Turbulence measurements over the sea by a tethered-balloon technique

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**SUMMARY**

Turbulent fluctuations of wind, temperature and humidity at heights up to a few hundred metres have been measured over the open sea by instruments supported by a tethered balloon. Wave-induced motion of the ship from which the balloon was flown produced unwanted contributions to the measured turbulence, but in light or moderate winds it has been possible to make allowances for these effects and to obtain useful estimates for vertical fluxes of heat and moisture: the calculated momentum fluxes in contrast appear less reliable and to obtain satisfactory values will require stabilization of the tethering point of the balloon cable. Some observations made in generally unstable conditions near OWS 'J' yield a value of $(1.8 \pm 0.7) \times 10^{-3}$ for $C_F$ in the bulk aerodynamic formulation for evaporation ($E = \rho C_F (q_0 - q_1)(U_{10} - U_0)$) and suggest that compared to the vertical moisture flux the heat flux may often have an insignificant role in determining the buoyancy flux and hence the turbulence structure in the bulk of the boundary layer.

**NOTATION**

$C_D$ drag coefficient for momentum transfer

$C_{H}, C_F$ coefficients for heat and moisture transfer

$c_p$ specific heat

$e$ vapour pressure

$E$ vertical moisture flux

$f$ reduced frequency ($n \pi / U$), Coriolis parameter

$H$ vertical heat flux

$k$ von Kármán's constant

$L$ Monin-Obukhov length

$n$ frequency

$T$ air temperature

$T_*$ $-H/\rho c_p u_*$

$T_W$ wet-bulb temperature

$p$ atmospheric pressure

$q$ specific humidity

$q_*$ $-E/\rho u_*$

$S_{uu}(n), S_{ww}(n), S_{TT}(n), S_{qq}(n)$ spectral estimates for turbulent fluctuations

$S_{uu}(n), S_{Tw}(n), S_{qy}(n)$ cospectral estimates

$u_*$ friction velocity

$u, v, w$ velocity components

$U$ horizontal velocity

$V_g$ geostrophic wind speed

$\alpha$ geostrophic wind veer

$\rho$ air density

$\theta$ potential temperature

$\phi$ term adjusting vertical gradients to diabatic conditions

$\psi$ term adjusting log profiles to diabatic conditions

$\tau$ vertical momentum flux

$\alpha_{u, w, \alpha T, \alpha q}$ standard deviations of turbulent fluctuations
In most investigations of atmospheric turbulence over the open sea the sensors have been mounted on ships, masts or buoys (e.g. Hasse 1968) and the measurements confined therefore to within a few metres of the surface. Corresponding observations at greater heights which would be required in any comprehensive investigation of the vertical structure of the turbulent boundary layer are by comparison rare, and are restricted mainly to a few studies using aircraft (e.g. Bunker 1955, 1960). It is now argued that a tethered balloon technique for supporting the turbulence instrumentation, despite limitations in altitude, offers certain advantages over aircraft in this context: for example it can be used remotely from land if a suitable ship is available, and measurements can be made at different heights simultaneously.

Studies of turbulence over land using balloon-supported instrumentation have been made for a number of years at Cardington, England in particular (Smith 1961; Thompson 1966; Readings and Rayment 1969), while Yokoyama (1969) has described a tethered balloon system that has been used at sea (Yokoyama, Hayashi and Ogura 1969). The present author, however, has not been able to find reports in the open literature of the use of tethered balloon techniques to obtain vertical fluxes of heat, moisture and momentum. In part this may be due to the inherently unstable nature of the support platform for instruments which is offered by a balloon and its tethering cable. Over land at least, in unstable conditions the characteristic motion of balloons such as those used at Cardington (about 25 m long) is a periodic crosswind migration initiated presumably by fluctuations of wind azimuth, the period being a function of the height of balloon above ground: there are indications that these motions can result in significant contributions to the measured vertical fluxes of momentum (Thompson 1969; Readings and Butler 1971). The magnitudes of both mechanically and thermally produced turbulence are usually much less over the sea and balloon motions of this nature are therefore likely to be of less significance in their effect on measured fluxes.

Two important limitations of the technique specifically when used at sea stem from firstly the impossibility of obtaining convincing turbulence data close to the surface due to the disturbing influence of the ship, and secondly the wave-induced motion of this ship. It is not clear what is the lowest height at which the ship’s influence may still be neglected: in the present experiments it has been taken arbitrarily to be about 100 m when headed into wind, or 50 m when lying-to but further data will be required before this point can be resolved. Wave-induced motion of the ship and hence of the balloon tethering point provide a much more important obstacle to the satisfactory measurement of turbulence by instruments attached to the cable. Generally the space available for inflation limits the maximum size of balloon which can be used with the result that the balloon is small and often fairly heavily laden. With instruments mounted at more than one level the result is substantial ‘sag’ in the cable so that vertical motions of the tethering point produce not only vertical displacements of the sensor systems but also comparatively large-amplitude oscillations (and therefore accelerations) in the horizontal. This introduces particular difficulties if the instrumentation for measuring vertical wind fluctuations is gravity-orientated by a pendulum system.

This paper is concerned with the results from a series of measurements near OWS ‘J’ of turbulent fluctuations in the lowest 200 m. Observations were made over a wide range of meteorological conditions, sometimes at two heights simultaneously, and have been used to derive spectra and cospectra, and vertical eddy fluxes.

2. Instrumentation

The helicopter flight deck of the H-class naval survey vessel HMS Hecla was used for inflating and tethering the balloon. This last was about 10 m long, 80 m$^3$ volume and helium-filled with a free lift of about 50 kg. It was tethered by 1,000 kg steel cable controlled by an electric winch. The system was designed for use in winds up to at least 25 m s$^{-1}$ and survived squalls of 20 m s$^{-1}$ without harm, though ultimately it was damaged by such a squall while being topped-up on deck. The main limitation on altitude was due to weight (0.05 kg per m)
and drag of the tethering cable. Maximum altitude with a single turbulence-measuring package was in excess of 400 m but this could be increased to about 600 m in spells of settled weather by using a lighter cable.

The turbulence-measuring packages each weighed about 3.5 kg; they were vane-like instruments about 1 m long which pivoted about the cable and measured directly or indirectly horizontal and vertical wind speeds, temperature and humidity. The wind sensors were a miniature photoelectric cup anemometer mounted with its rotor axis horizontal which measured magnitude of the wind vector, and a heated-wire device (Jones 1961) mounted on a damped pendulum to sense the inclination of this vector to the horizontal. Time constants in a 5 m s\(^{-1}\) wind were around 0.1 and 0.01 s respectively. Air temperature and humidity were obtained by fine-wire wet and dry resistance thermometers connected in an AC bridge circuit. The wet bulb was made from 25 μm platinum wire around which were wound two fine cotton threads in the form of a double helix: the whole was twisted with a further cotton thread and mounted on wetted vertical cotton wicks in a similar way to that described by Dyer and Maher (1965). The measured diameter of the element when wet was about 0.5 mm and the time constant therefore near 1 s (Taylor 1963). A simple shield was used to screen the element from direct solar radiation. Wind-tunnel tests showed variations of less than 0.02 °C in indicated wet-bulb temperature for speeds between 2 and 15 m s\(^{-1}\). Time constant of the 'dry-bulb' thermometer which was constructed from similar wire was around 0.02 s. The signals were converted to an audio-frequency multiplex and transmitted either via electrical cables or a 400 MHz radio link. The received signals were recorded in the ship's laboratory on a multi-track magnetic tape recorder, and were also demultiplexed and discriminated for display on UV recorders.

3. Data and data processing

The observations to be described were obtained during the JASIN expedition, a Royal Society sponsored air-sea interaction experiment near OWS 'J' involving three ships carrying out a number of meteorological and oceanographic investigations (Royal Society Report 1971). Turbulence data were obtained over a period of about a week. The start of the period coincided with light winds and generally stable density stratification near the surface but conditions then became more unsettled with stronger winds, almost continuous convection, and significant air-sea exchanges. The duration of the experimental runs and associated meteorological data are listed in Table 1.

The observations of temperature and humidity from the ship were made at hourly intervals using a ventilated psychrometer and bucket thermometer both deployed to avoid the disturbing influence of the ship as far as possible. The buoy data were obtained by a prototype self-recording system. Sea temperatures from this were thought to be closer estimates of near-surface conditions than those from the ship in the first part of the observational period when there was a marked increase of temperature near the surface due to absorption of solar radiation: the stirring action of the ship by contrast tended to destroy this surface thermocline. Sea temperatures obtained from the ship were preferred at other times because of marked horizontal temperature gradients which were observed across the area, coupled with the separation of ship and buoy of typically 5–10 km.

The tethered balloon was flown continuously apart from topping-up every day or so. Turbulence measurements were made with the ship sailing up or downwind, or else hove-to or lying-to. All usable data were obtained with observations at either one level, or two levels simultaneously, at heights between 45 m and 185 m (the minimum distance between balloon and a turbulence package was 30 m). The maximum period of continuous observation (determined by magnetic-tape length) was in excess of three hours but the actual lengths of data analysed were normally considerably shorter, usually being longer runs subdivided to ensure that meteorological conditions were as far as possible stationary or to eliminate sections where ship-maneuvres were taking place.

On playback of the turbulence data the signals were demultiplexed and after frequency-
### TABLE 1. Experimental runs and mean meteorological data

<table>
<thead>
<tr>
<th>Run No.</th>
<th>Date</th>
<th>Period (Z)</th>
<th>Instrument heights (m)</th>
<th>Measured (a) wind speed ms(^{-1})</th>
<th>Buoy wind (b) speed ms(^{-1})</th>
<th>Air temp (c) (°C)</th>
<th>Specific (c) humidity x10(^3)</th>
<th>Sea temp (c) (°C)</th>
<th>Sea temp (d) (buoy) (°C)</th>
<th>(\theta_b - \theta_{10})</th>
<th>(q_b - q_{10}) x10(^3)</th>
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<td>8.9</td>
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<td>1.7</td>
<td>2.8</td>
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</table>

(a) corrected for ship's drift velocity  
(b) at 2 m  
(c) measured from ship at about 10 m  
(d) at \(-0.7\) m
to-voltage conversion were then filtered by 4-stage RC low-pass networks with cut-off \((-12 \text{ db})\) at the Nyquist frequency, prior to sampling at one-second intervals by an analogue-digital convertor with punched paper tape output (short sections of some records were also sampled at five times a second – details are given later). Calculations were performed on a KDF9 computer. The convertor outputs were first scaled and converted to speed, inclination, temperature and specific humidity, non-linearities in the sensors and telemetry system being represented by cubic functions. Specific humidities were obtained from vapour pressures calculated by Regnaut's formula

\[
e = (e_o)_w - Ap(T - T_w)\]

where \(A = 6.60 \times 10^{-4}(1 + 1.15 \times 10^{-2}T_w)\) (Smithsonian Meteorological Tables): \((e_o)_w\)
was calculated using the approximation \((T_w^2 - 488.56T_w + 60009.3)/(0.0361622T_w^2 - 24.29T_w + 4104.45)\) given by Langlois (1967). Wind vector magnitude and inclination were converted to horizontal and vertical speeds after an axis rotation to make \(\langle w_r \rangle = 0\). Linear trends were removed and the vertical turbulent fluxes of heat, moisture and momentum were then calculated by the eddy correlation method. Power and cross spectra were also obtained using the Fast Fourier Transform (Rayment 1970).

4. Results and discussion

The UV chart records had shown before the detailed analysis described above was carried out that the effects of ship motion on measured speed and inclination were very severe, the typical spurious contributions being around \(\pm 1 \text{ m s}^{-1}\) and \(\pm 0.2 \text{ radians}\). These were not primarily the results of vertical displacements of the instruments but as explained earlier were due to predominantly horizontal movements which caused changes of measured wind speed and displacements of the inclinometer pendulum from the true vertical. This substantial source of noise had been anticipated from data obtained in 1969 by a prototype of the present system during the Atlantic Trade Wind Expedition (ATEX) (Thompson 1971a). It appears to preclude the satisfactory calculation of turbulent fluxes but the majority of the noise is concentrated in a fairly narrow band of frequencies, and empirical data on the frequency range of cospectra measured over the sea (Miyake, Donelan and Mitsuta 1970; Miyake, Stewart and Burling 1970) suggest that a cospectral 'gap' is likely to exist between the peaks due to noise and genuine turbulent fluctuations. It will become evident that this is an oversimplification because significant noise is also produced by sensor motion at subharmonically-related frequencies but in spite of this the simple assumption provides a useful basis for eliminating much of the spurious contribution. Miyake et al.'s data support the idea that contributions to momentum and moisture fluxes are negligible at reduced frequencies \(f = nz/u\) greater than about 1 but show by contrast the cospectra for the heat flux significant up to \(f \approx 4\). The main ship motion was at a frequency of about \(0.15 \text{ Hz}\) and if it is assumed that reduced-frequency scaling will produce effective coincidence of normalized cospectra obtained at different heights (Panofsky and Mares 1968) then in winds up to about \(10 \text{ m s}^{-1}\) there should be little difficulty in isolating contributions due to sensor motion at the characteristic frequency of the ship from those due to genuine turbulent fluctuations in the cases of momentum and moisture fluxes at heights as low as \(50 \text{ m}\). There would be some ambiguity for the heat flux unless winds were light or the measurement level was at least \(100 \text{ m}\).

Variance spectra extend to higher frequencies and therefore will have significant contributions in the region strongly influenced by sensor motion. Spectra obtained on a convective day during JASIN at a height of \(90 \text{ m}\) (Fig. 1(a)) show this quite clearly. The majority of the variance of both horizontal and vertical fluctuations is seen to be associated with the strong noise peak centred between \(0.1\) and \(0.2 \text{ Hz}\) but there are also lesser contributions at around \(0.08\) and \(0.04 \text{ Hz}\), apparently sub-harmonics of the main peak. The temperature spectrum shows substantial noise at higher frequencies (>\(0.3 \text{ Hz}\) produced by the telemetry and recording system, but in contrast to \(U\) and \(w\) the main contributions from ship motion are at frequencies below \(0.1 \text{ Hz}\). These result from vertical displacements in a nearly adiabatic
atmosphere \((\partial T/\partial z \approx -0.01 \, ^{\circ}C \, m^{-1})\). The relative importance of the sub-harmonics in this case suggests that sensor motion is generally nearly horizontal and that substantial vertical motion only occurs as 'beats' that result from differences between wave-frequency and the ship's natural frequency. By comparison the humidity spectrum shows only residual effects due to ship motion. This follows from similarity arguments using very rough approximations for the true variances (thus \(\sigma_T/\sigma_e \approx \sigma_q/\sigma_e\) and hence \(\sigma_T/\sigma_e \approx (\partial T/\partial z)/(\partial q/\partial z)\)). Because \(|\partial T/\partial z| > |\partial \theta/\partial z|\) vertical sensor motion will produce therefore relatively larger effects on the temperature than the humidity spectrum). There is some noise above 0.3 Hz but apart from this the ultimate slope is less than the \(-2/3\) expected in the inertial subrange: this is because of fall-off of wet-bulb sensor response at higher frequencies.

The corresponding cospectra (Fig. 1(b)) all show significant contributions which can be unambiguously associated with sensor motion. The \(Uw\) cospectrum is very severely 'contaminated,' as would be expected from the spectra of the velocity components but there is the cospectral gap referred to above and it is obviously possible to isolate the main noise contribution. The cospectrum for \(Tw\) is less affected partly because of the smaller noise content of the temperature fluctuations than of the horizontal velocity fluctuations at frequencies below about 0.2 Hz, and also because \(T, w\) correlation coefficients are larger than those for \(U, w\) in unstable conditions (Pond, Phelps, Paquin, McBean and Stewart 1971; Phelps and Pond 1971). However, there are clearly significant noise peaks at 0.08 and 0.04 Hz, frequencies where contributions from turbulent fluctuations are almost certainly not negligible. Sensor motion has comparatively little effect on the \(q, w\) cospectrum though the spurious contributions can still be recognized, at around 0.04 Hz in particular.
(a) Measured momentum fluxes

The vertical eddy-fluxes of momentum are $\tau_{xz} = \rho \langle uw \rangle$, $\tau_{yz} = \rho \langle vw \rangle$, with the x-axis aligned along the mean wind direction. The instrumentation used in the present investigation was not able to resolve the horizontal wind into its two components, and the calculated flux was therefore $\rho \langle Uw \rangle$, or $\tau_{xz}(1 + <v^2>/2 <U^2>)$ approximately (Pasquill and Tyldesley 1967). The total flux ($\tau_{xz} + \tau_{yz}$) can be represented with sufficient accuracy by $\tau_{xz}(1 + \tau_{yz}^2/2\tau_{xz}^2)$ and the error in equating this to $\rho \langle Uw \rangle$ is therefore $\tau_{xz}(\tau_{yz}^2/2\tau_{xz}^2 - <v^2>/2 <U^2>)$. McBean (1971) found $<v^2>/u_*^2 \sim 5$ in near-neutral conditions, and using the rough approximation $u_*^2 \sim 10^{-9} <U^2>$ (see below), the second term inside the brackets is therefore negligible. Swinbank's (1970) analysis of the Leipzig wind profiles suggests that the preceding term is very small (less than 0.01) at all heights in the boundary layer, but Carson and Smith's (1971) analysis of the same data showed a rapid increase in the size of this term with increasing height, being 0.06 at 200 m and 0.23 at 400 m where the total stress was 0.4 times the surface value. Over the sea in middle latitudes the geostrophic veer is around 5 degrees (Mendenhall 1967) compared with around 25 degrees for the Leipzig profiles and the approximation $\tau = \tau_{xz} = \rho \langle Uw \rangle$ is probably accurate to a few percent at heights up to several hundred metres, but in trade wind areas where the veer may be as large as 30 degrees (Charnock, Francis and Sheppard 1956) the errors may be substantially larger.

Fluxes (Table 2) were calculated in three ways, (a) by eddy correlation using the original unaveraged data, (b) by eddy correlation after block-averaging the data into overlapping 8-second sections (equivalent to low-pass filtering with cut-off (−3 db) at 0.06 Hz) and (c) by planimetry of the cospectra which had been smoothed at their high-frequency end to zero at $f = 1$. A comparison of (a) with (b) or (c) shows the important contribution to the raw 'fluxes' from sensor motion. Estimates from methods (b) and (c) agree closely in most cases, but (c) is preferred in spite of its subjectivity because it is less likely to remove genuine turbulent contributions. The Table also lists estimates for the surface fluxes calculated using the conventional bulk aerodynamic formula

$$\tau_0 = -\rho C_D(\langle U_{10} \rangle - \langle U_0 \rangle)^2$$

(1)

$U_0$ (the mean surface drift velocity) is about 0.04 $\langle U_{10} \rangle$ (Keulegan 1951), and $C_D$ is around $1.3 \times 10^{-3}$ (Hasse 1968; Smith 1970). $\langle U_{10} \rangle$ was estimated from the wind measured by the turbulence instrumentation (corrected for ship's drift) using the constant-flux layer relation

$$\langle U_{10} \rangle = \langle U_z \rangle - u_*(\log(z/10) + \psi_M(10/L) - \psi_M(z/L))/k$$

(2)

Values for $\psi$ were obtained from Dyer and Hicks (1970) and Webb (1970) in unstable and stable conditions respectively. For this purpose a preliminary value of $L(-u_*(c_D\rho T/\rho) (H+0.61 \, T_C p E))$ was determined with sufficient accuracy by the relation (Thompson 1971 b)

$$L = -2.4 <U_{10}'>^4/(\theta_o - \theta_{10} + 175 (q_0 - q_{10}))$$

(3)

which follows from Eq. (1) and the bulk aerodynamic formulation for heat and moisture transfer over the sea:

$$H_o = \rho C_E H(\theta_o - \theta_{10}) (\langle U_{10} \rangle - \langle U_0 \rangle)$$

(4)

$$E_o = \rho C_E (q_0 - q_{10}) (\langle U_{10} \rangle - \langle U_0 \rangle)$$

(5)

$\theta_o$ and $q_o$ were related directly to the measured bulk water temperature without any attempt to correct for 'skin' effects (Hasse 1971). $C_E$ and $C_H$ were assumed here to be equal to $1.3 \times 10^{-3}$ (Robinson 1966; Chamberlain 1968): $<U_{10}'>$ was a rough estimate of the wind at 10 m, usually taken as 0.9 $\langle U_z \rangle$. The values for $C_D$ actually used in Eq. (1) were based on a near-neutral value of $1.3 \times 10^{-3}$ but were corrected for stability (Deardorff 1968; Thompson 1971 b). The final listed $L$'s were obtained using Eqs. (1), (4) and (5) with $C$'s corrected for stability.

The final column of Table 2 compares measured fluxes with estimated surface values
## TABLE 2. MOMENTUM FLUXES

<table>
<thead>
<tr>
<th>Run No.</th>
<th>Instrument heights (m)</th>
<th>$\langle U_{10} \rangle$ (ms$^{-1}$)</th>
<th>(a) Unaveraged data</th>
<th>(b) 8-sec av. data</th>
<th>Fluxes (Nm$^{-2}$)</th>
<th>(c) Cospec smooth to zero at $f = 1$</th>
<th>(d) Estimated surface flux</th>
<th>$L$ (m)</th>
<th>Measured flux</th>
</tr>
</thead>
<tbody>
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<td>1</td>
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(i) Sensor motion too violent for proper functioning of inclinometer
for the second half of JASIN where because of the generally higher wind speeds any errors in estimating ship's drift velocities were insignificant in their effect on the calculated $\tau_0$. The results for runs 17 and 18 were unreliable because of an anemometer fault which resulted in 'clipping' of the highest wind speeds and although this produced only small errors in measured vertical velocities the horizontal fluctuations were significantly underestimated: this in turn resulted in spurious positive contributions to the momentum fluxes. Discarding these two runs the average of the flux ratios is 0.74 and the mean measurement height 100 m.

In these generally unstable conditions the variation of $\tau$ with height may be approximated by $\rho f V_p \sin \alpha$ in the lowest 100 m or so, and assuming $V_p = 8.5$ m s$^{-1}$ ($<U_{10}> = 7.3$ m s$^{-1}$) and $\alpha = 5^\circ$ for a near-neutral lapse (Mendenhall 1967) then $\tau_{100}/\tau_0 \approx 0.88$. The measured fluxes therefore appear to be too small (the difference is significant at about the 15 per cent level). The explanation seems to lie with small or sometimes positive values which were found in many of the cospectra at frequencies between about 0.01 and 0.05 Hz, possibly the result of sensor motion at sub-harmonics of the main ship's frequency, though $Uw$ correlations introduced by periodic migration of the balloon acrosswind cannot be ruled out at these frequencies (it is seen that the results from run 15, the experiment with the highest wind speed, have been discarded also: sensor motion became very violent on this occasion and included lateral oscillations of the vane about its pivot on the balloon cable which prevented the instruments functioning properly). The measured values for $\tau$ when corrected for variation with height and used in conjunction with the estimated 10 m winds gave an average drag coefficient of $1.12 \pm 0.43 \times 10^{-3}$.

The corresponding cospectral estimates, normalized by $<Uw>$ and plotted against reduced frequency are shown in Fig 2: the data at natural frequencies above about 0.1 Hz have not been used in order to remove the largest spurious contributions due to sensor motion. Periodicities such as those shown in Fig. 1(b) for values of $f$ between about 0.1 and 1 are concealed by the frequency scaling because of the variety of measuring heights and wind speeds in the experiments, but in spite of the scatter their overall contribution can be recognized by the small size on average of the cospectral estimates for $f > 0.1$ (Miyake et al. 1970b). It is almost certain therefore that the measurement technique underestimates the momentum fluxes. The large fluctuations at low frequencies reflect the comparatively short sampling periods for some of the runs and make a close comparison with other 'universal' cospectra obtained over the sea but nearer to the surface (Miyake et al. 1970a,b; Pond et al. 1970) very difficult: however, the frequency of the cospectral peak (between $f = 0.1$ and $f = 0.01$) and a rapid decrease in cospectral values as $f$ approaches 0.001 are consistent with the results of other investigations.

(b) heat fluxes

Table 3 lists the eddy fluxes of heat $(\rho c_p <Tw>)$ obtained in three ways from the turbulence data, and also the corresponding surface values estimated using Eq. (4). The preferred measured flux is again that deduced from the cospectrum (here smoothed to zero at $f = 4$) but differences between values obtained in this way and those calculated directly from the raw (unaveraged) turbulence data are small in most cases, demonstrating that sensor motion produced much smaller contributions than in the case of momentum flux.

The ratios of fluxes measured at about 100 m to estimated surface values on five occasions when sea-air temperature differences were around 1 degree C or greater showed a good deal of scatter but the average of 0.8 suggests that the measurement technique is reasonably satisfactory (a value of less than 1 is to be expected in view of the vertical flux divergence found when measurements were made at two levels in thermally unstable conditions).

Cospesrtal estimates normalized by $<Tw>$ were plotted against reduced frequency for all the runs but there was no evidence for a universal cospectrum. An explanation for this is provided by Fig. 3 where cospectra from run 10 are shown for the two heights of measurement: on this occasion the vertical heat flux changed sign between the lower and upper level as a result of increasing negative contributions at lower frequencies but the small-scale flux was directed upwards at both heights (estimates above about 0.06 Hz are very significantly
### TABLE 3  Heat and moisture fluxes

<table>
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<tr>
<th>Run No.</th>
<th>Inst. hts. (m)</th>
<th>(a) Unav data</th>
<th>(b) 8-sec av. data</th>
<th>Heat fluxes (W/m²) smoothed to zero at f = 4</th>
<th>Measured flux</th>
<th>(c) Cospec (mg m⁻² s⁻¹) smoothed to zero at f = 1</th>
<th>Moisture fluxes</th>
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N. THOMSON
affected by noise produced by sensor motion). Some of the other runs with measured upward heat flux also produced cospectra with negative contributions at low frequencies.

Normalized cospectra for the four runs with the largest upward fluxes and similar instrumental heights are shown in Fig. 4. The scatter at low frequencies is considerable, as in Fig. 2, and it suggests that measuring periods should be several hours long to obtain stable cospectral estimates when measurements are made above the surface layer. Contributions due to sensor motion can be recognized at higher frequencies, especially near \( f = 0.5 \) and \( f = 1 \) (estimates at natural frequencies above 0.1 Hz have not been plotted). There is otherwise broad agreement with the results of Miyake et al. (1970b) and Phelps and Pond (1971), with the major contributions to the heat flux between reduced frequencies of 0.1 and 0.01, and a rapid decrease in cospectral values at lower frequencies.

(c) moisture fluxes

These showed smaller variations with height than the heat fluxes (Table 3) and on convective occasions it is likely that the divergence up to 100 m or so is small usually so that surface values will differ little from those measured a few tens of metres above the surface. The cospectra showed only small contributions from sensor motion and the ‘raw’ flux
estimates agreed closely with those from the smoothed cospectra. The table includes values for $C_E$ obtained using Eq. (5) in ten unstable and two slightly stable cases where the surface wind could be estimated with reasonable accuracy. The mean value of $(1.78 \pm 0.68) \times 10^{-3}$ may be compared with $1.25 \times 10^{-3}$ obtained by Robinson (1966) from heat budget considerations in predominantly unstable conditions, and $1.32 \times 10^{-3}$ by Chamberlain (1968) from wind-tunnel studies, and thus appears rather large. It is heavily weighted, however, by the results from runs 14 and 16 and eliminating these would reduce the mean to $(1.51 \pm 0.42) \times 10^{-3}$, not significantly different from Robinson's and Chamberlain's values. The comparatively large standard deviation of the estimates for $C_E$ is probably due to the sampling errors inherent in measuring fluxes above the immediate surface layer in generally convective conditions.

Corresponding cospectra (normalized by $<q_w>$) are shown in Fig. 5 for natural frequencies below 0.1 Hz. There is considerable scatter with important contributions from sensor motion above reduced frequencies of about 0.3. Otherwise the cospectra are similar to those found by Phelps and Pond (1971) and Miyake et al. (1970a).

Fig. 6 gives $q_w$ cospectra for run 10 at both heights of measurement. If the obvious contributions from sensor motion are smoothed out the cospectra have then broadly similar shapes, in sharp contrast to those for $T_w$ (Fig. 3). It is interesting to note that the frequencies of the major contributions to both heat and moisture fluxes were the same at each level ($5.5 \times 10^{-3}$ and $1.7 \times 10^{-2}$ Hz at 45 m, $4.0 \times 10^{-3}$ Hz at 140 m) even though the sign of the heat flux contributions reversed between the two heights: this demonstrates that the same eddies were transferring the heat and moisture in the vertical, at least at these frequencies, and suggests that the upper level of measurement was in a region with sub-adiabatic lapse rate.

During JASIN there was cumulus development on a number of occasions when measured vertical heat fluxes were small or negative at heights well below cloud base (runs 10–13, 16) and the convection must have been supported then entirely by the moisture flux. In these
circumstances downward heat fluxes below cloud base might be anticipated to result from release of latent heat in the convective condensation process. The sensible heat required to evaporate droplets from the decaying clouds and hence prevent the released latent heat being re-absorbed was obtained presumably from warmer air which was turbulently entrained from the inversion capping the boundary layer.

(d) spectra

Typical $U$ and $q$-spectra obtained near the surface over the sea (Miyake et al. 1970b; Phelps and Pond 1971) and plotted in the form $nS(n)$ against $f$ have peaks at reduced frequencies around $10^{-2}$. $T$-spectra show considerable variations due to radiative effects (which depend on humidity) but in middle latitudes where humidities are comparatively low the peak frequency is also around $10^{-2}$. $w$-spectra in contrast have their peak at frequencies about an order higher. The spectra plotted in Fig. 1(a) have features similar to those but apart from $q$ they are strongly contaminated by noise due to sensor motion for $f$ greater than about 1: this is a region where genuine turbulent fluctuations are not insignificant and thus any filtering procedure is unable to remove the noise without removing also valid contributions to the variances. However, the results of, for example, Pond et al. (1971) and Phelps and Pond (1971) show that an inertial subrange starts at reduced frequencies around or somewhat lower than 1 and this suggests that the present spectra might be corrected for the major noise contribution provided that a reliable spectral estimate can be obtained anywhere in the
inertial subrange. This cannot be done very convincingly at the lower-frequency end because of sub-harmonics (Fig. 1(a)) so attempts were made to obtain spectral estimates for \( U, \ w \) and \( T \) at frequencies above those plotted in the Figure. Short lengths of data equally spaced through runs 10, 11 and 12 were sampled at 5 times-per-second after 4-stage RC filtering with cut-off at the Nyquist frequency and the spectral estimates from each section of data were averaged before plotting. An example is shown in Fig. 7: here, as in the previous Figures, the estimates have been corrected for attenuation due to RC filtering. The periodicities which appear at higher frequencies are harmonically related to those in the main peak and because of them it is not possible to decide with any certainty what the spectrum should be in the inertial interval in the absence of sensor motion. This in turn prevents the reliable estimation of variances of the genuine turbulent fluctuations of wind and to a lesser degree of temperature also. Wind spectra plotted in Fig. 8 are normalized therefore by \( u_\text{ref}^2 \) rather than by the corresponding variances (the spectra are not extended to natural frequencies above about 0.1 Hz to eliminate the major contribution from sensor motion). Noise due to sensor motion dominate the spectra at high frequencies. At low frequencies there is a measure of agreement with results from near-surface studies, apart from the scatter in the \( u \)-spectra which is caused presumably by incomplete sampling of meso-scale features.

Normalized temperature spectra are plotted in Fig. 8 for the four runs with the largest measured heat flux, where noise contributions to the variance of the temperature fluctuations were comparatively small. Measured variances for runs 17 and 18 (\( \sim 0.01(\text{C})^2 \)) were several times larger than those found for the other runs and the signal-to-noise ratios were correspondingly more favourable. This is reflected in the spectra for the two final runs, which (for \( f \)
Figure 6. $q_w$ - cospectra, run 10.

Figure 7. $w$ - spectrum at 140 m, run 10.
greater than about 1) have slopes close to the minus two-thirds expected in the inertial subrange; in contrast the other spectra are markedly affected by noise over the same frequency range. The general spectral shapes and the suggested low-frequency limit for the inertial interval are similar to those found by Phelps and Pond off San Diego, where specific humidities were similar to those in JASIN.

The normalized humidity spectra (Fig. 8) show considerable variations in magnitude resulting from meso-scale variability in the humidity field and this leads probably to the scatter in the normalized estimates at high frequencies. Allowing for a fall-off in response of the wet-bulb sensor at higher frequencies it appears that the inertial interval extends to values of natural frequency somewhat less than 1; a similar result was found by Phelps and Pond, and Miyake et al. (1970a).

5. CONCLUSIONS

The limited data which have been obtained suggest that the tethered balloon system in its present form can be used to measure fluxes of heat and moisture at heights up to a few hundred metres in winds below about 10 m s\(^{-1}\). Ship motion on the one occasion with stronger winds was found to produce sufficiently violent movements of the sensors attached to the balloon cable to prevent satisfactory functioning of the equipment. \(\text{Uw}\) correlations were much more strongly influenced by sensor motion than those for either \(\text{T}\) or \(\text{q}\) and although it was found possible to identify and remove the major spurious contributions to the measured momentum fluxes, the corrected (downward) values were on average somewhat smaller than expected.

Rapid decreases or even reversals of heat flux with height were found on three occasions with thermal instability at the surface but moisture fluxes varied less rapidly from their estimated surface values in these conditions. The measured value of \((1.8 \pm 0.7) \times 10^{-3}\) for the constant in the bulk-aerodynamic formulation of evaporation is in moderately good agreement with estimates by other investigators.

The technique is not suitable for obtaining reliable spectra of horizontal and vertical wind speed and (to a lesser degree) temperature because the main spurious contributions due to
ship motion affect regions with significant spectral density and so cannot be convincingly filtered out. However, the introduction of servo-stabilization of the tethering point of the balloon cable would enhance considerably the utility of this technique.

Thus a reduction of noise contributions to the fluxes by about an order (corresponding to a reduction of the vertical displacements of the tethering point by a factor of about 3) would probably be sufficient to obtain satisfactory momentum fluxes by the method outlined above but because of the greater high-frequency content of spectra a noise reduction nearer two orders of magnitude (vertical motion reduced by a factor of 10) would be required for their satisfactory measurement. Servo-stabilization would also allow proper operation in winds stronger than 10 m s$^{-1}$.

Acknowledgments

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References


1960 'Heat and water-vapour fluxes in air flowing southward over the western North Atlantic Ocean,' Ibid., 17, pp. 52-63.


Jones, J. I. P. 1961 'The measurement of turbulence near the ground,' Porton Technical Paper No. 786.


Pond, S., Phelps, G. T., Paquin, J. T., McBean, G. and Stewart, R. W.
1971a Unpublished – Meteorological Office, Met. O. 14 Turbulence and Diffusion Note No. 11.