Measurements of Turbulent Fluxes in Bass Strait

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ABSTRACT

Measurements of turbulent momentum, heat and moisture fluxes have been made in Bass Strait from a stable platform, at a height of approximately 5 m above water. Direct measurements of these fluxes are compared with estimates obtained from spectra of velocity, temperature and humidity fluctuations with the use of the inertial dissipation technique. Directly measured momentum and moisture flux values are in reasonable agreement with inertial dissipation values. The sensible heat flux obtained by the inertial dissipation technique is about twice as large as the directly measured heat flux. The dependence on wind speed of bulk transfer coefficients of momentum, heat and moisture and of variances of velocity and scalar fluctuations is discussed and compared with available data.

1. Introduction

Transfer rates of momentum, heat and moisture through the marine boundary layer have a large effect on the global weather system, oceanic and atmospheric circulations. Measurements of momentum, heat and moisture fluxes are therefore important inputs to numerical weather prediction schemes. These measurements are best performed from a fixed stable platform as wave-induced motions on any other type of platform can interfere significantly with the signals (especially those associated with the vertical w fluctuation) obtained from instrument sensors. Flux estimates from stable platforms have been reported by De Leonibus (1971) who used the Argus Island tower at a water depth of 60 m and by Zubkovskii *et al.* (1974) who used the Caspian Sea platform at a water depth of 40 m. In Australia, an almost ideal location for the measurement of fluxes over the ocean is provided by the offshore oil rig platforms in Bass Strait. Momentum and heat flux measurements from the Marlin platform in Bass Strait (40 km off the Victorian coast at a water depth of about 60 m) were reported by Hicks and Dyer (1970). They concluded that valuable and accurate measurements, particularly under high wind speed conditions, could be obtained using drilling platforms at sea. In the present investigation momentum, heat and moisture flux measurements were made, as part of a larger air-sea interaction study program (project B.A.S.S. for Basic Air-Sea Studies), from the Kingfish B oil-production platform, also in Bass Strait, at a water depth of about 70 m. Some of the activities included in the air-sea interaction study in Bass Strait have been described in Antonia *et al.* (1974).

Estimates for all fluxes have been determined by two different methods. Direct calculations of covariances were made from recorded velocity, temperature and humidity fluctuations. Indirect covariance estimates were made from a knowledge of the spectra of these fluctuations in the inertial subrange and by assuming a certain form for the budgets of turbulent energy and of variances of scalar fluctuations. The results are presented in the form of bulk transfer coefficients and are compared with other results in the literature.

2. Experimental techniques

The experiment was conducted on Kingfish B (Fig. 1a), the ESSO-BHP natural gas platform which stands in Bass Strait (148° 9′E, 38° 36′S) about 80 km off the Gippsland coast of Victoria. Instruments used for the flux measurements were mounted at the end of a horizontal boom fastened to one of the platform legs on the western side of Kingfish B (Fig. 1b). A vertical mast (5 cm diameter aluminium pipe) onto which most of the instrument sensors were fastened, was attached to the end of the boom. The boom was of sufficient

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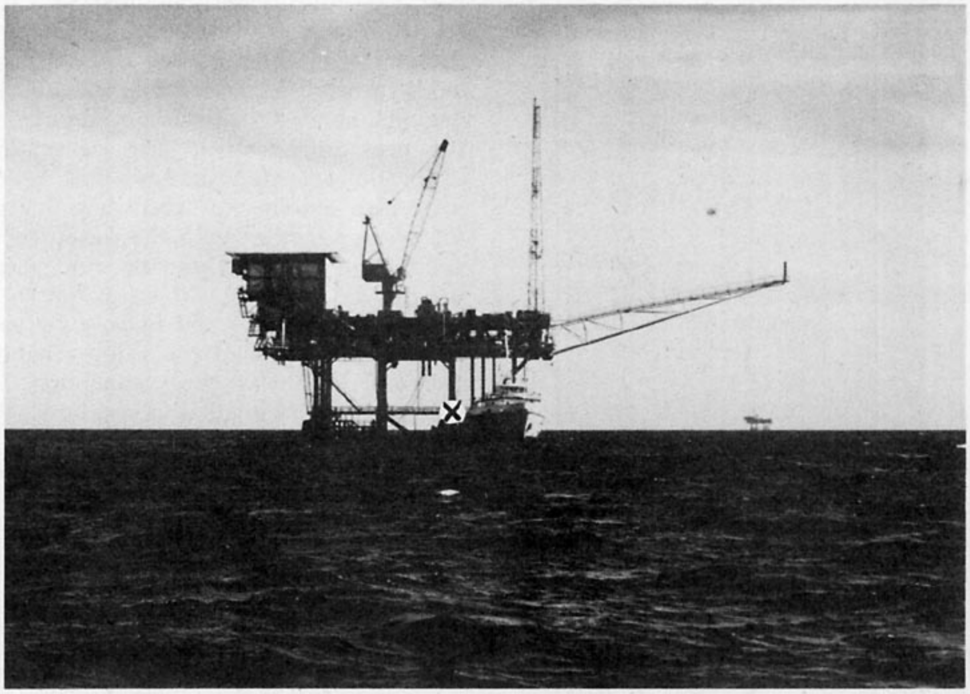

FIG. 1a. General view of Kingfish B oil production platform looking from the east. × indicates location of boom on west side of platform.

length (~15 m) to allow the measurements to be made clear of the disturbance of the platform. Preliminary experiments in a wind tunnel on a model (scale 1:300) of the platform indicate negligible interference at a position corresponding to the location of the mast. The boom could easily be raised to allow easy access to the instruments on this mast from one of the main decks of the platform. A gill propeller array was mounted at a height of about 5 m above the water surface to measure relatively low-frequency u (streamwise), v (lateral) and w (normal) velocity fluctuations. When positioned on the boom for an experimental run the u propeller was aligned to the wind direction and the w propeller was leveled by eye to the horizon. The adequacy of the propeller alignment to the wind direction is discussed in Section 4. Also mounted with the propeller array were a Lyman-alpha humidiometer to measure humidity fluctuations q and a thermistor to measure relatively low-frequency temperature fluctuations θ. The plane of the w propeller was 40 cm above the axis of the u propeller. The thermistor and Lyman-α meter were located 27 and 33 cm, respectively, below the axis of the u propeller. High-frequency velocity and temperature fluctuations were obtained with a hot wire and cold wire, respectively. The hot and cold wires were mounted respectively at the levels of w propeller and Lyman-α humidiometer, at a lateral separation of about 40 cm. The hot wire (5 μm diameter platinum) was operated by a DISA 55M01 constant temperature anemometer while the cold wire (0.6 μm diameter platinum, 0.8 mm length) was operated by a constant current anemometer. The value of the current was low enough (0.1 mA) for the wire to be sensitive to temperature fluctuations only. Neither the hot wire anemometer nor the Lyman-α humidiometer was linearized. In the case of the hot wire, no corrections were made for either the temperature sensitivity (neglible as the anemometer bridge was operated at an overheat ratio of 0.8) or the lateral velocity sensitivity.

Mean wind velocity profiles were measured with an array of cup anemometers. A propeller anemometer was fixed to the radio tower (see Fig. 1b) to provide a continuous record of the wind velocity at a reference height of 49 m above the sea level. The mean water level was continuously monitored with a resistance wire gage suspended from the boom but at a separation of about 2 m from the vertical mast so as to minimize any possibility of wave-induced motion of the instrument mast. The sea surface temperature T_s was also continuously monitored with the use of a Hewlett-Packard quartz crystal thermometer (Model 2801A) located in a sea surface temperature bucket attached to a buoy. The mean level of the probe below the water surface was about 0.3 m. The air temperature and dew point were measured with an EG & G (Model 110S) unit,

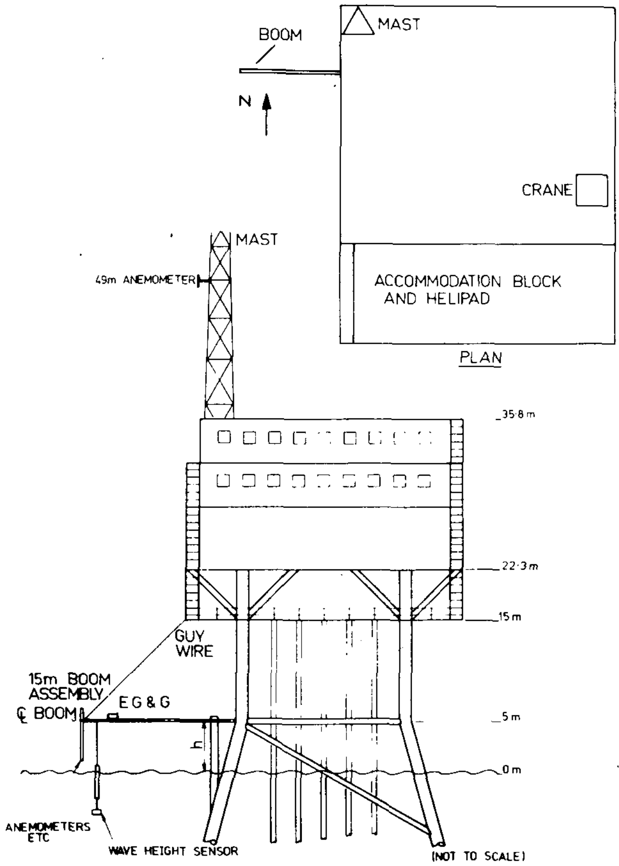

FIG. 1b. Layout of instrumentation on Kingfish B oil production platform.

located on the boom 3 m from the mast. The dew-point voltage output of this instrument was used for an *in situ* calibration of the Lyman-α hygrometer. Static calibrations of the wires (both hot and cold) were obtained in the laboratory (University of Newcastle) prior to the field trip to the platform. Most of the wires that survived the field trip were recalibrated in the laboratory. There was, in general, satisfactory agreement between the pre- and postfield trip calibrations. *In situ* calibrations in the field of hot and cold wires were also obtained by recording fluctuations from the wires together with the outputs from the cup anemometer closest to the hot wire and from the EG & G unit.

Voltages proportional to u, w, q and θ fluctuations were recorded on a four-channel Hewlett-Packard 7060 FM tape transport. Most of the recordings were made at a speed of 24 mm s^{-1} (−3 dB point of tape recorder $\approx$ 375 Hz), while a few runs required for the direct dissipation technique were recorded at a speed of 380 mm s^{-1} (−3 dB point $\approx$ 6 kHz). The tape recorder has a dynamic range of ± 5 V and a nominal signal/noise ratio of 46 dB at the recording speed of 24 mm s^{-1}. The tapes were later played back and digitized at various sampling frequencies (ranging from 1.5 Hz to 10 kHz) in the Faculty of Engineering Computing Centre of the University of Sydney. The digital records were processed on a PDP 11/45 computer and a 1904 ICL computer at the University of Newcastle. Prior to digitization, the signals were low-pass filtered with the −3 dB cutoff frequency set at one-half the sampling frequency. The majority of the records used for direct flux estimates were digitized at a sampling frequency of 20 Hz as only minor contributions to the covariances are expected from frequencies greater than 10 Hz. Most of the digital records analyzed had a 30 min duration but records as long as 66 min and as short as 20 min were examined to determine the covari-

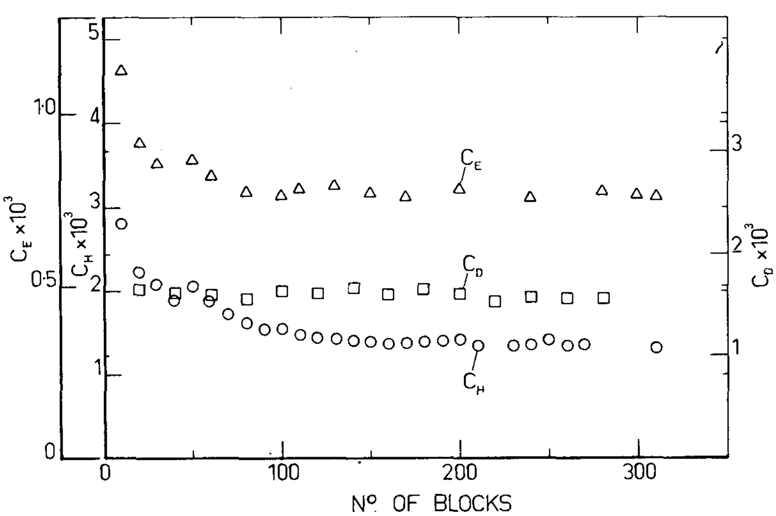

FIG. 2. Variation of flux covariances with record duration (one block represents ~12.8 s).

ance. For a typical case, Fig. 2 shows the dependence on record duration of the running mean values for the important covariances $\overline{uw}$, $\overline{wq}$ and $\overline{w\theta}$. A record duration of 30 min (approximately 140 data blocks in Fig. 2) appears to be sufficient to achieve stationary averages.

3. Direct flux measurements

Most of the flux estimates were made on 23 August 1976, for which the wind direction (northwest–west–southwest sector) was favorable with respect to the fixed location of the boom on the platform. The wind direction (Fig. 3) was reasonably steady over most of the measurement period. Values of U_5 (wind velocity at the 5 m height) and U_{49} (wind velocity at the 49 m height, as obtained by the wind vane on the radio tower) are shown in Fig. 3 for the period of the experiment. A rough estimate of the roughness length z_0 has been obtained with the use of Charnock's (1955) formula $z_0 = \alpha u_*^2/g$, where u_* is the friction velocity $(-\overline{uw})^{1/2}$. The constant α was chosen to be 0.016, the value obtained by Hicks (1972) for fairly low wind speeds (4 m s^{-1} < U_{10} < 7 m s^{-1} in Bass Strait. Values of z_0 obtained from the present direct momentum flux values are seen in Fig. 3 to increase as a result of a general increase in U_5 and fairly constant C_D during the experimental period.

Momentum, heat and moisture fluxes are presented in Fig. 4 in terms of the bulk coefficients C_D, C_H and C_E. The kinematic shear stress $-\overline{uw}$, assumed here equal to the surface stress, is related to U_5 by

$$-\overline{uw} = C_D U_5^2. \tag{1}$$

Values of C_D obtained from (1) using $\overline{uw}$ derived

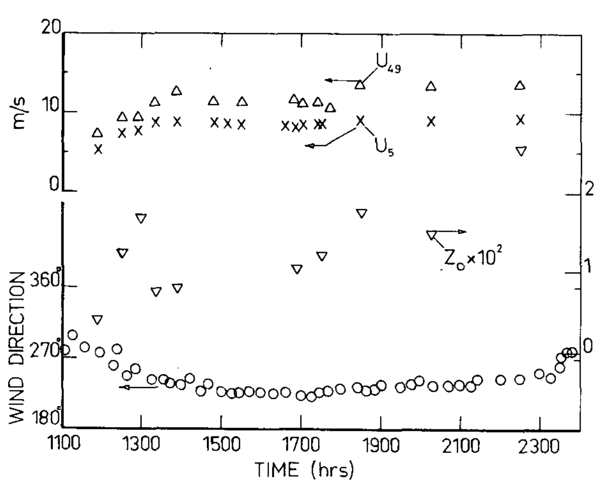

FIG. 3. Wind speed, direction and roughness height during observational period.

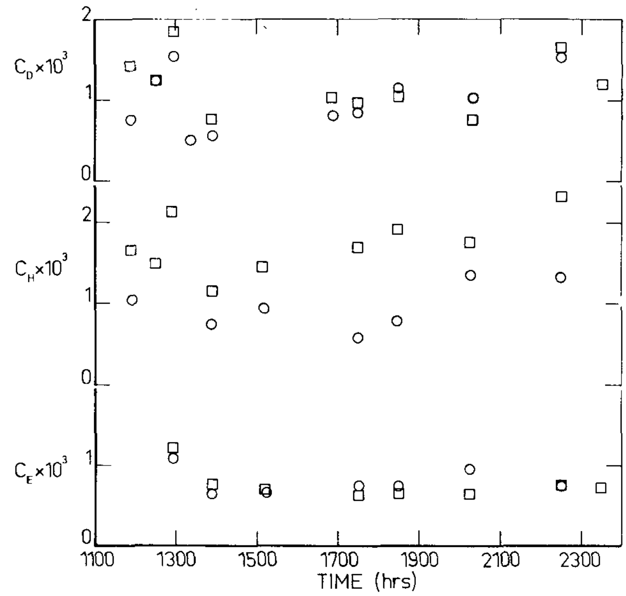

FIG. 4. Bulk coefficients of momentum, heat and moisture fluxes: circles, direct measurement; squares, inferred from inertial dissipation technique.

from the hot wire u or propeller u and propeller w signals are shown in Fig. 4. The thermometric heat flux $\overline{w\theta}$ (with θ obtained from thermistor or cold wire) is represented by

$$\overline{w\theta} = C_H U_5(T_s - T_5), \tag{2}$$

where T_5 is the potential temperature at 5 m height. Values of C_H (average value $\approx 0.89 \times 10^{-3}$ given in Table 1) are slightly lower than C_D but approximately equal to the moisture flux bulk transfer coefficient C_E, defined here by

$$\overline{wq} = C_E U_5(Q_s - Q_5), \tag{3}$$

where Q_s is the sea surface absolute humidity (saturation value at T_s) and Q_5 is the absolute humidity of the air at 5 m. Other nondimensional bulk transfer coefficients associated with other combinations of velocity and scalar fluctuations are given in Fig. 5. These coefficients are defined by

$$C_{u\theta} = \overline{u\theta}/U_5(T_s - T_5), \tag{4}$$

$$C_{uq} = \overline{uq}/U_5(Q_s - Q_5), \tag{5}$$

$$C_{q\theta} = \overline{q\theta}/(T_s - T_5)(Q_s - Q_5). \tag{6}$$

The average values of C_{uq}, $C_{q\theta}$ and $C_{u\theta}$ over the period of the experiment all exceed (Table 1) 4×10^{-3} and are significantly larger than C_D, C_H or C_E. It should be noted that no corrections for the surface drift velocity have been made in Eqs. (1)–(5).

TABLE 1. Regression lines of bulk coefficients on wind speed U_5.

Parameter	Mean ± standard deviation	Linear regression ± standard deviation	Correlation coefficient	Standard deviation Intercept	Standard deviation Slope
$10^3 C_D$	1.05 ± 0.35	$0.74 + 0.034U_5$ (±0.39)	0.10	±0.86	±0.10
$10^3 C_D^*$	1.25 ± 0.31	$1.91 - 0.084U_5$ (±0.33)	0.34	±0.70	±0.08
$10^3 C_H$	0.89 ± 0.24	$1.13 - 0.03U_5$ (±0.29)	0.18	±0.70	±0.08
$10^3 C_H^*$	1.67 ± 0.37	$1.61 + 0.008U_5$ (±0.41)	0.02	±0.93	±0.11
$10^3 C_E$	0.82 ± 0.15	$2.09 - 0.14U_5$ (±0.15)	0.49	±1.02	±0.12
$10^3 C_E^*$	0.75 ± 0.17	$2.47 - 0.19U_5$ (±0.14)	0.66	±0.73	±0.08
$10^3 C_{uq}$	4.34 ± 0.63	$11.74 - 0.85U_5$ (±0.30)	0.86	±1.95	±0.22
$10^3 C_{q\theta}$	4.96 ± 1.55	$17.46 + 1.43U_5$ (±1.61)	0.53	±10.03	±1.14
$10^3 C_{u\theta}$	4.30 ± 2.30	$-4.62 + 1.04U_5$ (±2.46)	0.25	+13.95	±0.59

4. Inertial dissipation results

To estimate $-\overline{uw}$ using the inertial dissipation technique, the mean turbulent energy dissipation ϵ is first inferred from the inertial subrange spectral density of u and the shear stress is subsequently deduced from the turbulent kinetic energy budget equation. The inertial subrange of u, θ and q is clearly in evidence[1] in Fig. 6. Here we assume the validity of the turbulent kinetic energy budget obtained for the overland Kansas experiment by Wyngaard and Coté (1971) and recently confirmed, for a Minnesota overland experiment, by Champagne *et al.* (1977). The form obtained by Wyngaard and Coté for the stability dependent imbalance (due apparently to the pressure diffusion term), between the production and dissipation of turbulent energy, is assumed together with the stability dependence of the mean velocity gradient as given by Businger *et al.* (1971). The resulting expression for u_* (see Champagne *et al.*, 1977) is given by

$$u_* = \left[\frac{\kappa z \epsilon}{(1 + 0.5|z/L|^{2/3})^{3/2}}\right]^{1/3}, \quad (7)$$

where L is the Monin-Obukhov length and κ the von Kármán constant, assumed here equal to 0.35, the value used by Wyngaard and Coté (1971) and Champagne *et al.* (1977). The length L is computed here using the expression (Champagne *et al.*, 1977).

$$L = \frac{-u_*^3[1 + 0.472 \times 10^{-3}(T/273)Q]}{\kappa g[(\overline{w\theta}/T) + 0.472 \times 10^{-3}(T/273)\overline{wq}]}, \quad (8)$$

where Q is the mean humidity (g m^{-3}) and T the absolute temperature (K). Values of L obtained at a nominal 5 m height using directly measured values of $\overline{uw}$, $\overline{w\theta}$ and $\overline{wq}$ are shown in Fig. 7 together with the temperature and humidity differences between sea level and 5 m. After a substantial increase during the initial period of the experiment, L remained approximately constant with a value of about -100 m, suggesting only moderately unstable conditions during the period of observation. Estimates of ϵ were obtained from the inertial subrange behavior of the spectrum of u, *viz.*,

$$\phi_u(k_1 z) = \beta_u(\epsilon z)^{2/3}(k_1 z)^{-5/3}, \quad (9)$$

where k_1 is the one-dimensional wavenumber $2\pi n/U$ (n is the frequency) and β_u the Kolmogorov constant, here taken as 0.5. This value for β_u appears to have received substantial support in the literature and is very nearly equal to the average value inferred, in this study, from the measured spectra of u and direct estimates of ϵ using the isotropic assumption

$$\epsilon = 15\nu\overline{(\partial u/\partial t)^2}/U^2.$$

These estimates were made from four sets of recorded velocity time derivative signals, digitized at twice the Kolmogorov frequency (~1.5 kHz).

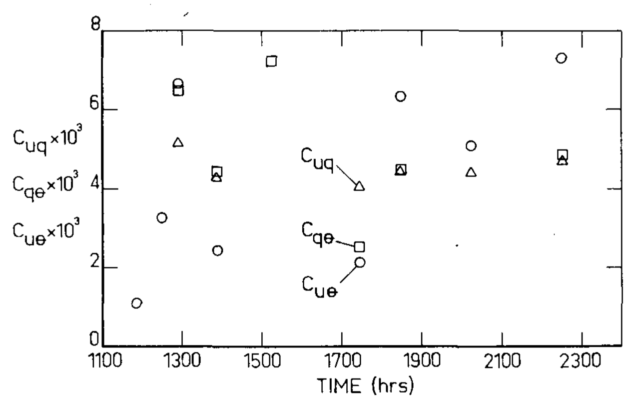

FIG. 5. Bulk transfer coefficients associated with combinations of horizontal velocity and scalar fluctuations.

[1] It should be noted that the Lyman-α instrument used here for humidity measurements is the same instrument which was used in the Minnesota experiment by Champagne *et al.* (1977). These authors stated that the frequency response of the instrument was no better than dc to 10 Hz. The present results on the other hand suggest that the upper frequency response (set essentially by the spatial separation between the Lyman-α detector and the source tube) is good up to about 100 Hz. Evidently the response of the device is not well understood.

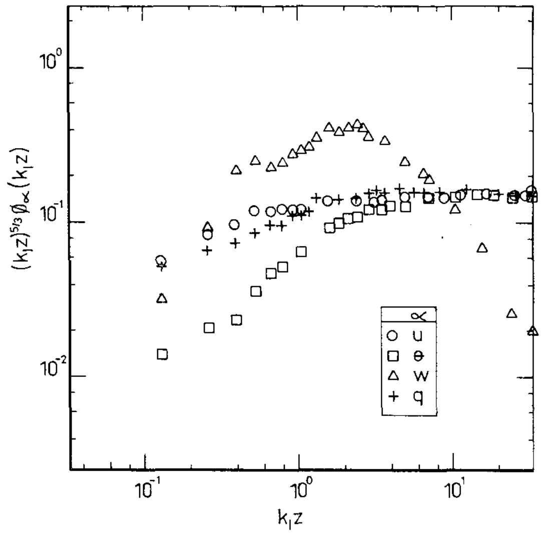

FIG. 6. Spectral densities of horizontal velocity, vertical velocity, temperature and humidity (sampling frequency = 20 Hz).

Because of the relatively low turbulence levels, no correction was made for ϵ in the previous equation to allow for the possible inadequacy of Taylor's hypothesis (see, e.g., Champagne *et al.*, 1977). Using average values of $\sigma_u/U \approx 0.1$ and $\sigma_w/U \approx 0.045$ (see Table 1), and making the conservative assumption that $\sigma_v/U \approx \sigma_u/U$, the present values overestimate ϵ by about 3.5%.

Inertial dissipation estimates of $\overline{w\theta}$ and $\overline{wq}$ are obtained here by assuming equality of production and dissipation of the variance of temperature and humidity, *viz.*,

$$-2\overline{w\theta}\frac{\partial T}{\partial z} = \chi_\theta, \tag{10}$$

$$-2\overline{wq}\frac{\partial Q}{\partial z} = \chi_q. \tag{11}$$

Using the form for the nondimensional mean temperature gradient obtained by Businger *et al.* (1971),

$$-\frac{\kappa z}{T_*}\frac{dT}{dz} = \frac{1}{1.35}\left(1 - 9\frac{z}{L}\right)^{-1/2} \equiv \gamma, \tag{12}$$

and a similar form for the dimensionless mean humidity gradient,

$$-\frac{\kappa z}{q_*}\frac{dQ}{dz} = \gamma, \tag{13}$$

where $T_*(=\overline{w\theta}/u_*)$ and $q_*(=\overline{wq}/u_*)$ are friction temperature and friction humidity, respectively, Eqs. (10) and (11) may be written as

$$u_*T_* = \left(\frac{\kappa z\chi_\theta u_*}{2\gamma}\right)^{1/2}, \tag{14}$$

$$u_*q_* = \left(\frac{\kappa z\chi_q u_*}{2\gamma}\right)^{1/2}. \tag{15}$$

Estimates of χ_θ and χ_q have been determined from

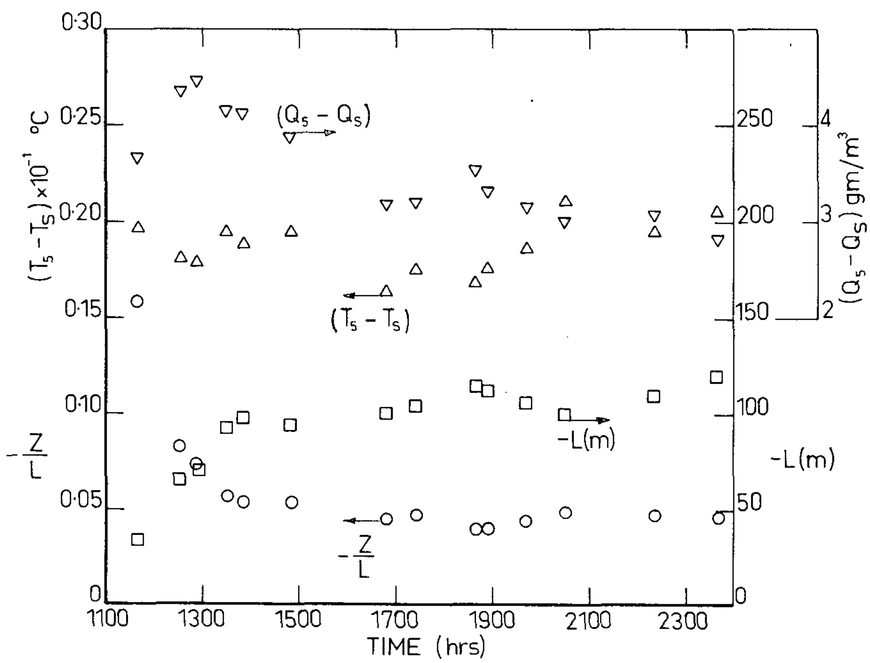

FIG. 7. Temperature and humidity difference between the 5 m height and the sea surface, Monin-Obukhov length and Richardson number.

inertial subrange estimates of spectral densities of θ and q, *viz.*,

$$\phi_\theta(k_1z) = \beta_\theta(\epsilon z)^{-1/3}\chi_\theta z(k_1z)^{-5/3}, \quad (16)$$

$$\phi_q(k_1z) = \beta_q(\epsilon z)^{-1/3}\chi_q z(k_1z)^{-1/3}, \quad (17)$$

where the constants β_θ and β_q are here assumed to be both equal to 0.5. There is some support in the literature for this numerical value in the case of β_q but the situation with respect to β_θ remains confused (see later in this section), even though direct estimates of χ_θ are more easily obtainable than those of χ_q.

Values of u_*, u_*T_* and u_*q_* obtained from Eqs. (7), (14) and (15) are also presented in Fig. 4 in the form of C_D^*, C_H^* and C_E^*. Values of C_D^* are generally higher than those of C_D during the initial part of the experiment [relatively low wind speeds (Fig. 3) and larger instability (Fig. 7)] but are in reasonable agreement with C_D for higher wind speed conditions. The average value of C_D^* is 1.22×10^{-3}, which is about 20% higher than the mean value of C_D. While C_E^* is in good agreement with C_E, the average value of C_H^* is almost twice as large as the mean value of C_H (Table 1).

Before these discrepancies can be usefully discussed, some attention should be given to possible errors that arise as a consequence of a misalignment to the wind direction of the w propeller and of its limited frequency response. The mean vertical velocity $\bar{w}$ obtained from the propeller signals was found to be in the range of $\pm 0.02\ U$. This corresponds to a tilt α of about $\pm 1°$ to the perceived wind direction. Using Deacon's (1968) result that the tilt error E in the covariance estimate $\overline{uw}$ is given by

$$E = \tfrac{1}{2}\sin 2\alpha[(\sigma_u/u_*)^2 - (\sigma_w/u_*)^2],$$

a maximum error of $\pm 15\%$ may be expected for $\overline{uw}$. For $\overline{w\theta}$ and $\overline{wq}$ the maximum error due to tilt is 20 and 15%, respectively. During the experimental period, $\bar{w}$ was of either sign so that the tilt error could be of either sign.

An estimate can also be made of the effect of limited upper frequency response (Fig. 6) of the u and w propellers on C_D, C_H and C_E. Fig. 6 indicates that the departure from inertial subrange behavior in the w spectrum occurs for $k_1z \gtrsim 2.5$. A similar departure occurs in the w spectrum given by Fichtl and Kumar (1974). Also, the expression obtained by Fichtl and Kumar for the time constant of propellers suggests that the departure between the true and measured spectra occurs at $k_1z \approx 2.5$ for the present wind velocity and turbulence levels. For $k_1z \lesssim 2.5$, the present w spectrum is similar in shape to that measured (using sonic anemometers) by Kaimal *et al.* (1972) for an equivalent value of z/L. The u propeller response was found to be slightly better, suggesting that the limiting factor in $\overline{uw}$ estimates is the w propeller response. Assuming that the high-frequency attenuation of the w propeller response is equivalent to applying a low-pass cutoff at $k_1z = 2.5$, the cospectra of uw and $w\theta$ measured by Kaimal *et al.* (1972) show that the present values of $\overline{uw}$ (and hence C_D) are likely to have been underestimated by 8–10%. It is estimated that similar effect on $\overline{w\theta}$ should represent about a 10% reduction in C_H. Note that unlike the tilt error, this error always results in reduced values of C_D, C_H and C_E, and thus accounts for almost half the discrepancy between C_D and C_D^*. However, the present discrepancy between C_E and C_E^* is likely to be slightly increased by the inadequate frequency response of the w propeller.

It should be noted that during those runs where u was obtained both from the hot wire and the propeller, the two estimates of $\overline{uw}$ were well within the overall standard deviation (Table 1) from each other. This applies also to estimates of $\overline{w\theta}$ using θ obtained from the cold wire and the thermistor. This further indicates that errors in $\overline{uw}$ and $\overline{w\theta}$ due to phase shift between the vertical propeller and the other sensors are well within the experimental accuracy.

The above considerations suggest that while discrepancies between C_D and C_D^* and also C_E and C_E^* could (at least in part) be traced to instrumentation problems and measuring errors, the discrepancy between C_H and C_H^* cannot be accounted for in the same manner. It should be noted that Champagne *et al.* (1977) obtained good agreement between C_E and C_E^* (with $\beta_q = 0.5$), and also between C_H and C_H^* (with direct measurement of χ_θ). The present discrepancy between C_H and C_H^* might imply a numerical value of β_θ that is significantly greater than 0.5 over water.

It should be mentioned that Pond *et al.* (1971) found that inertial dissipation estimates of sensible heat flux during BOMEX were a factor of 2 or more larger than direct flux estimates. Dunckel *et al.* (1974) found that they required a constant $\beta_\theta = 1.05$ to bring their estimates of C_H^* into agreement with C_H. Leavitt and Paulson (1975) suggested that a value of β_θ in excess of 0.8 was necessary to force agreement between C_H^* and estimates of the heat flux by the temperature profile method. Friehe and Schmitt (1976) suggested that the presence of "cold" spikes in Pond *et al.*'s (1971) temperature signals might be responsible for their large estimates of C_H. Schmitt *et al.* (1977) presented the hypothesis that temperature measurements over the ocean, particularly those exhibiting "cold spikes", are likely to be erroneous due to humidity sensitivity of spray-coated temperature sensors. This problem affects both C_H and C_H^* measurements, and appears to result in $C_H^* > C_H$. In the present experiments, however, there was little evidence of cold spikes in the temperature signals. It has been suggested that the well-established anisotropy of the streamwise temperature derivative (as indicated by the measured nonzero skewness) might account for the large values of χ_θ since the anisotropy may act in such a direction as to make the actual value of χ_θ smaller than the isotropic value (Champagne *et al.*, 1977), i.e.,

$$\chi_\theta < 6\alpha\overline{\left(\frac{\partial\theta}{\partial x}\right)^2},$$

where α is the thermal diffusivity of air. Recent measurements in a laboratory turbulent boundary layer of all three components of χ_θ by Sreenivasan *et al.* (1977) indicate that χ_θ should, in fact, be larger than the isotropic value by as much as 20%. The question of the anisotropy of atmospheric $\partial\theta/\partial x$ records is only one of several questions that need to be clarified before we can attempt to reconcile existing discrepancies between C_H and C_H^*. It therefore seems inappropriate at this stage to suggest yet another value for β_θ.

5. Discussion of bulk-transfer coefficients and other statistics of velocity and scalar fluctuations

Linear regressions for the drag coefficients C_D and C_D^* on wind speed U_5 are given in Table 1 together with values of the correlation coefficient and standard deviation of both intercept and slope of the regression lines. The increase in C_D with wind speed is not as pronounced or as significant as that obtained by Smith and Banke (1975) for a much larger data base obtained over water. Over a wind speed range which extends to values of U_{10} in excess of 20 m s^{-1}, Smith and Banke obtained

$$10^3 C_D = 0.63 + 0.066U_{10}(\pm 0.23),$$

where C_D is referred to the 10 m height. The present inertial dissipation values of C_D^* show a decrease with U_5, whereas Smith and Banke have commented that the "dissipation method" yields a smaller increase in the drag coefficient with wind speed. Opposite trends with respect to U_5 are also obtained for C_H and C_H^*, but the correlation coefficient is, in both cases, quite small. The trends of C_E and C_E^* with U_5, however, are in good agreement with each other, a decrease in the moisture flux transfer coefficient being predicted for increasing wind speeds. Other significant trends in Table 1 are the decrease of C_{uq} and the increase in $C_{q\theta}$ with U_5.

Values of the correlation coefficient R_{xy} over the

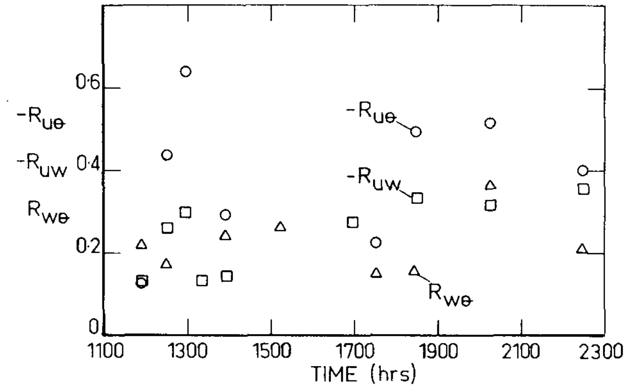

FIG. 8. Correlation coefficients of $u\theta$, uw and $w\theta$.

period of the experiment are shown in Figs. 8 and 9. The correlation coefficient R_{xy} is defined by

$$R_{xy} = \frac{\overline{xy}}{\sigma_x\sigma_y},$$

where x and y stand for either u, w, q or θ and σ_x, σ_{xy} are the standard deviations of x and y, respectively. Mean values of R_{xy} are shown in Table 2, together with linear regressions of R_{xy} on U_5. The coefficient R_{uw} is smaller in magnitude, but shows a more significant increase with wind speed, than that given by Smith and Banke. The temperature-humidity coefficient $R_{q\theta}$ has been found (Friehe *et al.*, 1975) to be important in determining statistics of optical refractive index fluctuations in the marine boundary layer. $R_{q\theta}$ and R_{uq} have the largest magnitudes of all the coefficients investigated here and both show a decrease with U_5, although this decrease is marginal for R_{uq}.

Values of rms velocity and scalar fluctuations normalized by friction velocity and friction scalar quantities are given in Fig. 10. The average value

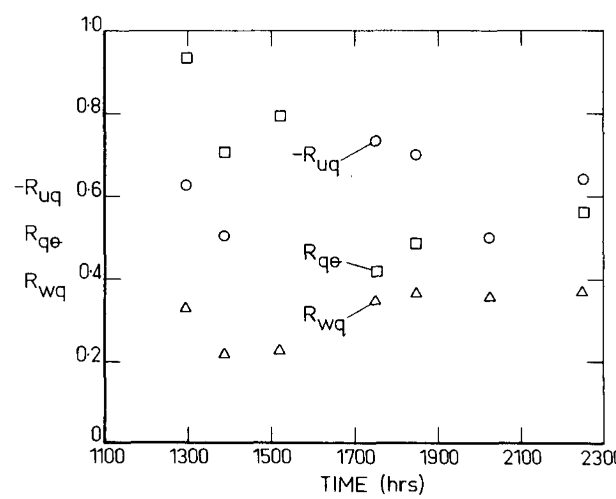

FIG. 9. Correlation coefficients of uq, $q\theta$ and wq.

TABLE 2. Regression lines of correlation coefficients and turbulence intensities on wind speed U_5.

Parameter	Mean ± standard deviation	Linear regression ± standard deviation	Correlation coefficient	Standard deviation Intercept	Standard deviation Slope
$-R_{uw}$	0.25 ± 0.08	$-0.006 + 0.031U_5$ (±0.079)	0.45	±0.180	±0.021
$-R_{u\theta}$	0.39 ± 0.16	$-0.017 + 0.050U_5$ (±0.164)	0.41	±0.375	±0.045
$R_{w\theta}$	0.23 ± 0.06	$+0.174 + 0.006U_5$ (±0.073)	0.12	±0.169	±0.020
$-R_{uq}$	0.62 ± 0.09	$+0.654 - 0.004U_5$ (±0.107)	0.02	±0.744	±0.084
R_{wq}	0.42 ± 0.18	$-1.525 + 0.221U_5$ (±0.166)	0.62	±1.116	±0.126
$R_{q\theta}$	0.65 ± 0.18	$2.892 - 0.256U_5$ (±0.139)	0.77	±0.945	±0.107
σ_u/u_*	3.30 ± 0.60	$3.656 - 0.043U_5$ (±0.670)	0.08	±1.516	±0.181
σ_w/u_*	1.36 ± 0.32	$2.914 - 0.187U_5$ (±0.269)	0.65	±0.608	±0.073
σ_θ/T_*	4.01 ± 1.42	$-0.743 + 0.579U_5$ (±1.408)	0.55	±3.304	±0.397
σ_q/q_*	2.68 ± 0.27	$3.045 - 0.01U_5$ (±0.320)	0.07	±2.149	±0.244

of σ_w/u_* (Table 2) is slightly larger than the (near neutral) value of 1.2 estimated by Monin and Yaglom (1971) for a large amount of available overland data. It is in fair agreement with the usually higher values (e.g. Pond *et al.*, 1971) of σ_w/u_* obtained over the ocean. The decrease of σ_w/u_* with U_5, shown in Table 2, is supported by the data of Smith and Banke. The average value of σ_u/u_* is somewhat larger than the values of 2.3 given by Monin and Yaglom (1971) (near neutral conditions overland) and 2.5 of Smith and Banke (1975) (overwater). The average value of σ_q/q_* is in plausible agreement with the trend of the data ($z/|L| > 0.1$) of Leavitt and Paulson (1975) but is rather small compared with the values obtained by Miyake and McBean (1970) at their lower values of $z/|L|$. The values of σ_θ/T_* of Fig. 10 show a large amount of scatter, with a mean value of 4.0 being twice as large as the value reported by Wyngaard (1973) for $z/|L| = 0.05$.

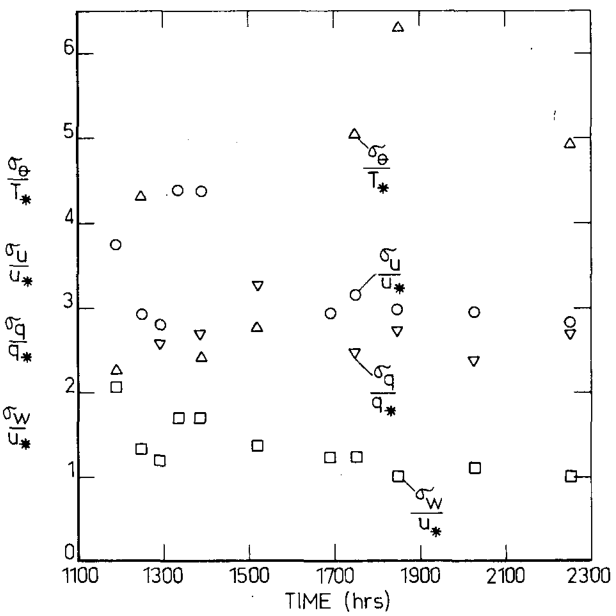

FIG. 10. Values of rms velocity and scalar fluctuations normalized by friction velocity and friction scalar quantities.

6. Summary of results

Turbulent fluxes of momentum, temperature and humidity were measured directly, under weakly unstable conditions, from a fixed oil-rig platform in Bass Strait. They were used to compute bulk transfer coefficients, which were also estimated by the inertial dissipation technique from measured inertial subrange spectral densities of velocity, temperature and humidity fluctuations, using a Kolmogorov constant of 0.5 for both velocity and scalar fluctuations. The two estimates were generally in reasonable agreement, except in the case of heat flux, for which the inertial dissipation technique yielded a value about twice as high as the direct method. This result points to the possibility that the Kolmogorov constant β_θ in Eq. (16) is larger than 0.5, and/or to inadequate assumptions made in the inertial dissipation technique. Within the range of measurements reported here, the variations of the bulk coefficients C_D and C_H with wind speed is not pronounced, and shows conflicting trends depending on the method used for estimating these coefficients. The moisture flux coefficient C_E is found to decrease with increasing wind speed, the result being independent of the method used to determine $\overline{wq}$. Bulk transfer coefficients associated with direct measurements of $\overline{uq}$, $\overline{u\theta}$ and $\overline{q\theta}$ are significantly larger than C_D, C_H or C_E. Of all the measured second-order correlation coefficients, those between humidity and temperature fluctuations and between humidity and streamwise velocity fluctuations have the largest magnitude, and decrease slowly with wind speed.

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