulent fluxes. The corresponding sampling error for 1-hr averages at 10 m with a mean stability ($z/L \approx -0.3$) typical of AMTEX 1975 conditions is approximately 12% for momentum and 7% for heat and water vapour (using approximate formulae derived by Wyngaard, 1973). These statistical uncertainties account for the additional hourly scatter (that are not accounted for by errors in technique) in all three flux comparisons but not the

TABLE II

Tarama–Miyako eddy flux comparison

Sensor height	≈ 10 m
Sensor separation	60 km (30°–60° to wind)
Momentum:	
No. of samples	153 (1 hr)
τ_T/τ_M	0.78
Scatter* (%)	33
Heat flux:	
No. of samples	142 (1 hr)
H_T/H_M	0.77
Scatter (%)	37
Latent heat flux:	
No. of samples	165 (1 hr)
L_wE_T/L_wE_M	0.91
Scatter (%)	23
Mean wind:	
No. of samples	241 (1hr)
U_T/U_M	0.92
Scatter (%)	13

* r.m.s. of differences, corrected for the difference in the means

$$\left(\text{i.e., } \sqrt{\frac{\sum(x-y)^2}{N} - (\bar{x}-\bar{y})^2}\right).$$

reduced agreement in the overall long-term averages. The mean wind speed measured at Miyako is significantly higher than that at Tarama as shown in Table II. The resulting inhomogeneity in the mean wind field accounts for 15% larger momentum flux, and 8% increases in heat and latent heat fluxes at Miyako (based on their wind-speed dependence discussed in Section 4). In contrast, the 13-day averages of air-sea temperature and humidity differences at the two sites agree to within 2%.

Thus the differences in the long-term average fluxes measured by the three independent sets of sensors, and shown in Tables I and II, are essentially explained by calibration uncertainties and mean wind-speed differences between the sites. Comparison of simultaneous hourly measurements between any two sets of sensors shows an inherent scatter which is increased when the sensor separation exceeds the scale of eddies corresponding to one hour.

4. Surface-Layer Bulk Transfer Relations

The conventional bulk transfer method utilising measurements at 10 m above the surface, is based on the following relations,

$$\begin{aligned} \tau &= \rho C_D U_{10}^2 \\ H &= -\rho c_p C_H U_{10}(\theta_{10} - \theta_{\text{sea}}) \\ (L_w E) &= -\rho L_w C_E U_{10}(q_{10} - q_{\text{sea}}) \end{aligned} \tag{1}$$

where ρ, c_p, L_w are air density, specific heat of air at constant pressure and latent heat of vaporisation for water, respectively.

The observations of hourly mean values of wind (U_{10}), potential temperature (θ_{10}) and specific humidity (q_{10}) were made routinely at Tarama (Sahashi and Ohtaki, 1975), as was that of surface temperature (θ_{sea}), from which q_{sea} may be derived. The coefficients C_D, C_H, C_E may be expressed in terms of the product of the 10-m neutral coefficients C_{DN}, C_{HN}, C_{EN} respectively and known stability functions.

The stability functions, via the Businger–Dyer dimensionless gradients (e.g., Dyer, 1974) are calculated in terms of a bulk Richardson number, using U_{10} and the virtual potential temperature at 10 m.

We have taken C_{DN} over sea as

$$C_{DN} = 5.1 \times 10^{-4}\ U_{10}^{0.46}, \tag{2}$$

based on a recent review for $5 < U_{10} < 20\ \text{m s}^{-1}$ by Garratt (1977); this is close to the relation of Wu (1969) for $U_{10} < 15\ \text{m s}^{-1}$, and reflects increasing surface roughness with wave development. Following Pond *et al.* (1974) we use

$$C_{HN} = C_{EN} = 1.5 \times 10^{-3} \tag{3}$$

inferred from limited data for a range of stabilities.

In Table III, the bulk flux estimates at Tarama are compared to the mean Tarama fluxes. It is of great interest that the scatter in hourly values is not significantly different from those in the Tarama two-sensor flux comparison (this is also reflected in the correlation coefficients, typically 0.8–0.9, between the two sets of fluxes directly measured at Tarama, and between one of these sets and the bulk estimates).

TABLE III

Tarama eddy flux – surface bulk flux comparisons

Sensor height	≈ 10 m
Momentum:	
No. of samples	199 (1 hr)
τ_T/τ_{Bulk}	1.11
Scatter* (%)	25
Heat flux:	
No. of samples	199 (1 hr)
H_T/H_{Bulk}	0.81
Scatter (%)	33
Latent heat flux:	
No. of samples	196 (1 hr)
L_wE_T/L_wE_{Bulk}	1.00
Scatter (%)	18

* r.m.s. of differences corrected for the difference in means

$$\left(\text{i.e., } \sqrt{\frac{\Sigma(x-y)^2}{N} - (\bar{x}-\bar{y})^2}\right).$$

This implies that under the AMTEX 1975 conditions, the bulk formulae are as effective as the direct measurements for determining all three surface fluxes for a unique set of constants describing the neutral coefficients. For momentum and latent heat, the assumed values of C_{DN}, C_{EN} give long-term flux averages within about 10% of the direct flux measurements. The bulk heat fluxes are 19% higher than the Tarama eddy correlation heat fluxes (which are in excellent agreement with eddy correlation heat fluxes measured independently by K. Sahashi at the same site (Sahashi *et al.*, 1978)), a difference too large and in the wrong direction (see Section 3) to be explained by uncertainties in the eddy flux measurements.

In view of this, best estimates of the neutral coefficients are determined by re-comparing the two sets of measurements with the stability parameter now determined directly from the flux measurements, (i.e., z/L, L being the Monin–Oboukhov length). The results, with (i.e., C_{DN}, C_{HN}, C_{EN}) and without the stability correction (i.e. C_D, C_H, C_E as in Equation (1)), are expressed in terms of a linear regression against mean wind speed, with standard errors of the regression coefficients in parentheses, in Table IV.

TABLE IV

Surface-layer bulk transfer coefficients with and without stability corrections

AMTEX 1975	$C = a_0 + a_1 \times U_{10}$ (U_{10} in m s^{-1})		Mean value ($\times 10^3$)	AMTEX 1974 mean value ($\times 10^3$)
	$a_0 \times 10^3$	$a_1 \times 10^3$		
C_{DN}	0.77 ($\pm$0.13)	0.085 ($\pm$ 0.014)	1.5 ($\pm$0.4)	1.6 ($\pm$0.5)
C_D	1.00 ($\pm$0.14)	0.078 ($\pm$0.016)	1.7 ($\pm$0.5)	
C_{HN}	0.48 ($\pm$0.11)	0.083 ($\pm$0.013)	1.2 ($\pm$0.4)	1.2 ($\pm$0.3)
C_H	0.68 ($\pm$0.15)	0.081 ($\pm$0.017)	1.4 ($\pm$0.5)	
C_{EN}	1.24 ($\pm$0.09)	0.032 ($\pm$0.010)	1.5 ($\pm$0.3)	1.6 ($\pm$0.3)
C_E	1.73 ($\pm$0.11)	0.005 ($\pm$0.012)	1.8 ($\pm$0.3)	

C_{DN} agrees well, in mean value and wind-speed dependence, with Equation (2). C_{HN} is significantly lower than the 1.5×10^{-3} used in Equation (3) and both C_{HN}, C_H show a wind-speed dependence which has similar significance (at the 99.0% confidence level) and slope to that in C_D, C_{DN}. C_{EN} is in substantial agreement with Equation (3), the wind-speed dependence being significant at the 75% confidence level only and insignificant in the case of C_E. Mean values of neutral bulk transfer coefficients from AMTEX 1974 (Garratt and Hyson, 1975), for similar mean wind speeds, are included in Table IV and show excellent agreement with the present results. The wind-speed dependence of C_{DN} is evident in the 1974 data and a similar behaviour in C_{HN} is suggested in their Figure 4, particularly for wind speeds above 5–6 m s^{-1}.

A comprehensive review of C_H measurements (which in general are reported with no correction for stability to obtain C_{HN}) is given in Friehe and Schmitt (1976). Their conclusion of $C_E/C_H = 1.36$ for a wind speed range $U_{10} \lesssim 8$ m s^{-1} is supported by the AMTEX data where the 1974–1975 averages give for $5 < U_{10} < 15$ m s^{-1}

$$C_{EN}/C_{HN} = 1.30 \pm 0.18 \,. \tag{4}$$

However, extrapolation to higher wind speeds must be made with extreme caution because of the observed variation of C_{HN} with wind speed (not confirmed in the review of data made by Friehe and Schmitt) and the near constant C_{EN}. It has been suggested (Hicks and Hess, 1977) that differences in C_H and C_E could result from a systematic error in the surface temperature measurement when using mercury-in-glass thermometers. For wind speeds exceeding 5 m s^{-1}, however, such an error is unlikely to be more than a few tenths of a °C (e.g., Hasse, 1971). In any case, for the AMTEX conditions, the effect on C_H and C_E (through $\Delta\theta$ and Δq) is similar.

Certainly the different behaviour of C_{HN}, C_{EN}, if confirmed elsewhere, suggests different mechanisms for the transfer of heat and water vapour from the surface to the air. As mentioned by Sheppard (1969), the effects of spray, for instance, may contribute to the heat and water vapour transfer but are difficult to investigate experimentally.

5. Boundary-Layer Bulk Transfer Relations

Arya (1975, 1977) and Yamada (1976) have discussed the momentum and heat flux boundary-layer relations

$$\begin{aligned} \hat{\tau}_0/\rho &= u_*^2 = C_{GD}\hat{V}^2\,, \\ \frac{\hat{H}_0}{\rho c_p} &= u_*\theta_* = C_{GH}\hat{V}(\hat{\theta}_0-\hat{\theta})\,, \end{aligned} \tag{5}$$

where u_*, θ_* are velocity and temperature scaling parameters, respectively, and where the circumflex refers to a large area average and, in the present context, the subscript (0) to sea surface values. Both Yamada (1976) and Arya (1977) chose vertical averages $\langle V_g\rangle$, $\langle\theta\rangle$ of the geostrophic wind V_g, and potential temperature,

$$\begin{aligned} \langle V_g\rangle &= \frac{1}{h}\int_{z_0}^{h} V_g\,\mathrm{d}z \\ \langle\theta\rangle &= \frac{1}{h}\int_{z_0}^{h} \theta\,\mathrm{d}z \end{aligned} \tag{6}$$

as appropriate wind and temperature parameters in Equations (5) in the presence of baroclinity. Here z_0 is the scale of surface roughness and h the height of the boundary layer (generally taken as the level of the lowest temperature inversion).

Nomograms of the bulk transfer coefficients C_{GD}, C_{GH} for different values of h/z_0, and for a range of stabilities given by the bulk Richardson number

$$\mathrm{Ri}_B = \frac{(g/\theta_a)h(\langle\theta\rangle-\theta_0)}{\langle V_g\rangle^2} \tag{7}$$

were presented by both authors.

In the presence of trajectory curvature (as evidenced in Figure 1), the geostrophic wind $\langle V_g\rangle$ should be replaced by a gradient wind V, where, following Hess (1959, p. 185),

$$V = -\frac{rf}{2}+\sqrt{\frac{f^2r^2}{4}+fr\langle V_g\rangle}\,. \tag{8}$$

For normal anticyclonic flow in the Northern Hemisphere, the Coriolis parameter f

and the radius of curvature r are positive, and $V, \langle V_g \rangle$ are negative in this expression.

The present analysis applies a linear regression equation to sea-level pressures and air temperatures, observed at 6-hrly intervals at each of the eight stations defining the grid of Figure 1, to obtain mean horizontal pressure and temperature gradients at the surface. At corresponding times for all stations, plots of the radiosonde temperature profiles are used to determine a boundary-layer height (at the lowest inversion), mean temperature $\langle \theta \rangle$, and confirmation that horizontal temperature gradients are essentially constant throughout the boundary layer. Using an average, h, of the boundary-layer heights (0.8 to 1.8 km with a station-to-station scatter $\approx 17\%$), the horizontal temperature gradient values provide a thermal wind as a function of height. When vectorially added to the surface geostrophic wind (from the pressure gradient) and integrated from 0 to h km, these give an areal average of the boundary-layer vertically averaged wind $\langle \hat{V}_g \rangle$ defined in (6).

To obtain V for use in Equations (5) and (7), which incorporates the effect of trajectory curvature, we have used Equation (8) for each 6-hrly value of the large-scale $\langle \hat{V}_g \rangle$, with the long-term mean radius, r, increasing from 1200 km at the surface (as in Figure 1) to infinity at the height at which geostrophic balance is achieved (~ 1.5 h). The subjectivity of manual chart analyses and the lack of detailed information on the pressure gradients with height do not warrant a more detailed estimate. The gradient wind is on average 7% higher than $\langle \hat{V}_g \rangle$.

In the absence of routine sea-surface measurements at all stations, we have used mean February sea-surface isotherms for the period 1953 to 1957 (GARP Publication Series No. 13, 1973). These agree to within 0.5 °C with 1975 measurements at Miyako and Tarama and give $\hat{\theta}_0$, which with $\langle \hat{\theta} \rangle$ from the radiosonde data gives $(\hat{\theta}_0 - \langle \hat{\theta} \rangle)$, (2–12 °C with a 19% station-to-station scatter) for use in Equations (5) and (7).

Using 6-hr averages of the directly measured surface fluxes at Tarama and Miyako for $\hat{\tau}_0, \hat{H}_0$ in (5) we have calculated the C_{GD}, C_{GH} values summarized in Table V. These are averaged for 3 Ri_B ranges to facilitate comparison with values taken from the $u_*/\langle V_g \rangle$ and $\theta_*/\Delta\theta$ nomograms of Yamada (1976) and Arya (1977), using $h/z_0 \simeq 5 \times 10^6$ applicable in our case.

The $C_{GD} (= (u_*/\langle V_g \rangle)^2)$ values are in good agreement with the nomogram values of Arya, and not incompatible with the lower Yamada values. The C_{GD} increase with increasing instability, through a linear regression, is similar to the variation shown in the nomograms. It is important to note that agreement worsens, with C_{GD} values higher, if the layer-averaged geostrophic wind is used rather than the analogous gradient wind.

The AMTEX C_{GH} values (Table V) show less scatter than the C_{GD} values (probably reflecting the small scatter in $\Delta\theta = (\theta_0 - \langle \theta \rangle)$ from station to station), and exhibit a similar tendency with instability.

Although there is excellent agreement in the C_{GH} $(= \theta_*/\Delta\theta \cdot u_*/\langle V_g \rangle)$ values of Table V, the discrepancies in C_{GD} above imply $\theta_*/\Delta\theta$ values of Yamada and Arya

TABLE V

Boundary-layer bulk transfer coefficients as a function of stability

$-\mathrm{Ri}_B$	$C_{GD} \times 10^3$			$C_{GH} \times 10^3$			$C_{GE} \times 10^3$
	AMTEX '75	Yamada	Arya	AMTEX '75	Yamada	Arya	
$0 \to 3$	1.2 ± 0.3	0.7	1.0	1.3 ± 0.3	1.2	1.2	1.5 ± 0.2
$3 \to 5$	1.6 ± 0.7	0.8	1.2	1.4 ± 0.4	1.6	1.5	1.7 ± 0.2
>5	1.4 ± 0.5	0.9	1.4	1.4 ± 0.4	1.8	1.8	1.9 ± 0.3
All data	1.4 ± 0.5			1.4 ± 0.4			1.7 ± 0.3

some 20–30% above our values. This difference reveals an error in the derivation of the nomogram values, resulting in an overestimate of this order in $\theta_*/\Delta\theta$. The nomograms are based on the relation

$$\frac{\theta^*}{\Delta\theta} = k\left(\ln\left(\frac{h}{z_0}\right) - C(\mathrm{Ri}_B)\right)^{-1} \tag{9}$$

with the similarity function $C(\mathrm{Ri}_B)$ determined from the Wangara data set (Clarke *et al.*, 1971), with $h/z_0 \simeq 5 \times 10^5$. The values were determined from a surface temperature θ_0, not measured directly, but found by extrapolation of temperature profiles to a height z_0. Garratt and Hicks (1973), Thom (1972) and Garratt (1978a), for example, argue that the actual surface temperature is more closely related to θ_0 at z_T which is typically an order of magnitude smaller than z_0 in aerodynamically rough flow over the Wangara site (here z_T is a roughness scaling length for heat, analogous to z_0 for momentum). Both Yamada and Arya used data corresponding to unstable conditions at Wangara, for which we find $\theta_0(z_T) - \theta_0(z_0) \sim 1$–$2\,°\mathrm{C}$, when values of $\Delta\theta$ are typically ~ 4–$6\,°\mathrm{C}$. There is then a corresponding 30% overestimate in $\theta_*/\Delta\theta$. (A more detailed revision of values of $C(\mathrm{Ri}_B)$ in Equation (9), based on $\theta_0(z_T)$, will be given elsewhere.)

In order to obtain a crude estimate of C_{GE} we have adopted a similar procedure to that used for temperature, to obtain $(q_0 - \langle\hat{q}\rangle)$ the humidity difference between the surface value (in our case the saturated vapour pressure at θ_0), and the averaged value up to a height h. Only two radiosonde stations (with complete dew-point profiles) are employed, and the C_{GE} values for three Ri_B ranges are included in Table V. We observe again an increase in C_{GE} with increasing instability.

It is striking that the C_{GE}/C_{GH} ratio of 1.26 is similar to C_{EN}/C_{HN} for the surface-layer data, prompting a final comment. As discussed above, the consistency of results from several independent sets of flux measuring sensors employed on two separate occasions (AMTEX 1974, 1975) provides the main evidence for correct determination of H/L_wE and hence the above ratios. Additional support is found in long-term measurements of $(H+L_wE)$ over land using the same instrumentation, which satisfy energy balance requirements with values to within 5% of net radiation minus ground heat flux $(R-G)$ (Garratt, 1978b).

6. Summary

Results presented in previous sections relate specifically to air-sea transfer occurring during the prolonged cold air outbreak of the AMTEX 1975 observation period. They may be summarized as follows:

(1) Comprehensive hourly observations of vertical fluxes using the eddy correlation technique show that long-term averages of the fluxes, over an area of scale at least 60 km, are determined to within 10–15% when the non-uniformity in the large-scale mean wind field is taken into consideration. This uncertainty is of instrumental origin.

The accuracy of the large-scale fluxes determined from a single hourly observation is further influenced by sampling errors, and is about 30%.

(2) Surface-layer bulk transfer formulae predict momentum and latent heat fluxes to an accuracy comparable with the direct eddy correlation method, using transfer coefficients from the recent literature. The data essentially confirm C_{DN} increasing with U_{10} and a constant $C_{EN}=1.5\times10^{-3}$. In contrast, use of $C_{HN}=1.5\times10^{-3}$ considerably overestimates heat fluxes. The data suggest that C_{HN} increases significantly with wind speed over the observed range 5–15 m s^{-1}, and on average is 30% lower than C_{EN}.

(3) The 6-hrly averaged surface fluxes are used to obtain preliminary estimates of boundary-layer transfer coefficients, employing vertical averaging of the geostrophic wind and measured mean temperature and humidity profiles to account for baroclinity. Both C_{GD} and C_{GH} are in broad agreement with recently published values, but with differences which identify physically (if not statistically) significant corrections. Without the inclusion of the effects of trajectory curvature the AMTEX C_{GD} values exceed the published values by an additional 15–20%. If the differences in C_{GD} are taken into account in the comparison of C_{GH} $(=\theta_*/\Delta\theta\cdot u_*/\langle V_g\rangle)$, the published $\theta_*/\Delta\theta$ values exceed the AMTEX values by 20–30%. A discrepancy of this magnitude in the published values is identified with inappropriate surface temperatures used in the $\theta_*/\Delta\theta$ derivation.

Finally, for water vapour, tentative values of C_{GE}, with $\bar{C}_{GE}/\bar{C}_{GH}=1.2\pm0.3$, are determined.

Acknowledgements

The cooperation of Dr K. Sahashi, Okayama University, in providing the data used in the surface-layer bulk transfer estimates, is gratefully acknowledged. Mr. G. Grauze assisted with the collection of data at Miyako.

References

AMTEX '75, *Data Report, Vol. 2, Aerological Data*, Ed. by K. Ninomiya, Meteorological Research Institute, Tokyo.

AMTEX '75, *Data Report, Vol. 7, Analysis Charts*, Ed. by A. Katayama and Y. Okamura, Meteorological Research Institute, Tokyo.

Arya, S. P. S.: 1975, 'Geostrophic Drag and Heat Transfer Relations for the Atmospheric Boundary Layer', *Quart. J. Roy. Meteorol. Soc.* **101**, 147–161.

Arya, S. P. S.: 1977, 'Suggested Revisions to Certain Boundary-Layer Parameterization Schemes Used in Atmospheric Circulation Models', *Mon. Wea. Rev.*, **105**, 215–227.

Arya, S. P. S. and Wyngaard, J. C.: 1975, 'Effect of Baroclinicity on Wind Profiles and the Geostrophic Drag Law for the Corrective Planetary Boundary Layer', *J. Atmos. Sci.* **32**, 767–778.

Clarke, R. H. with Dyer, A. J., Brook, R. R., Reid, D. G., and Troup, A. J.: 1971, 'The Wangara Experiment – Boundary Layer Data', *Division of Meteorological Physics Tech. Paper No. 19*, CSIRO, Australia.

Delsol, F., Miyakoda, K., and Clarke, R. H.: 1971, 'Parameterized Processes in the Surface Boundary Layer of an Atmospheric Circulation Model', *Quart. J. Roy. Meteorol. Soc.* **97**, 181–208.

Dyer, A. J.: 1974, 'A Review of Flux-Profile Relationships', *Boundary-Layer Meteorol.* **7**, 363–372.

Dyer, A. J. and Hicks, B. B.: 1972, 'The Spatial Variability of Eddy Fluxes in the Constant Flux Layer', *Quart. J. Roy. Meteorol. Soc.* **98**, 206–212.

Friehe, C. A. and Schmitt, K. F.: 1976, 'Parameterizations of Air-Sea Interface Fluxes of Sensible Heat and Moisture by the Bulk Aerodynamic Formulas', *J. Phys. Ocean*, **6**, 801–809.

Gadd, A. J. and Keers, J. F.: 1970, 'Surface Exchanges of Sensible and Latent Heat in a 10-level Model Atmosphere', *Quart. J. Roy. Meteorol. Soc.* **96**, 297–308.

GARP Publication Series No. 13, *The Air-Mass Transformation Experiment*, WMO-ICSU, July 1973.

Garratt, J. R.: 1977, 'Review of Drag Coefficients Over Oceans and Continents', *Mon. Wea. Rev.* **105**, 915–929.

Garratt, J. R.: 1978a, 'Transfer Characteristics for a Heterogeneous Surface of Large Aerodynamic Roughness', *Quart. J. Roy. Meteorol. Soc.* **104**, in press.

Garratt, J. R.: 1978b, 'Flux Profile Relations Over Tall Vegetation', *Quart. J. Roy. Meteorol. Soc.* **104**, in press.

Garratt, J. R. and Hicks, B. B.: 1973, 'Momentum, Heat and Water Vapour Transfer to and from Natural and Artificial Surfaces', *Quart. J. Roy. Meteorol. Soc.* **99**, 422, 680–687.

Garratt, J. R. and Hyson, P.: 1975, 'Vertical Fluxes of Momentum Sensible Heat and Water Vapour During the Air Mass Transformation Experiment (AMTEX) 1974', *J. Meteorol. Soc. Japan* **53**, 149–160.

Hasse, L.: 1971, 'The Sea Surface Temperature Deviation and the Heat Flow at the Sea-Air Interface', *Boundary-Layer Meteorol.* **1**, 368–379.

Hicks, B. B. and Hess, G. D.: 1977, 'On the Bower Ratio and Surface Temperature at Sea', *J. Phys. Ocean.* **7**, 141–145.

Hyson, P., Garratt, J. R., and Francey, R. J.: 1977, 'Algebraic and Electronic Corrections of Measured *uw* Covariance in the Lower Atmosphere', *J. Appl. Meteorol.* **16**, 43–47.

Koprov, B. M. and Sokolov, D. Yu.: 1973, 'Spatial Correlation Functions of Velocity and Temperature Components in the Surface Layer of the Atmosphere', *Izvestia Atmos. Ocean. Phys.* **9**, 178–182.

Pond, S., Fissel, D. B., and Paulson, C. A.: 1974, 'A Note on Bulk Aerodynamic Coefficients for Sensible Heat and Moisture Fluxes', *Boundary-Layer Meteorol.* **6**, 333–339.

Priestley, C. H. B., and Taylor, R. J.: 1972, 'On the Assessment of Surface Heat Flux and Evaporation Using Large Scale Parameters,' *Mon. Wea. Rev.* **100**, 81–92.

Sahashi, K. and Ohtaki, E.: 1975, 'Heat Flux Measurement at Tarama Island – Preliminary Report of AMTEX '74', *Bulletin of School of Education, Okayama University* **43**, 61–70.

Sahashi, K., Maitani, T., Francey, R. J., and Garratt, J. R.: 1978, 'Turbulence Sensor Comparisons at AMTEX 1975', to be submitted for publication.

Sheppard, P. A.: 1969, 'The Atmospheric Boundary Layer in Relation to Large Sclae Dynamics', *The Global Circulation of the Atmosphere* (Ed. G. A. Gorby), Publ. Roy. Meteorol. Soc. 91–112.

Thom, A. S.: 1971, 'Momentum Absorption by Vegetation', *Quart. J. Roy. Meteorol. Soc.* **97**, 414–428.

Tsvang, L. R., Koprov, B. M., Zubkovskii, S. L., Dyer, A. J., Hicks, B. B., Miyake, M., Stewart, R. W. and McDonald, J. R.: 1973, 'A Comparison of Turbulence Measurements by Different Instruments: Tsimlyansk Field Experiment 1970', *Boundary-Layer Meteorol.* **3**, 499–521.
Wu, Jin: 1969, 'Wind Stress and Surface Roughness at Air-Sea Interface', *J. Geophys. Res.* **74**, 444–455.
Wyngaard, J. C.: 1973, 'On Surface-Layer Turbulence', *Workshop on Micrometeorology* (Ed. Duane A. Haugen), Publ. American Meteorol. Soc., 191–149.
Yamada, Tetsuji: 1976, 'On the Similarity Functions *A*, *B* and *C* of the Planetary Boundary Layer', *J. Atmos. Sci.* **33**, 781–793.