Wind-Wave-Induced Disturbances in the Marine Surface Layer

R. A. Antonia and A. J. Chambers

Department of Mechanical Engineering, University of Newcastle, New South Wales 2308, Australia

(Manuscript received 2 July 1979, in final form 4 December 1979)

ABSTRACT

Measurements of velocity and scalar fluctuations were made using a stable platform in Bass Strait at a height of ~5 m above the mean surface of the ocean. These measurements were obtained over a period of time where the wind velocity increased steadily before reaching an approximately constant value that was maintained for a duration of ~10 h. During the initial period of the experiment, induced wave fluctuations, centered about the dominant wave frequency, are observed on spectra of the longitudinal u and vertical w velocity fluctuations, and on the uw cross spectrum. The cross spectrum indicates a relatively important transfer of momentum in the direction sea to air, at the wave frequency. No measurable wave influence is detected on either temperature θ or humidity q spectra, but $w\theta$ and wq cross spectra are either negligible or show a change of sign at the wave frequency. Although the effect on w persists, wave-induced disturbances on the cross spectra are small enough to be neglected when the ratio of wave phase speed to friction velocity becomes smaller than ~40.

1. Introduction

The influence of waves on the turbulent surface layer needs to be carefully assessed before accurate estimates can be made of the sea surface stress and mass, temperature and humidity fluxes. Laboratory and atmospheric studies have firmly established that wave-induced perturbations can be recognized in measurements of the longitudinal or streamwise u velocity and especially of the vertical w fluctuation in the turbulent flow above the waves. Most of the experimental evidence (e.g., Kitaigorodskii, 1973) has been in the form of power spectral densities of w and u which reveal a distinct peak at the frequency at which the spectral density of the sea surface wave displacement is maximum or at the main frequency of mechanically generated waves in air-water tunnels. The uw cospectrum can also exhibit a peak at the dominant wave frequency corresponding to peaks in the w and u spectra.

The laboratory investigation by Lai and Shemdin (1971) of air turbulence above mechanically generated waves has indicated that wave-induced perturbations decay with height above the wave surface and has shown that the rate of decay depends on the relative magnitude of wind and wave speeds. The uw cospectrum was found to exhibit a change of sign near the wave frequency, indicative of a transfer of momentum in the direction water to air, when the local air velocity was less than the phase velocity C corresponding to the dominant wave frequency. No such effect was noted when the wind velocity exceeded C although a peak in the cospectrum remained prominent at the wave frequency, thus suggesting an enhancement of overall momentum transfer from wind to water. More specifically, Lai and Shemdin concluded that the magnitude and direction of the wave-induced stress above the waves depended critically on $(U_1 - C)$ where U_1 can be taken as the average wind speed at the average wave crest level. The direction of stress was stated to be upward when $U_1 < C$. No significant peak in the uw cospectrum can be observed in McIntosh *et al.*'s (1975) observations over wind waves and mechanically generated waves in an air-water tunnel, although distinct peaks, at the dominant wave frequency, are noticeable in spectra of u and w. The data of Takeuchi *et al.* (1977), obtained in the same facility as that used by McIntosh *et al.*, indicate a significant negative spike in the uw cospectrum at the wave frequency,[1] although no peak is found on the w spectrum. The absence of this peak was attributed to the near equality between the local wind velocity and the phase velocity C.

Benilov et al. (1974) reported measurements from a stable platform in the Caspian Sea (see Zubkovskiy *et al.*, 1974) which indicate a positive excursion in the uw cospectrum at the main wave frequency.[2] The wave-induced contribution to

[1] A significant positive spike in the cospectrum occurred at the second harmonic of the wave frequency.

[2] Davidson and Frank (1973) observed that the wave-related momentum transfer could change sign with height above the surface of a lake. Close to the surface, the momentum transfer at the wave frequency could be either enhanced or reduced.

0022-3670/80/040611-12$07.00

$-\overline{uw}$ was determined using a linear filtering technique (see also Benilov and Filiushkin, 1970; and Benilov and Zaslavskiy, 1974), based on a correlation analysis of turbulence and sea waves. Benilov *et al.* found that the wave-induced contribution represented a reduction of 15% of the Reynolds shear stress. This reduction, which corresponded to increases of 29 and 18% in the variances σ_w^2 and σ_u^2 of w and u, respectively, was found over attenuated wind-generated waves for which the ratio C/U_* (U_* is the friction velocity $= -\overline{uw}^{1/2}$) is fairly large (85 in their case). This ratio was suggested (Kitaigorodskii, 1973) to be a relevant parameter characterizing the dynamic interaction between sea and wind. For developing wind-generated waves ($C/U_* = 20$), Benilov *et al.* (1974) found that the wave contribution to $-\overline{uw}$ was almost negligible while contributions to σ_w^2 and σ_u^2 were only about half those obtained for swell conditions.

It should be noted that the wave influence is not confined to u and w fluctuations. Elder *et al.* (1970) observed significant energy concentration at the wave frequency in spectra of v, the transverse velocity fluctuation. Elder *et al.* suggested that this observation supported the three-dimensional nature of the disturbance due to the wave field. However, it should be mentioned that although this peak in the v spectrum was accompanied by a peak in the w spectrum, no peak was observed in the u spectrum. Not surprisingly perhaps, the wave-induced contribution to pressure fluctuations has been found to be significant (e.g., Elliott, 1972b). Elliott observed a large hump in pressure spectra at the wave frequency with the amplitude and rate of vertical decay of this hump increasing and decreasing, respectively, as the mean wind speed increases.

Whereas the influence of waves on the velocity and pressure field has received some attention, the possible influence on scalar fluctuations, such as temperature θ and humidity q, seems to have been less well-documented. Before we briefly discuss the available experimental evidence, it is worth restating Benilov *et al.*'s suggestion that the amplitude of the wave-induced perturbation of any quantity depends, not only on distance from the surface height of the wind waves or relative magnitude between wind and wave phase velocities[3] but also on the vertical gradient of mean velocity (or presumably temperature or humidity). While induced velocity fluctuations could still be present under exactly neutral conditions, there is no obvious mechanism to explain the possibility of generation of temperature disturbances as a result of wave motion. Volkov (1969) (also, Kitaigorodskii, 1973) presented temperature spectra, measured from a ship at a height z of 2 m above the ocean surface, which indicate a small but distinct peak at the main wave frequency under slightly unstable stratification ($z/L \approx -0.025$, where L is the Monin-Obukhov length). Kitaigorodskii estimated that the wave-induced contribution to the temperature spectrum is of the same order of magnitude as the wave-induced temperature, as inferred from his relation (3.37). Benilov *et al.* (1974) noticed no significant peak in the humidity spectrum at a height $z = 2.45$ m above a sea surface characterized by $C/U_* \approx 85$. At $C/U_* = 20$, the data at $z = 4.25$ m exhibit a small bump or at least a distinguishable change in spectral slope at the main wave frequency. However, the calculations by these authors showed that, for both values of C/U_*, the contribution from wave-induced fluctuations to the variance of q is only ~4%. McIntosh *et al.* (1975) presented several temperature spectra for various heights and wind speeds above wind and mechanically generated waves. Only one of these spectra, that for the mechanical waves, seemed to exhibit a peak at the wave frequency. But the experimental spectral accuracy at that frequency appears to be insufficient to claim a significant wave disturbance. While the wave influence on θ or q spectra may be small, the effect on $w\theta$ and wq cospectra may be more significant. The wq cospectrum of Benilov *et al.* (1974) exhibited a significant reduction at the main wave frequency. One of Young *et al.*'s (1973) $w\theta$ cospectra (run V24) revealed a pronounced reduction at the dominant frequency of the wind-generated waves but no significant peak was noticeable in the θ spectrum.

Deviations of the behavior of parameters such as σ_u/U_*, σ_w/U_* and $\sigma_\theta/|T_*|$ ($T_* \equiv \overline{w\theta}/U_*$ is the friction temperature) from that prescribed by Monin-Obukhov similarity have been determined by Volkov (1969) as a function of the ratio C/U_*. The ratios σ_u/U_* and σ_w/U_* were found to increase significantly with C/U_* when this latter parameter exceeded a value of ~25. This increase (also observed by De Leonibus, 1971, and Davidson, 1974) is brought about by the decrease in $-\overline{uw}$ as C/U_* increases. In the case of temperature, $\sigma_\theta/|T_*|$ remained constant for $C/U_* \geqslant 25$ but increased significantly for $C/U_* \leqslant 25$. Davidson confirmed that a value of C/U_* near 25 is associated with minimal wind-wave coupling influence but could not support Volkov's increase in $\sigma_\theta/|T_*|$ at low values of C/U_* in view of lack of data in this range of C/U_*. De Leonibus (1971) presented some evi-

[3] It should also be noted that Kondo *et al.* (1972) found that when the wind blows in the opposite direction from that of the wave propagation, the wind fluctuation is in phase with respect to the wave motion and the amplitude of wave-induced wind component is relatively large.

dence of the dependence of the drag coefficient on dimensionless wave phase velocity and dimensionless wave height.

It is possible that both direct covariance and inertial dissipation techniques for estimating the sea surface stress can be in error if the velocity fluctuations have a significant wave-induced component. Stress estimates from the dissipation technique are generally higher than the directly measured values (e.g., Schmitt *et al.*, 1978b). However, it is not clear that the discrepancy between these two estimates is entirely due to the suggestion by Schmitt *et al.* (1978b) that the anisotropy of the velocity field, as caused by the interaction between surface waves and the turbulence, invalidates the application of the inertial dissipation technique.

Benilov *et al.* (1974) have indicated that on-site experimental studies to determine the role of wave-induced perturbations of the turbulent flow are rare. Zubkovskiy *et al.* (1974) pointed out the advantages of using a stable platform for such an experimental investigation and the data from their Caspian Sea platform constitute an important body of knowledge on the influence of waves on turbulence. The aim of this paper is to place on record some simultaneous measurements of u, w, θ and q obtained from a stable oil-rig platform in Bass Strait. These measurements were made over a period of time where the wind velocity increased steadily before reaching a value that remained approximately constant over a period of 10 h. The measured spectra and cospectra of u and w are significantly affected at or near the main wave frequency. These spectral results and those associated with temperature and humidity fluctuations are compared, when possible, with the data of Benilov *et al.* (1974) and others.

2. Experimental arrangement

Measurements of u, w, θ and q were made from Kingfish B, 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, Australia. The instruments for recording the above signals were mounted at a height z of ~5 m above the mean water level (on a vertical pipe), supported at the end of a horizontal boom fastened to one of the western platform legs. The boom was of sufficient length (15 m) to enable the measurements to be made clear of the disturbance of the platform. The response of the boom to the wave and wind fields was negligible. Measurements of the platform's acceleration obtained by ESSO-BHP indicated undetectable platform movement during conditions similar to those of the present experiment. A more detailed description of the instruments and their relative spatial location may be found in Antonia *et al.* (1978b). The horizontal velocity fluctuation u was obtained with a hot wire (5 μm diameter, 0.8 mm long) operated by a DISA 55M01, constant-temperature anemometer. The vertical velocity fluctuation w was obtained using a Gill propeller array which also provided another measurement of u. 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 is discussed in Antonia *et al.* (1978b). The tilt error (w propeller) was estimated to be ±15, ±20 and ±15%, respectively, for $\overline{uw}$, $\overline{w\theta}$ and $\overline{wq}$. Temperature θ was measured with a cold wire (0.6 μm diameter, 0.8 mm long) 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. Low-frequency temperature fluctuations were also obtained by a thermistor. The humidity fluctuation q was obtained using a Lyman-alpha humidiometer. Neither the hot wire anemometer nor the Lyman-alpha humidiometer was linearized. The hot and cold wires were mounted, respectively, at the levels of the w propeller and the Lyman-alpha humidiometer, at a lateral separation of ~40 cm. Schmitt *et al.* (1978a) suggested that temperature measurements over the ocean exhibiting "cold spikes" are likely to be in error due to the humidity sensitivity of the spray-coated temperature sensors. These would lead to errors in $\overline{w\theta}$. In the present experiment there was little evidence of cold spikes in the temperature signal. Schacher and Fairall (1976) also indicate that salt deposits are rarely of sufficient thickness to affect the calibration of hot and cold wires.

Fairall and Schacher (1977) found, for non-linearized hot wires at constant mean velocity, that the slope $s(=u/e$, where e is the fluctuating bridge voltage) increased by 30% due to salt buildup for a group of wires used daily for several days on a cruise. As s increased, the frequency response of the hot-wire anemometer system decreased. However, they expect the system frequency response to be adequate for inertial dissipation (and therefore direct) estimates of the surface shear stress if s is allowed to increase to 60% provided *in situ* wire calibrations are obtained. For the present experiment, the hot wires were exposed for ~18 h and changes in the hot-wire calibration constants were less than 10%.

The mean water level was continuously monitored with a resistance wave gage, consisting of a non-conducting tube, ~2.5 cm in diameter, with a conducting wire wrapped spirally around it. This wave gage, which provided an output proportional to the instantaneous surface displacement η, was suspended from the boom at a location 2 m downwind

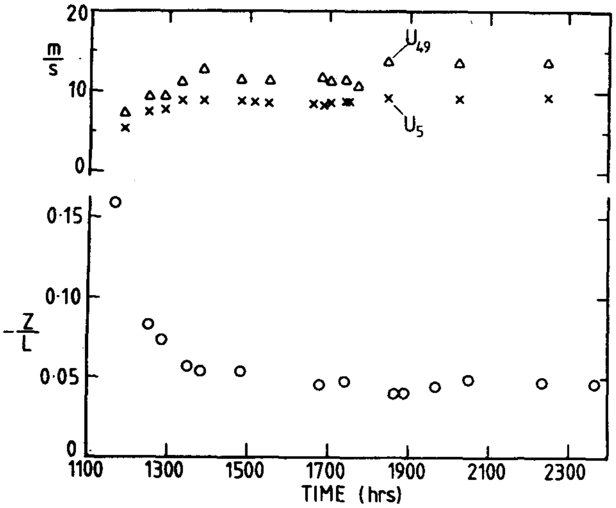

FIG. 1. Wind speed and z/L variation during observational period.

from the vertical mast to reduce the likelihood of wave-induced motion of the instrument mast.

Voltages proportional to u, w, q and θ fluctuations were usually recorded on a four-channel Hewlett-Packard 3960 FM tape recorder. For a number of runs, θ (or q) was replaced by the surface displacement η. The recording speed was 24 mm s^{-1} (−3 dB point of tape recorder 375 Hz). The tapes were played back in the laboratory at a speed of 16 times the speed recording and digitized at a sampling frequency of 64 Hz (real time equivalent is 4 Hz) on a DEC PDP 11/20 computer. Prior to digitization, the signals were low-pass filtered with the −3 dB cutoff frequency set at half the sampling frequency. Data computed from these signals included mean values of these quantities and of their products as well as probability density functions, spectra and cross spectra.

The variation with time of the wind velocity U_5 at 5 m and U_{49} at 49 m, as measured by a cup and propeller anemometer, respectively, is shown in Fig. 1 (typically 30 min averages of the cup output). It is noticeable that following an initial increase, U_5 remains approximately constant, equal to ~9.2 m s^{-1}, over the period 1300–2300 LST. Also, shown in Fig. 1 is the variation of L [for details of this computation see Antonia *et al.* (1978b)] over the same period.

3. Experimental results and discussion

Spectra of the wave surface elevation η were obtained at a number of different times throughout the observational period. The frequency n_0 of the main peak in the spectrum was found to be approximately constant, equal to ~0.13 Hz. It follows that the phase velocity C associated with the energy containing waves was also approximately constant, equal to about 12 m s^{-1} ($C = g2\pi n_0$). The error, in the estimate of C, resulting from a finite water depth was negligible as the water depth at the location of the platform is ~70 m. As the mean wind velocity at 5 m is always less than C (Fig. 1), all measurements presented in this section were obtained well below the critical layer height. The horizontal velocity measured at 49 m was less than C during 1100–1300, approximately equal to C during 1300–1800 and greater than C in the period 1800–2400 (all times local standard). Thus, for the major portion of the experiment, the critical layer height was close to or greater than 49 m. It should be noted that the ratio C/U_* given in Table 1 does not vary significantly over the experimental period 1300–2300. Larger values of C/U_* were encountered over the period 1100–1300 as a result of lower values of U_*, in the range 0.15–0.25 m s^{-1}.

Spectra of u, w and θ for run 1 (see Table 1) are shown in Fig. 2 in the form ϕ_α vs k_1z. The spectral density ϕ_α (where α stands for u, w, θ, q or η) has been normalized such that

$$\int_0^\infty \phi_\alpha d(k_1z) = 1,$$

where the wavenumber $k_1 = 2\pi n/U$. Fluctuations u, θ presented in Figs. 2 and 3 were obtained using the hot wire and cold wire, respectively. Also, shown in Figs. 2 and 3 is the η spectrum. An equilibrium form for this spectrum, as indicated by a −5 slope, exists for $k_1z > 2$. The main peak occurs near $k_1z = 1$. At or near this value, there are pronounced peaks in ϕ_u and especially ϕ_w. There is no noticeable bump in ϕ_θ, the variations in the spec-

TABLE 1. Mean and turbulence parameters for runs 1–4 (BASS V).

Run	Start (LST)	Duration (min)	U (m s^{-1})	U_*†† (m s^{-1})	$-z/L$	$\frac{C}{U_*}$	$C_D \times 10^3$†	$C_D \times 10^3$††	$\frac{\sigma_u}{U_*}$	$\frac{\sigma_w}{U_*}$	$\frac{\sigma_\theta}{T_*}$
1	1137	32	5.31	0.15	0.13	81.6	1.42	0.77	3.77	2.06	2.2
2	1217	26	7.33	0.26	0.08	46.0	1.23	1.27	2.94	1.32	4.3
3	1759	58	9.13	0.31	0.05	38.6	1.06	1.16	2.90	1.0	4.6
4	2200	60	9.38	0.37	0.05	32.6	1.66	1.54	2.83	0.98	4.9

† Inertial dissipation values.
†† Direct values.

trum near $k_1z \approx 1$ being of the same order as the scatter exhibited at lower frequencies. Also shown in Figs. 2 and 3 are the products $(k_1z)^{5/3}\phi_\alpha$, with $\alpha = \theta$ or u, to identify the inertial subrange region for temperature and velocity fluctuations. It is clear that while $(k_1z)^{5/3}\phi_u$ exhibits a constant value (~0.12) for $k_1z > 2$, the peak in ϕ_u associated with n_0 interferes severely with the extent and level of the inertial subrange plateau. The inertial subrange, in the case of temperature, does not start before k_1z is ~2.5 and, even if a bump in ϕ_θ had existed, interference with the inertial subrange[4] plateau would have been significantly smaller than in the case of velocity. It should be noted that the extent of the inertial subrange does not terminate at $k_1z = 13$; redigitization of the same u and θ records indicated inertial subrange regions extending to at least 25 for u and 40 for θ (and q). The quantity $(k_1z)^{5/3}\phi_w$ has not been plotted in Figs. 2 and 3 in view of the relatively poor frequency response of the w propeller (cf. Antonia *et al.*, 1978b).

Spectra (Fig. 3) of u and θ for run 2 (Table 1) are similar to the corresponding spectra of Fig. 2 except that the perturbation of ϕ_u due to the waves is significantly smaller than that in Fig. 2 and the start of the temperature −5/3 subrange is slightly delayed in Fig. 3. The wave-induced perturbation of ϕ_w (Fig. 3) is significant, although smaller than that of Fig. 2. The behavior of the w spectrum to the left of the bump[5] in Fig. 2 is somewhat different from that in Fig. 3. The relatively constant spectral density and sharp dip immediately prior to the conspicuous rise in Fig. 2 are absent in the low-frequency end of ϕ_w in Fig. 3. Approximate values of contributions of wave-induced fluctuations to σ_u^2 and σ_w^2 were ob-

[4] Note that the u, w and q spectra of Benilov *et al.* (1974) do not show a convincing −5/3 inertial subrange.

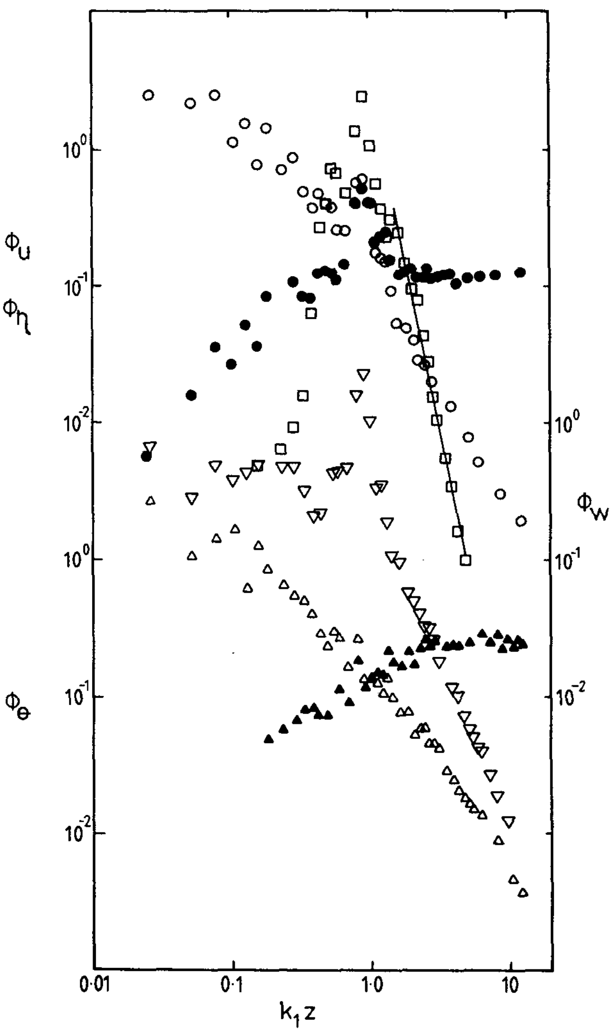

FIG. 2. Spectral densities of longitudinal, vertical velocity, temperature and wave height fluctuations, respectively, for run 1: (○) u, (▽) w, (△) θ, (□) η. Closed symbols represent $(k_1z)^{5/3}\phi_\alpha$. Straight line has slope of −5.

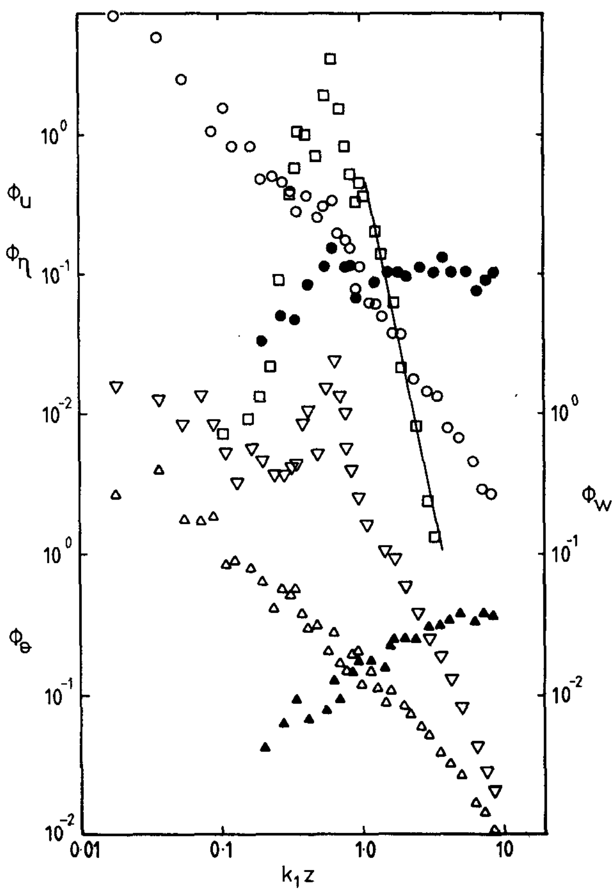

FIG. 3. Spectral densities of longitudinal, vertical velocity, temperature and wave height fluctuations for run 2. Symbols as in Fig. 2.

[5] It is difficult to identify a −5 region on the w spectrum immediately to the right of the bump, as mentioned by Zubkovskiy *et al.* (1974).

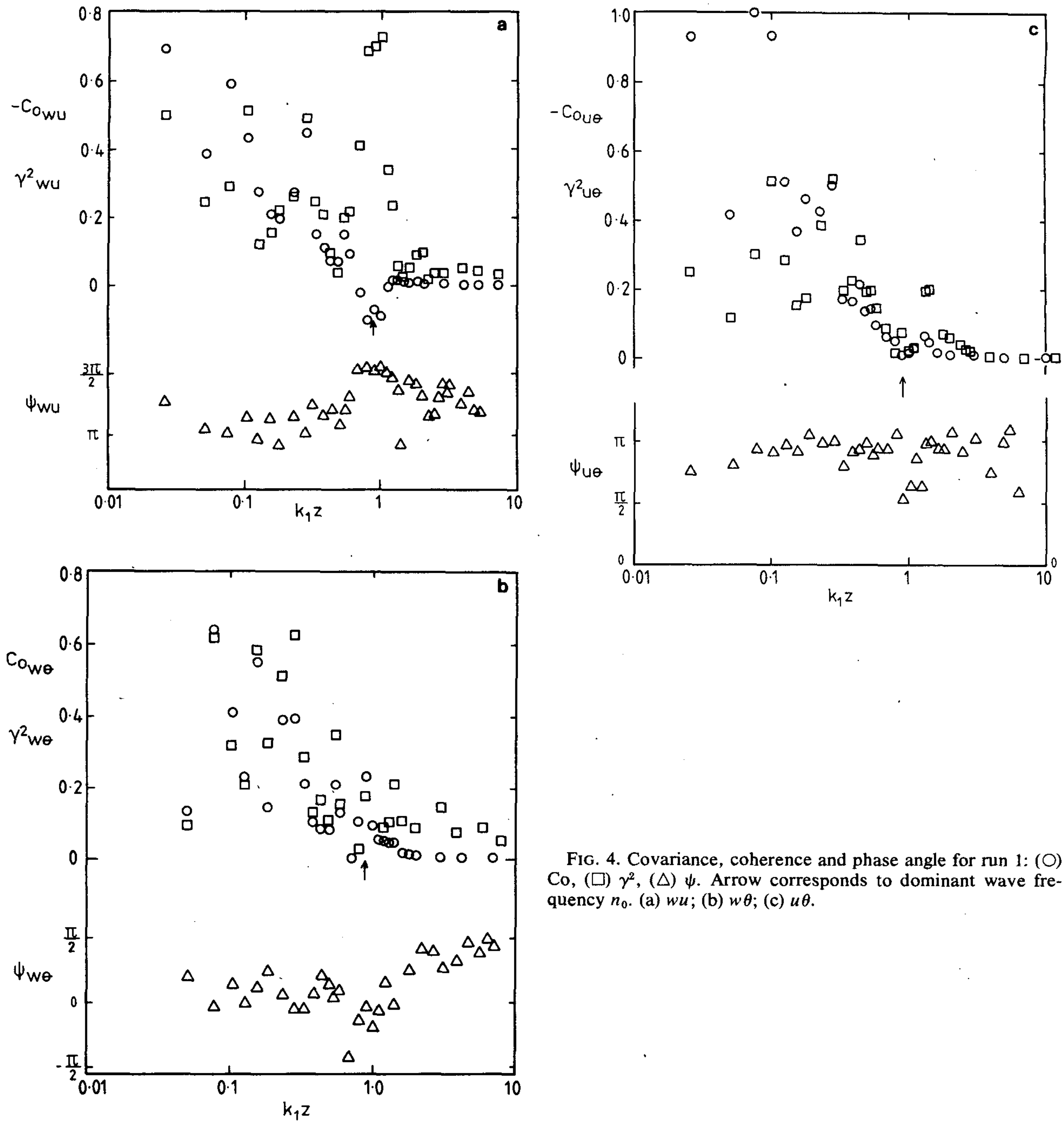

FIG. 4. Covariance, coherence and phase angle for run 1: (○) Co, (□) γ^2, (△) ψ. Arrow corresponds to dominant wave frequency n_0. (a) wu; (b) $w\theta$; (c) $u\theta$.

tained by estimating the spectral area enclosed by the bump near $k_1z \approx 1$ and a faired-in (by eye) interpolated curve joining the undisturbed spectral densities on either side of the bump. For Fig. 2 the contributions to the measured $\sigma_u{}^2$ and $\sigma_w{}^2$ are 13.6 and 52.5%, respectively. For Fig. 3 the corresponding values are 3.2 and 48.2%. Although these estimates are only crude, they are sufficiently representative of the wave-induced motion to allow comparison with the data of Benilov *et al.* (1974). These authors found that their linear filtering technique isolated wave noise contributions to spectra and cospectra not only near n_0 but also at higher and lower frequencies. The main wave noise contribution, however, was associated with n_0. For attenuated waves ($C/U_* \approx 85$), Benilov *et al.* found that the contributions to $\sigma_u{}^2$ and $\sigma_w{}^2$ were 18 and 29%, respectively. While the wave noise contribution to $\sigma_u{}^2$ for run 1 is not very different to that of Benilov *et al.*, the present contribution to $\sigma_w{}^2$ is much larger. Although the ratio C/U_* for run 1 is relatively large ($\sim$82),[6] perhaps it should be noted that it was obtained for a history of wind speed which is different from that applicable to Benilov *et al.*'s experiment. In the latter, the mean velocity at $z = 10$ m was large (13 m s^{-1}) in the period preceding the experiment and equal to $\sim$6 m s^{-1} during the experimental run.

Cross spectra between the various quantities of interest were computed for runs 1 and 2 and are shown in Figs. 4 and 5. The cross spectrum is given by Co + iQ, where Co and Q are the cospectrum and quadrature spectrum respectively. The normalization of the figures is such that

$$\int_0^\infty \mathrm{Co}\,d(k_1 z) = R(0),$$

the cross-correlation coefficient at zero time delay τ, where $R_{\alpha\beta}(\tau) = \overline{\alpha(t)\beta(t-\tau)}/\sigma_\alpha\sigma_\beta$. Also, shown in these figures are the phase ψ (=tan^{-1} Co/Q) between quantities α and β and the coherence γ^2, defined in the usual fashion,

$$\gamma^2 = \frac{\mathrm{Co}^2 + \mathrm{Q}^2}{\phi_\alpha \phi_\beta},$$

where ϕ_α and ϕ_β are the spectral densities of α and β, respectively. Conspicuous in Fig. 4a is the change in sign of Co_{uw} near $k_1 z \approx 1$. The positive (and relatively large) values of Co_{uw} near the main wave frequency indicate a transfer of momentum in the direction water to air, as observed by Benilov *et al.* (1974) over the ocean ($C/U_* \approx 85$) and Lai and Shemdin (1971) over mechanically generated waves when the air velocity was smaller than the wave phase velocity. The coherence γ^2 near $k_1 z \approx 1$ exceeds 0.7 (significantly higher than values measured at lower frequencies where a reasonably high correlation is expected between u and w). Lai and Shemdin report a value of γ^2 at their wave frequency in excess of 0.8,[7] whereas, Benilov *et al.* somewhat inexplicably, state that γ_{uw}^2 is close to zero near n_0, although they report values of $\gamma_{u\eta}^2$ and $\gamma_{w\eta}^2$ equal to 0.8 and 0.6, respectively. The phase ψ (Fig. 4a) is such that, at or near n_0, w leads u by nearly $3\pi/2$ or, alternatively, u leads w by $\pi/2$. This is essentially in agreement with the observation of Lai and Shemdin who also find that ψ, at $n = n_0$, decreases with increasing height above the surface. In Fig. 5a ($C/U_* \approx 46$), Co_{uw} does not change sign but is nearly zero near the wave frequency. Nevertheless, the coherence γ^2 is still large (0.6) and u leads w by $\pi/2$. Wave-induced contributions to $\overline{uw}$ were estimated from the area enclosed by the depression in Co_{uw}, near $k_1 z \approx 1$, and a linear interpolation (by eye) joining the "undisturbed" Co_{uw} on either side of the depression ($0.5 < k_1 z < 1.2$, Fig. 4a; $0.45 < k_1 z < 0.65$, Fig. 5a). The contribution represents a reduction in $-\overline{uw}$ of 24.9 and 5.4% for runs 1 and 2, respectively.

The $w\theta$ and $u\theta$ cospectra (Figs. 4b, 4c; 5b, 5c) indicate that, near the main wave frequency, $\mathrm{Co}_{u\theta}$ is negligible, while $\mathrm{Co}_{w\theta}$ does, in fact, become negative, suggesting a transfer of heat from the air toward the slightly warmer sea surface. Coherences near n_0 ($k_1 z \approx 0.9$, Fig. 4; 0.6, Fig. 5) are nearly zero, while θ leads w by nearly $\pi/2$, in reasonable agreement with the data of Benilov *et al.* It should finally be noted that the main features of the data in Figs. 2–5 have, in fact, been verified with the use of different instruments. Spectra and cross spectra obtained with the use of a u propeller (instead of the hot wire) and a thermistor (instead of the cold wire) are in close agreement with those in Figs. 2–5. It must be noted that the behavior of correlation coefficients $R_{wu}(\tau)$ and $R_{w\theta}(\tau)$ as a function of time delay τ is such that maximum values for these coefficients occur at positive values of τ so that θ and u lead w for the overall time record.

Spectra of q and cross spectra between $q\theta$ and wq are presented in Fig. 6. The quantities of interest were measured prior to run 1 (1100–1115 LST) and θ was obtained from a thermistor located beside the Lyman-alpha humidiometer. There is no apparent bump in ϕ_q or $\mathrm{Co}_{\theta q}$ near $k_1 z \approx 1$, which corresponds to the peak in the wave spectrum (Fig. 6a). The inertial subrange of q appears well established at $k_1 z \approx 3$ and $\mathrm{Co}_{\theta q}$ exhibits a $-5/3$ inertial subrange behavior for $k_1 z > 3$. The behavior of the wq cross spectrum is qualitatively similar to $\mathrm{Co}_{w\theta}$ for runs 1 and 2. Both Co_{wq} and γ_{wq}^2 tend to zero and q appears to lead w by about $\pi/2$ at $k_1 z \approx 1$ (Fig. 6b). There is no apparent wave effect on the $\mathrm{Co}_{\theta q}$, with $\gamma_{\theta q} \approx 0.9$ and $\psi_{\theta q} \approx 0$ over the major part of the $k_1 z$ range (Fig. 6c). Behavior of the peak values in $R_{wq}(\tau)$ and $R_{\theta q}(\tau)$ is such that, for the whole record, q leads w and is in phase with θ.

For runs 3 and 4, typical of data in the 10 h period in which U_5 and C/U_* were practically constant, no significant effect was noted on cospectra or coherences associated with quantities u, w, q at or near the wave frequency, although ϕ_w continued to show a bump at this frequency. In particular, ϕ_θ and ϕ_q as well as the cospectrum $\mathrm{Co}_{\theta q}$ remained unaffected throughout this period. Values of σ_u/U_* and σ_w/U_* shown in Table 1 indicate a significant increase for runs 1 and 2 for which C/U_* exceeds 40. This increase is a result of the increase in σ_u and σ_w and corresponding decrease in shear stress rela-

[6] This value was calculated using U_* inferred from the direct estimate of $\overline{uw}$. If the inertial dissipation value for U_* was used (see Table 1) C/U_* would be about 57.

[7] A similar value was obtained by Kononkova *et al.* (1973) at a height of 15 cm above the water surface in an air-water tunnel.

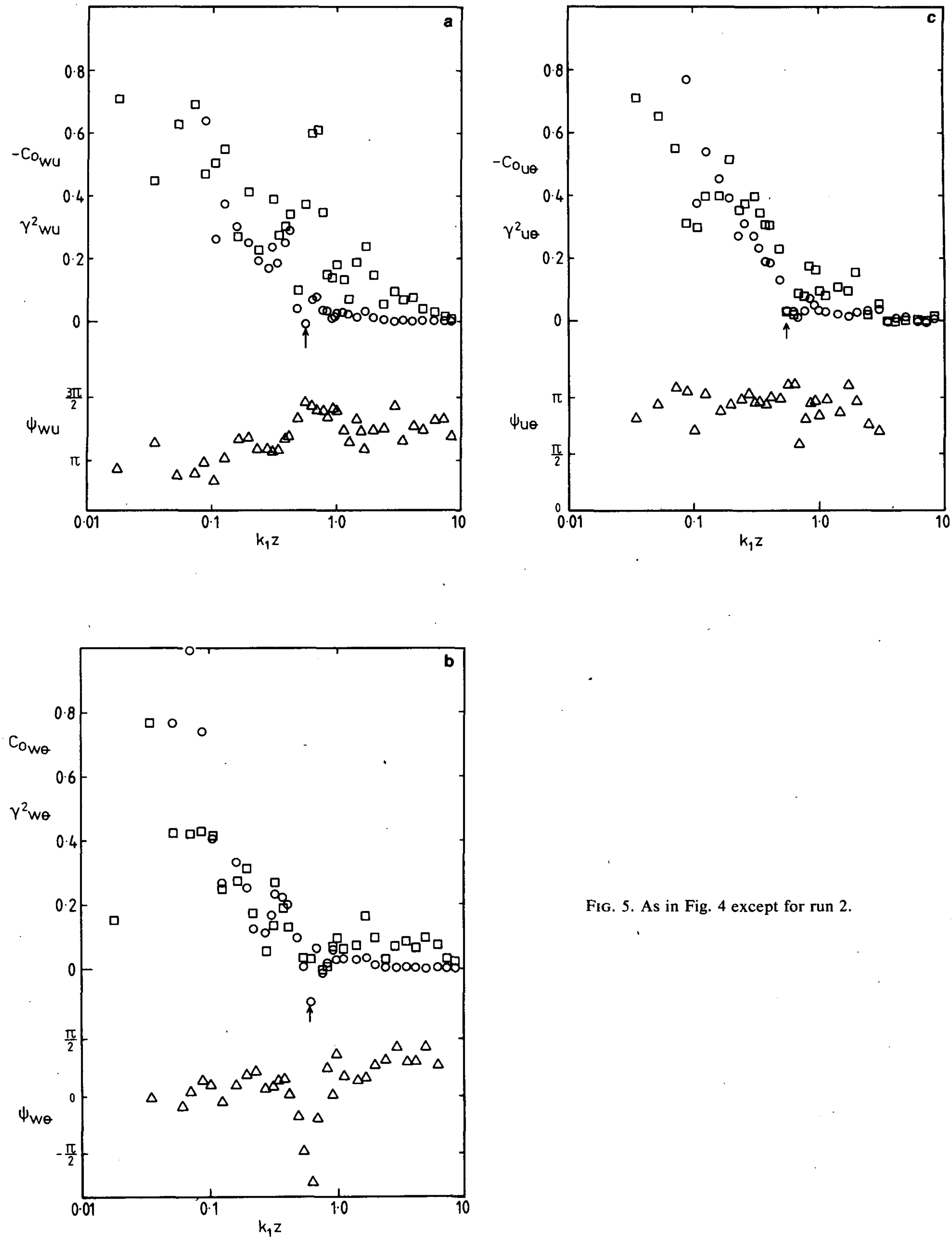

FIG. 5. As in Fig. 4 except for run 2.

tive to the undisturbed values of runs 3 and 4. The increase cannot be explained by the difference in z/L of runs 1–4 but is in qualitative agreement with the dependence, established by Volkov (1969), of $\sigma_u U_*$ and σ_w/U_* as a function of C/U_*. Values of σ_u/U_* and $\sigma_w U_*$ for $C/U_* < 40$ are in reasonable agreement with those measured over land for a similar value of z/L. The relatively small value of $\sigma_\theta T_*$ for run 1 can be almost entirely ascribed to the reduction in U_* for that run. The correlation coefficient $R_{uw}(0)$ is equal to 0.13 and 0.26 for runs 1 and 2 respectively. Although the variation of $R_{uw}(0)$ with C/U_* is in qualitative agreement with that of Volkov (1970), the present values of $R_{uw}(0)$ are significantly larger than those of Volkov at corresponding values of C/U_*. The present values of $R_{w\theta}(0)$, which can be inferred from Table 1, do not vary significantly from 0.2 over the range of C/U_* covered in the table. They are in reasonable agreement with the measurements of Volkov.

Skewness S_α ($\equiv\overline{\alpha^3}/\sigma_\alpha{}^3$) and flatness factors F_α ($\equiv\overline{\alpha^4}/\sigma_\alpha{}^4$) of u, w and θ are shown in Table 2. Values of S and F for runs 1 and 2, where the wave influence is noticeable in both ϕ_u and ϕ_w, are not significantly different from those of runs 3 and 4, where the only perceptible disturbance by the wave is in ϕ_w. This seems to be in agreement with the observation by Takeuchi *et al.* (1977) that probability density functions and other mean statistics measured above mechanically generated waves are in good agreement with those measured in a boundary layer over a smooth wall. These authors commented that the influence of the waves on the air may be obscured by ordinary time averaging. The numerical values of S_u, S_w, F_u, F_w in Table 2 are in fair agreement with those tabulated by Takeuchi *et al.* The present values of skewness and flatness factors of $(uw - \overline{uw})$ are a little higher than those of Takeuchi *et al.* It should also be pointed out (e.g., Sreenivasan *et al.*, 1978) that these values of $S_{uw-\overline{uw}}$ and $F_{uw-\overline{uw}}$ do not differ significantly from those corresponding to the probability density function of the product of two Gaussian variables and cannot simply be attributed, as stated by Takeuchi *et al.*, to the intermittent nature of the Reynolds stress production.

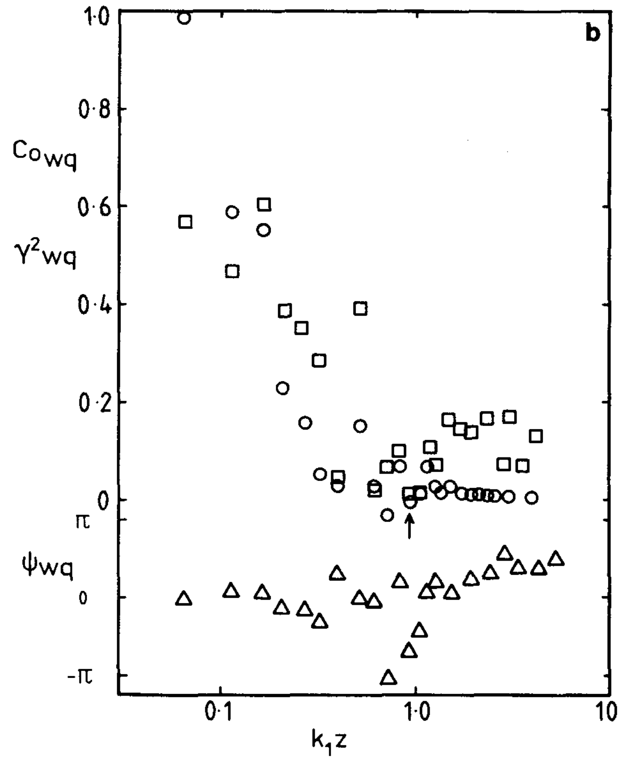

FIG. 6b. Covariance, coherence and phase angle for wq. Measurement prior to run 1. Symbols as in Fig. 4.

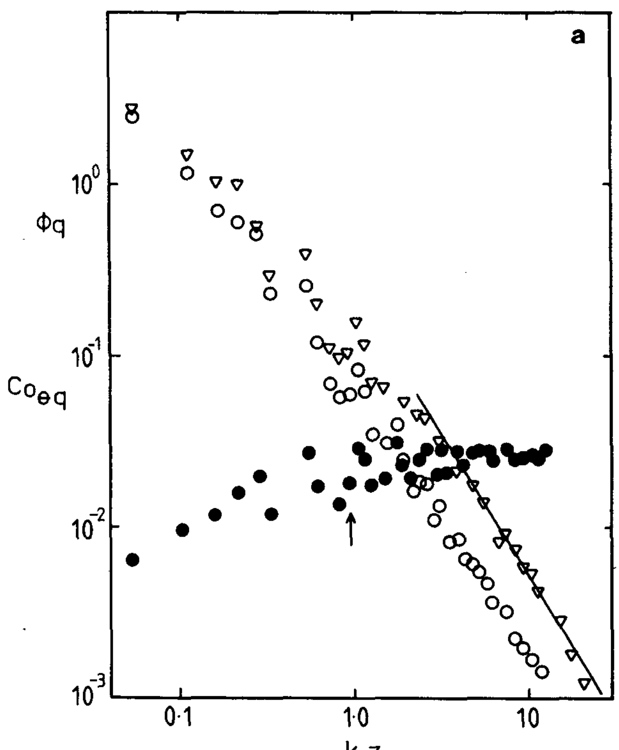

FIG. 6a. Spectral densities of humidity and covariance of humidity and temperature fluctuations measured prior to run 1: (○) q, (▽) $\mathrm{Co}_{\theta q}$, (●) $(k_1 z)^{5/3}\phi_q$. Straight line has a slope of $-5/3$.

4. Summary of results and concluding discussion

Spectra of u, w and uw cospectra measured at a height of ~5 m over the ocean show a considerable influence of the presence of the surface at or near the

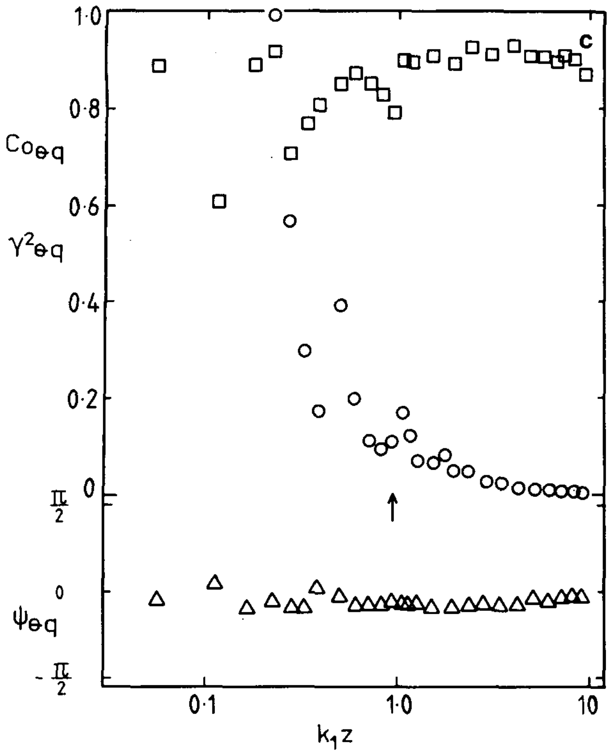

FIG. 6c. Covariance, coherence and phase angle for θq. Measurement prior to run 1. Symbols as in Fig. 4.

main wave frequency for a value of C/U_* of ~82. In particular, the estimated wave contributions to σ_u, σ_w and $-\overline{uw}$ are 14, 53 and −25%, respectively. These magnitudes are slightly larger than those reported by Benilov *et al.* (1974) for data corresponding to a similar value of C/U_*. As in the case of Benilov *et al.*, the *uw* cospectrum changes sign at or near the wave frequency n_0 and acquires relatively large positive values, indicating a momentum transfer from sea to wind. At or near n_0, the $w\theta$ and wq cospectra are nearly zero, the change of sign exhibited by these cospectra not being as evident as in the case of *uw*. The large coherence between *u* and *w* at $n = n_0$ and the phase shift of ~90° between *u* and *w* seems consistent with a description of the interaction between waves and turbulence as a quasi-potential wave perturbation of the turbulent flow. For $C/U_* \lesssim 40$, no discernible effect can be observed, at n_0, on ϕ_u or Co_{uw}, $\mathrm{Co}_{w\theta}$, Co_{wq}, but ϕ_w shows a distinct peak at the wave frequency. For all the present data, the wind velocity at the measurement height was smaller than the wave phase velocity. It seems, therefore, that Lai and Shemdin's (1971) recommendation that the direction of the induced stress is upward when the wind velocity at the height of the wave crest is smaller than C may be a necessary but not sufficient condition.

Spectra of θ and q and the θq cospectrum are essentially unaffected by the waves throughout the measurement period. It follows that no wave-induced disturbance is expected on spectra of refractive index fluctuations which are important when considering the propagation of visible radiation in the atmospheric surface layer. In particular, as a good −5/3 inertial subrange behavior is observed in ϕ_θ, ϕ_q and $\mathrm{Co}_{q\theta}$, the expectation of a −5/3 subrange in the refractive index spectrum is also good (e.g., Antonia *et al.*, 1978a). While the generation of fluctuations of temperature (or humidity) by the waves as a result of a fluctuating mean temperature gradient (e.g., Volkov, 1969; Kitaigorodskii, 1969) does not seem to be sufficiently important, the possibility of the generation of temperature fluctuations from static pressure fluctuations can also be dismissed. Friehe (1977, personal communication) estimated the level of $\overline{\theta^2}$ that can be generated in an adiabatic flow from turbulent static pressure fluctuations using the data of Elliott (1972a,b). The estimate may be written as

$$\sigma_\theta = 2.23 \times 10^{-3}\tau_s \text{ [°C]},$$

where τ_s is the mean surface shear stress (dyn cm^{-2}). Identifying τ_s with $-\overline{uw}$ (direct flux values in Table 1), the above relation represents contributions of ~0.5 and 1.4% to the measured standard deviations of runs 1 and 2, respectively. However, as suggested by Friehe, it is possible that the contribution to σ_θ due to pressure fluctuations may become important at higher wind speeds.

At $C/U_* \approx 82$, a good inertial subrange can be identified in the *u* spectrum although the bump at the wave frequency does affect the start of the inertial subrange. For θ and q, the beginning of the inertial subrange occurs at a frequency significantly larger than n_0 so that, even if a bump in ϕ_θ or ϕ_q did occur, it is unlikely that estimates of heat and humidity fluxes by the inertial dissipation technique

TABLE 2. Skewness and flatness factors for runs 1–4.

Run	S_u	F_u	S_w	F_w	$S_{uw-\overline{uw}}$	$F_{uw-\overline{uw}}$	S_θ	F_θ
1	−0.23	3.22	0.14	3.10	−1.69	14.12	0.39	2.72
2	−0.42	3.00	0.14	3.02	−1.11	8.35	0.14	2.81
3	−0.36	2.81	0.10	3.20	−1.94	13.95	0.86	3.54
4	−0.40	3.07	0.15	3.09	−1.78	11.71	0.35	2.65

would be influenced by the presence of waves. As the Kolmogorov constants for θ and q are not as reliably known as the Kolmogorov constant for u, there is little to be gained by the use of the inertial dissipation technique. In the case of u, it seems reasonable to believe that, provided the turbulence structure is not significantly influenced by the waves, the inertial dissipation technique should yield an adequate estimate of the shear stress that would apply in the absence of waves. As such, U_* values obtained from the dissipation technique, applied to u, would be estimates of the surface tangential stress. Table 1 gives estimates of the drag coefficient C_D, obtained by both direct and inertial dissipation methods, for runs 1–4. For run 1, $(C_D)_{\text{inertial}}$[8] is considerably higher than $(C_D)_{\text{direct}}$, as expected, since the wave contribution to $-\overline{uw}$ is such as to reduce the magnitude of the overall momentum flux. The difference between these two values of C_D represents ~46% of the inertial dissipation or unperturbed estimate of $-\overline{uw}$. The contribution caused by the depression in the uw cospectrum (Fig. 4) near n_0 was estimated in Section 3 to be ~25% of the unperturbed value of $-\overline{uw}$. This latter contribution is at best an approximation in view of the relatively crude method of determination. Also, it is possible that wave contributions to Co_{uw} can arise at frequencies significantly different from n_0 (e.g., Benilov *et al.*). It seems appropriate to comment on the suggestion by Schmitt *et al.* (1978b) that local anisotropy as measured by the ratio ϕ_w/ϕ_u in the inertial subrange, can lead to erroneous estimates of $(C_D)_{\text{inertial}}$. The inadequate response of the w propeller, which may have been responsible for a slightly reduced value of $(C_D)_{\text{direct}}$, does not enable us to calculate the ratio ϕ_w/ϕ_u accurately in the inertial subrange. However, Bradshaw (1967) found that the Kolmogorov constant remained unaffected down to a relatively low laboratory value of turbulence Reynolds number where local isotropy is clearly vitiated. It would appear that the inertial dissipation technique benefits from the relative insensitivity of the determination of the mean dissipation to, presumably, small departures from local isotropy. Thus it seems attractive to identify the difference between $(C_D)_{\text{inertial}}$ and $(C_D)_{\text{direct}}$ with the wave contribution to $-\overline{uw}$.

Acknowledgments. We acknowledge the generous cooperation of Dr. I. S. F. Jones, the Australian Navy Research Laboratory and ESSO/BHP. We are also indebted to Drs. D. Britz, C. A. Friehe, S. Rajagopalan and K. R. Sreenivasan for their assistance with the experiment and data reduction.

The research was supported by the Australian Research Grants Committee.

[8] A Kolmogorov constant of 0.5 was assumed (Antonia *et al.*, 1978b).

REFERENCES

Antonia, R. A., A. J. Chambers and C. A. Friehe, 1978a: Statistical properties of optical refractive index fluctuations in the marine boundary layer. *Bound.-Layer Meteor.*, **15**, 243–253.

——, ——, S. Rajagopalan, K. R. Sreenivasan and C. A. Friehe, 1978b: Measurements of turbulent fluxes in Bass Strait. *J. Phys. Oceanogr.*, **8**, 28–37.

Benilov, A. Y., and B. N. Filiushkin, 1970: Application of methods of linear filtration to an analysis of fluctuations in the surface layer of the sea. *Izv. Atmos. Ocean. Phys.*, **6**, 810–819.

——, and M. M. Zaslavskiy, 1974: Determination of wave and turbulent components of random hydrodynamic fields in the marine atmospheric surface layer. *Izv. Atmos. Ocean. Phys.*, **10**, 628–635.

——, O. A. Kuznetsov and G. N. Panin, 1974: On the analysis of wind wave-induced disturbances in the atmospheric turbulent surface layer. *Bound.-Layer Meteor.*, **6**, 269–285.

Bradshaw, P., 1967: Conditions for the existence of an inertial subrange in turbulent flow. ARC R & M No. 3603.

Davidson, K. L., 1974: Observational results on the influence of stability and wind-wave coupling on momentum transfer and turbulent fluctuations over ocean waves. *Bound.-Layer Meteor.*, **6**, 305–331.

——, and A. J. Frank, 1973: Wave-related fluctuations in the airflow above natural waves. *J. Phys. Oceanogr.*, **3**, 102–119.

De Leonibus, P. S., 1971: Momentum flux and wave spectra observations from an ocean tower. *J. Geophys. Res.*, **76**, 6506–6527.

Elder, F. C., D. L. Harris and A. Taylor, 1970: Some evidence of organized flow over natural waves. *Bound.-Layer Meteor.*, **1**, 80–87.

Elliott, J. A., 1972a: Microscale pressure fluctuations measured within the lower atmospheric boundary layer. *J. Fluid Mech.*, **53**, 351–383.

——, 1972b: Microscale pressure near waves being generated by the wind. *J. Fluid Mech.*, **54**, 427–448.

Fairall, C. W. and G. Schacher, 1977: Frequency response of hot wires used for atmospheric turbulence measurements in the marine environment. *Rev. Sci. Instrum.*, **48**, 12–17.

Kitaigorodskii, S. A., 1969: Small-scale atmosphere-ocean interactions. *Izv. Atmos. Ocean. Phys.*, **5**, 1114–1131.

——, 1973: *The Physics of Air-Sea Interaction*. Israel Program for Scientific Translations, 237 pp.

Kondo, J., Y. Fujinawa and G. Naito, 1972: Wave-induced wind fluctuation over the sea. *J. Fluid Mech.*, **51**, 751–771.

Kononkova, G. E., E. A. Nikitina, L. V. Poborchaya and D. A. Speranskaya, 1973: Experimental study of the turbulence structure of the air flow near a water surface. *Fluid Mech.-Sov. Res.*, **2**, 55–58.

Lai, R. J. and O. H. Shemdin, 1971: Laboratory investigation of air turbulence above simple water waves. *J. Geophys. Res.*, **76**, 7334–7350.

McIntosh, D. A., R. L. Street and E. Y. Hsu, 1975: Turbulent heat and momentum transfer at an air-water interface: The influence of surface conditions. Tech. Rep. No. 197, Dept. Civil Engineering, Stanford University, $$$ pp.

Schacher, G., and C. W. Fairall, 1976: Use of resistance wires for atmospheric turbulence measurements in the marine environment. *Rev. Sci. Instrum.*, **47**, 703–707.

Schmitt, K. F., C. A. Friehe and C. H. Gibson, 1978a: Humidity sensitivity of atmospheric temperature sensors by salt contamination. *J. Phys. Oceanogr.*, **8**, 151–161.

——, —— and ——, 1978b: Sea surface stress measurements. *Bound.-Layer Meteor.*, **15,** 215–228.

Sreenivasan, K. R., A. J. Chambers and R. A. Antonia, 1978: Accuracy of moments of velocity and scalar fluctuations in the atmospheric surface layer. *Bound.-Layer Meteor.*, **14,** 341–359.

Takeuchi, K., E. Leavitt and S. P. Chao, 1977: Effects of water waves on the structure of turbulent shear flow. *J. Fluid Mech.*, **80,** 535–559.

Volkov, Y. A., 1969: The spectra of velocity and temperature fluctuations in airflow above the agitated sea surface. *Izv. Atmos. Ocean. Phys.*, **5,** 1251–1265.

——, 1970: Turbulent flux of momentum and heat in the atmospheric surface layer over a disturbed sea-surface. *Izv. Atmos. Ocean. Phys.*, **6,** 1295–1302.

Young, M. B. O., E. Y. Hsu and R. L. Street, 1973: Air-water interaction: the nature of turbulent heat, mass and momentum transfer mechanisms in the air boundary layer. Tech. Rep. No. 163, Dept. Civil Engineering, Stanford University, 187 pp.

Zubkovskiy, S. L., O. A. Kuznetsov and G. N. Panin, 1974: Certain results from measurements of the fluctuations of temperature, humidity and wind velocity in the marine surface layer. *Izv. Atmos. Ocean. Phys.*, **10,** 655–660.