

AN INVESTIGATION OF TURBULENCE STRUCTURE AND FLUXES
OVER THE SEA BY A TETHERED-BALLOON TECHNIQUE

by

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ABSTRACT

A description is given of a method which has been developed for the measurement of turbulent fluctuations of wind, temperature and humidity at heights up to a few hundred metres over the sea using instruments supported by a tethered balloon. Wave-induced motions of the ship from which the balloon was flown produced unwanted contributions to the measured fluctuations 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 water vapour: the calculated momentum fluxes appeared to be less reliable. Observations made in generally unstable conditions yielded a value of $(1.8 \pm 0.7) \times 10^{-3}$ for C_E in the bulk-aerodynamic formulation for evaporation ($E = \rho C_E (q_s - q_{s0}) (U_{10} - U_0)$). It appears that the buoyancy flux and hence the turbulence structure in the bulk of the boundary layer may depend much more on the vertical water-vapour flux than on the heat flux because the latter generally decreases more rapidly with increasing height than the former. The nature of the variation of the heat flux with height is explicable in terms of a convection process which is dominated by moisture. Momentum fluxes obtained from the same series of observations suggest a drag-coefficient (referred to the wind at 10m) around 1.1×10^{-3} but there is evidence that this was an underestimate since Uw -cospectral estimates were often found to be small or positive at frequencies around 3×10^{-2} Hz, probably a consequence of sensor motion. Estimates of the terms in the balance equation for turbulent fluctuations did not generally support the neglect of the flux divergence terms.

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NOTATION

a_k, b_k, A_k	Fourier coefficients
A	Constant in psychrometric equation
c_p	Specific heat of air
C_D	Drag coefficient
C_G	Geostrophic drag coefficient
C_H, C_E	Transfer coefficient in bulk-aerodynamic formulae for heat and moisture transport
e	Vapour pressure
E	Vertical moisture flux
E'	Turbulent kinetic energy
f	Reduced frequency ($= nz/U$) or Coriolis parameter
F	Attenuation of unweighted running-average filter
h	Boundary layer depth
H	Vertical heat flux
k	von Karman's constant
L	Monin-Oboukhov length
M	Magnitude of wind vector
n	Frequency
p	Pressure
q	Specific humidity
q_*	Scaling humidity ($= -E/\rho u_*$)
s	Averaging time
$S_{uu}(n), S_{ww}(n), S_{TT}(n), S_{qq}(n)$	Spectral estimates
$S_{uw}(n), S_{Tw}(n), S_{qw}(n)$	Cospectral estimates
T	Air temperature, record length
T_w	Wet bulb temperature
T_*	Scaling temperature ($= -H/\rho c_p u_*$)
u, v, w	Wind components referred to x-axis aligned along mean wind

u_g, v_g	Geostrophic wind component along and perpendicular to surface wind direction
U	Horizontal wind speed
U_o	Sea surface drift velocity
$\underline{V}$	Horizontal wind vector
$\underline{V_g}$	Geostrophic wind vector
x, y, z	Right-handed coordinate system, z-axis upwards
z_o	Roughness length
α	Geostrophic wind veer
ϵ	Rate of dissipation of turbulent kinetic energy
ρ	Air density
θ	Potential temperature
$\Delta \theta, \Delta q$	Sea-air differences of potential temperature and specific humidity
k	Wave number
$\sigma_u, \sigma_v, \sigma_w, \sigma_T, \sigma_q$	Standard deviation of turbulent fluctuations
ϕ	Wind inclination to horizontal
$\phi_m(z/L), \phi_n(z/L), \phi_E(z/L)$	Empirical functions used to allow for effect of non-neutral density stratification on vertical gradients of wind, temperature and humidity eg $\partial U / \partial z = u_* \phi_m(z/L) / k z$
$\psi_m(z/L), \psi_n(z/L), \psi_E(z/L)$	Empirical functions adjusting vertical profiles to diabatic conditions eg $U = u_* (\log(z/z_o) - \psi_n(z/L)) / k$
χ_T, χ_q	Rates of dissipation of variance of temperature and humidity fluctuations
τ	Vertical momentum flux
u_*	Friction velocity (= $(\tau/\rho)^{1/2}$)

CHAPTER 1

INTRODUCTION: ATMOSPHERIC TURBULENCE AND TURBULENT FLUXES

OVER THE SEA

In recent years there has been much interest in the processes by which properties such as heat and momentum are exchanged between the atmosphere and the underlying surface and redistributed by atmospheric turbulence. The demands of numerical modellers have insured that increasing attention be paid to this complex problem of the inter-relation between gross boundary-layer structure and turbulence but progress in this has been made mainly as a result of the drastic assumption that the processes involved can be considered (statistically) stationary. While such an approach is of value in initial investigations it inevitably must give way to treatments which include specifically diurnal and synoptic variability. It has not been demonstrated that the problem is tractable using purely theoretical treatments and suitable experiments designed to provide all necessary empirical parameters are so complicated and costly that it seems unlikely effective solutions will be found for a number of years.

It is reasonable to make a fairly clear distinction between boundary-layer structure over land and sea. Typically the first shows very large diurnal variability with turbulence structure controlled essentially by vertical heat transfers, but over the sea in contrast the diurnal cycle is often very weak and structure development may be dominated by the water-vapour flux. Logistically it is certainly less difficult to carry out a comprehensive boundary-layer study over land (though there is no intention here to imply that any such study is an easy undertaking) : that reported by Clarke (1970) is the most noteworthy but even so has many weaknesses, for example lack of direct measurements of vertical turbulent fluxes away from the surface, a failure to incorporate satisfactorily the effects of baroclinicity, and the uncertainties introduced by diurnal effects. The practical and immediate stimulus provided by the development of forecasting

models means that even more complex investigations will be undertaken and presumably most of the effects isolated and quantified eventually. On the other hand if such efforts were directed towards the parallel problem over the open sea progress might be more rapid because of the more uniform nature of the underlying surface and the greatly-reduced significance of diurnal variability.

Over both land and sea most investigations of atmospheric turbulence and vertical transfers have been concentrated on the lowest few metres, the so-called "constant-flux" layer, and studies over substantial fractions of the boundary-layer are rare. Similarity theory (Monin and Oboukhov 1954) can be used in the constant-flux layer to obtain simple relations between vertical gradients and vertical turbulent fluxes : thus the wind profile is expressed in gradient form by

$$\partial U / \partial z = (u_* / k z) \phi_m (z/L) \quad (1)$$

with ϕ_m an empirical function introduced to allow for departures from neutral density stratification and now reasonably well defined in value as a result of numerous experimental studies over land (eg Dyer and Hicks 1970, Webb 1970). In neutral conditions ($\phi_m = 1$) eq (1) can be integrated to give, specifically over the sea,

$$U - U_0 = (u_* / k) \log(z/z_0) \quad (2)$$

where U_0 is the surface drift velocity (about $0.04U$, Keulegan 1951). Thus the downward flux of momentum is

$$\begin{aligned} \tau = \rho u_*^2 &= \rho k^2 (U - U_0)^2 / (\log(z/z_0))^2 \\ &= \rho C_D (U - U_0)^2 \\ (C_D &= (k / \log(z/z_0))^2) \end{aligned} \quad (3)$$

C_D is conventionally used in conjunction with the wind speed at 10m height. Observationally there is a clear correlation between wind speed and the state of sea and intuitively therefore it might be expected that z_0 and hence C_D would increase with wind speed. However, despite numerous experimental investigations any relations between C_D and wind speed are still

obscure. Earlier studies supported the intuitive idea, for example in 1958 Sheppard suggested the relation

$$C_D = (0.80 + 0.114 U_{10}) \times 10^{-3}$$

but some recent measurements appear to favour C_D being virtually constant for winds up to about 16 ms^{-1} (Brooks and Krugermeyer 1970, Smith 1970) with a value around 1.3×10^{-3} . On the other hand Kitaigorodskii (1970) concluded that complex relationships exist between C_D , wind speed and state of sea. Substantially more data will be required to resolve these conflicting points of view, but in the meantime it is thought that a constant value for C_D of 1.3×10^{-3} should usually be accurate to within about $\pm 25\%$ for winds up to near gale force. There appear to be no reliable data at higher speeds.

The effects of density stratification on size of drag coefficient decrease rapidly with increasing wind speed but in light winds there may be significant departures from neutral values. Figure 1 (taken from Thompson 1971a, incorporated in this thesis as paper 1 in the Appendix) shows clearly that in unstable conditions with typical magnitudes for the differences $\Delta\theta$ and Δq measured between sea surface and 10 m height (usually less than 5 deg C and 5×10^{-3} respectively) it is only in light winds that C_D will vary more than a few percent from its neutral value: a similar behaviour is found for stable density stratification also (Deardorff 1968). Similarity theory leads to the following expressions for the vertical gradients of temperature and specific humidity:

$$\partial\theta/\partial z = (T_*/kz) \phi_\theta(z/L) \quad (4)$$

$$\partial q/\partial z = (q_*/kz) \phi_q(z/L) \quad (5)$$

where $T_* = -H/e c_p u_*$ and $q_* = -E/e u_*$. The variations in magnitude of ϕ_θ , ϕ_θ and ϕ_q with stability are now reasonably well-known as a result of numerous experimental studies over land (eg Dyer and Hicks 1970, Webb 1970). In particular ϕ_θ and ϕ_q appear to be equal in size which suggests a similarity in the mechanisms of turbulent transport of heat and water vapour. However

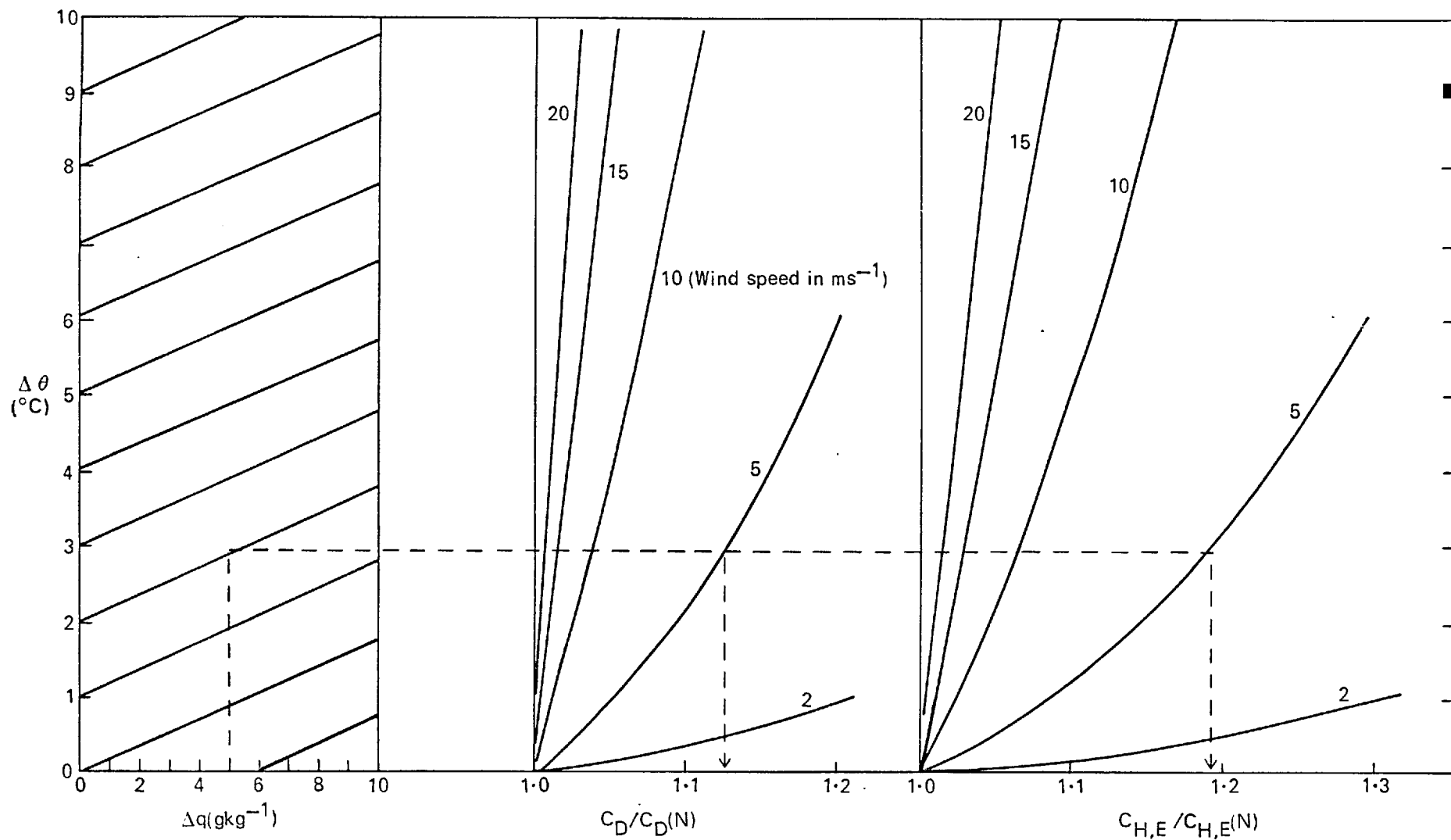


Figure 1 Estimated variations of transfer coefficients with temperature and humidity differences measured between surface and 10m. (The dashed lines demonstrate use of the diagram for $\Delta\theta = 2^{\circ}\text{C}$, $\Delta q = 5 \times 10^{-3}$ and $U = 5\text{ms}^{-1}$)

they are systematically smaller than ϕ_m in unstable conditions, showing that the process of vertical momentum transport is less efficient than those for heat or water vapour in these circumstances. In stable conditions the evidence is for near-equality in the $\phi's$. In near-neutral conditions $\phi_H, \phi_E \sim 1$ and equations (4) and (5) can be integrated to give

$$\theta - \theta_0 = (T_*/k) \log(z/z_T) \quad (6)$$

$$q - q_0 = (q_*/k) \log(z/z_q) \quad (7)$$

or, using equation (2)

$$H = \rho c_p C_D (\theta_0 - \theta) (U - U_0) (\log(z/z_0)) / \log(z/z_T) \quad (8)$$

$$E = \rho C_D (q_0 - q) (U - U_0) (\log(z/z_0)) / \log(z/z_q) \quad (9)$$

Equations (8) and (9) are often expressed in simplified form (the familiar bulk-aerodynamic formulae);

$$H = \rho c_p C_H \Delta\theta (U - U_0) \quad (8a)$$

$$E = \rho C_E \Delta q (U - U_0) \quad (9a)$$

where $\Delta\theta$ and Δq are the sea-air differences obtained using a bulk water temperature. Chamberlain (1968) from wind tunnel studies obtained values of the ratio $z_0/z_{T,q}$ for different wind speeds: the ratio increased with wind speed being around 2 at 10ms^{-1} and 3 at 20ms^{-1} and this gave values for C_H and C_E a few percent smaller than C_D (Thompson 1971a). Also the surface temperature is different usually from the bulk water temperature (Hasse 1971) and taking this into account reduces the coefficients still further, typically by a few percent. Calculated variations of the coefficients with unstable density stratification are larger than those for the drag coefficient (figure 1) and are in broad agreement with experimentally-determined variations given by Deardorff (1968) but again are insignificant except in light winds.

It should be pointed out that Stewart (1961) suggested that a significant fraction of the vertical momentum flux is likely to be transported to the sea to maintain and generate waves, the mechanism depending on a

suitable phasing of turbulent pressure fluctuations and wave height, and some experimental support for this has been supplied by Dobson (1971). The effective level at which momentum is given up to the waves is near that where the wind speed is equal to the phase speed of the waves and so the downward flux of momentum must decrease as the surface is approached. As a result there would be a reduction of turbulence at these low heights and vertical diffusion coefficients for heat and vapour transfer would then be less than measurements at greater heights might suggest. The consequence would be increases in the vertical gradients of heat and water vapour near the surface. It is unlikely therefore that the temperature or water vapour differences across this layer where the stress changes significantly with height would be given by the integrated forms of eq. (4) and (5) and hence the ratios $C_{H,E}/C_D$ might be changed by the effect.

Direct measurements of evaporation in a wind-tunnel by Chamberlain (1968) gave a value for C_E of 1.3×10^{-3} while Robinson (1966) by assuming $C_H = C_E$ and using a thermal budget method in conjunction with routine weather-ship data found a value for the coefficients of about 1.2×10^{-3} . Hasse (1968) measured vertical heat fluxes over the sea by an eddy-correlation method and concluded that $C_H \approx 1.0 \times 10^{-3}$. In general the recent data (obtained mainly in near-neutral conditions) broadly support the idea that the exchange coefficients for momentum, heat and water vapour are very similar in magnitude.

In strong winds spray droplets will be carried away from the surface by turbulent upcurrents and their subsequent evaporation will increase the upward vapour flux but reduce the sensible heat flux. Monahan (1968) showed that an abrupt increase of droplet production occurred at a wind speed of about 11 ms^{-1} , though he believed that most of this spray was produced by the jet-droplet process. As wind speeds increase, increasing numbers of spray droplets are formed by the destruction of wave crests and

it is likely that this is the dominant process which leads to the rapid decrease of visibility which is observed when winds increase to above about 20 ms^{-1} . Calculations using Monahan's data by Hall (1972) suggested that with typical sea-air humidity differences the evaporation of jet-droplet spray was unlikely to increase vertical water vapour fluxes by more than one or two percent but, bearing in mind that in unstable conditions the flux of latent heat is usually several times greater than that for sensible heat, this would produce a significant reduction in the upward sensible heat flux. In very strong winds with drops also being produced directly from breaking waves the enhanced evaporation from these could in unstable conditions reverse the flow of sensible heat except very close to the surface. The process would lead also to a decrease of temperature near the surface and hence would modify considerably the structure of the surface layer. Winds greater than 20 ms^{-1} occur frequently in association with middle-latitude depressions and hence the process just described should be of more than passing interest to those developing numerical models of the atmosphere.

Over the whole depth of the oceanic boundary layer the role of the vertical flux of water vapour in determining the turbulence structure has often been underestimated in the past (the part played by the vapour flux will be treated in detail in section 7 and will be discussed only briefly here). Water vapour being lighter than air will contribute to the buoyancy flux when transported vertically; near the surface the global average Bowen ratio is about 0.1 (Sellers 1965) and so usually this contribution will be of similar size to that due to the heat flux (Lumley and Panofsky 1964). Evaporation from the surface also increases the "thickness" of the overlying layer (i.e. the depth of the layer bounded by fixed constant-pressure surfaces), not apparently taken into account when surface exchanges were introduced into the Meteorological Office 10-level numerical-forecasting model (Gadd and Keers 1970). At greater

heights the release of latent heat carried upwards by the vapour flux may modify profoundly the turbulence structure. Despite these complications the mixed layer over the sea is perhaps simpler than the corresponding region over the land because of its much smaller temporal variability. As already noted the diurnal variability is often small enough to be neglected - usually impossible over land - and the synoptic range of density stratification is typically rather small as well. It is perhaps interesting in this context to use results from Clarke (1970) and Findlater et al. (1966) to find the change with stability of the drag coefficient (C_G) expressed in terms of the "frictionless" wind (or the wind at 900mb in the latter case). Over land there was a 50-fold variation in C_G , all of which could result from diurnal stability changes: over the sea, after correcting as far as possible Findlater et al.'s "surface" wind to a height of 10m and adjusting C_D for the likely range of stability as explained earlier, C_G varied by less than a factor of 10, and the effects of diurnal change would probably be undetectable in most cases.

Typical sensible-heat fluxes over the sea are small, for example over the Trade Wind region values are around 10 W m^{-2} (Sellers 1965), and are often insufficient to compensate for radiational cooling in the mixed layer. In higher latitudes over restricted areas where usually warm ocean currents lie close to cold continents, the sensible heat fluxes are often much larger, and with Bowen ratios around 1 the thermal and turbulent structure of the boundary layer must approach more closely that observed over land by day. In general however over the majority of the oceanic areas the part played by sensible heat transfer in determining boundary-layer structure is likely to be small.

CHAPTER 2

REVIEW OF TECHNIQUES FOR MEASURING TURBULENCE AND TURBULENT FLUXES OVER THE SEA.

2.1 INTRODUCTION

The semi-empirical similarity theory of Monin and Oboukhov for the constant-flux layer provides in principle a simple basis for estimating vertical turbulent fluxes from measurements of mean profiles of wind, temperature and specific humidity, but the depth of this layer may be only a small proportion of the turbulent boundary layer. There have been a number of recent attempts (conveniently summarized by Smith 1968) to extend similarity theory to the whole of the mixed layer but, as in the constant-flux case, empirical constants have to be found experimentally (their values are not yet accurately determined) and baroclinicity and diurnal or synoptic variability may invalidate the method. Hence these techniques for estimating vertical fluxes throughout the mixed layer, based on measuring "external" parameters such as geostrophic wind and temperature and humidity differences between surface and the top of the layer are not yet satisfactory. Attempts to estimate fluxes in regions away from the near-surface layer which involve measurements of vertical gradient used in conjunction with diffusion coefficients at present have little plausibility because of uncertainty about the size of the coefficients. In general therefore the profile techniques are restricted usually to a small part of the boundary layer and fluxes at greater heights have to be estimated from direct measurements (an exception is the geostrophic-departure method described in 2.3 below which is a profile-based technique that can be applied to the whole of the boundary-layer).

2.2 NEAR-SURFACE MEASUREMENTS

The majority of the experimental studies over the sea have been carried out in coastal waters using instruments mounted on platforms, towers or masts rigidly attached to the sea bottom (eg. Weiler and Burling 1967), but the shallow water and proximity to land make it doubtful whether the results are reasonably representative of conditions over the open sea. (Dyer and Hicks (1971) did obtain turbulence data from instruments attached to an offshore gas rig in comparatively deep water (60m) and calculated vertical fluxes of heat and momentum but in this case the large bulk of the rig must have introduced some disturbance to the airflow past the instrument). In general, representative near-surface measurements of turbulent fluxes over the open sea have to be made from floating platforms. The main techniques used have been either to measure vertical profiles of mean wind, temperature and humidity and then use eq. (1), (4) and (5) in conjunction with likely empirical values for the ϕ 's or alternatively to measure the turbulent fluctuations of horizontal and vertical wind speed, temperature and specific humidity and apply the following relations to the data

$$\tau = \rho \langle U_w \rangle \quad (10)$$

$$H = \rho c_p \langle T_w \rangle \quad (11)$$

$$E = \rho \langle q_w \rangle \quad (12)$$

$\langle w \rangle$ is of course equal to zero over the sea. The profile-measuring method is probably the more attractive because it needs somewhat less sophisticated instrumentation but because of the small vertical gradients which are observed over the sea the measurements have to be made with very great accuracy; also the values to be taken for the ϕ 's are still open to some dispute. The effects of buoy-motion on wind stress measured in this way were investigated by Pond (1968) who estimated that unless the buoy was unusually unstable the resulting errors would normally be a few percent at most. In recent years developments in instrumentation have resulted in the eddy-correlation

technique (fundamentally the more satisfactory) becoming more often used over the sea. Unfortunately buoy motions are likely to produce spurious correlations which will affect the measured momentum fluxes in particular, though large oscillating tilts (of the order of $\pm 15^\circ$) may be tolerated before errors in stress exceed 15%. However, errors due to a mean tilt of the instrumentation can be very serious, typical values being close to 10% for each degree of tilt. The eddy-correlation technique is therefore best suited to measurements made from rather stable platforms such as the "Flip" spar buoy (though the bulk of these above the waterline may produce significant disturbances to the airflow). An ingenious compromise has been to use a small, easily-handled buoy with servo-stabilization to maintain the vertical orientation of the sensors, with extra instrumentation to measure accelerations and any residual tilts (Hasse 1968). In strong winds and high seas the spar-buoy's great vertical stability can become an embarrassment however because of the risk of instrumentation becoming periodically submerged. In these conditions the amount of spray makes it virtually impossible to use the eddy-correlation technique and even the flux-gradient method breaks down for temperature and humidity measurements because of contamination of the sensors. (Attempts have been made to measure vertical profiles using instruments attached to a ship (eg Deacon et al 1956) and the eddy-correlation technique has also been used on a ship (Mitsuta et al 1969) but it is difficult to measure and eliminate with certainty the disturbing effect of the vessel).

The instruments normally used for profile measurements are cup-anemometers and wet and dry resistance-thermometers or thermocouples, but there has been more variety in the sensors used to measure turbulent fluctuations. Here the wind components are obtained usually with hot-wire or sonic anemometers (eg Weiler and Burling 1967, Miyake et al 1970a): thrust anemometers have also been used successfully from a stable platform (Smith 1970). Temperature

fluctuations are sensed acoustically (Miyake et al 1970a) or with fine-wire resistance thermometers (Hasse 1968). Until recently there has been no compact humidity sensor with a response sufficiently rapid to sample properly the fluctuations contributing significantly to the q-spectrum, but the Lyman- α device now appears to be suitable for this purpose (eg. Miyake et al 1970b). The quartz-crystal hygrometer described by Hicks and Goodman (1971) is, almost certainly, too sensitive to contamination to be used successfully at sea except in light winds.

The frequency response required if the sensors are to sample the majority of the high-frequency contributions to the variances falls off (roughly) inversely as the height above the surface (Busch and Panofsky 1968) and depends slightly on stability. In general the majority of the variance of wind and humidity fluctuations appears to be sampled if the sensors have a cut-off at a reduced frequency $f(=nz/U)$ higher than about 5, but the temperature spectrum may extend to somewhat higher frequencies (Miyake et al 1970 a,b). Turbulence contributions at these frequencies are incoherent and make negligible contributions to the cospectra which therefore have lower cut-off frequencies: Miyake et al suggest about $f=1$ for momentum and water vapour and $f=4$ for the heat flux. The cut-offs on the low-frequency side of the spectra and cospectra will determine minimum sampling periods for statistical reliability. There is an implied assumption here that a spectral gap exists or that micro- and mesoscale processes can be clearly separated, but this is not likely to be the case in general and in fact only the vertical velocity spectrum falls off rapidly at lower frequencies. Fortunately this produces cut-offs for the fluxes which are reasonably-well defined. The contributions to the heat flux appear to fall away more rapidly at lower frequencies than those for momentum and water vapour, but in all cases they seem to be negligible at reduced frequencies below 10^{-3} . Thus at a height of 2m in a 5ms^{-1} wind a minimum sampling period of about 7 minutes would be required, but for adequate statistical reliability would have to be several times longer.

2.3

MEASUREMENTS REMOTE FROM THE SURFACE

Most studies of turbulence at heights more than a few metres above the sea surface have been made using instruments attached to aircraft and the technique has now been developed sufficiently to produce apparently reliable data for the calculation of turbulent fluxes by the eddy-correlation method (Miyake et al 1970b). The instrumentation is complex because it must include accelerometers and gyroscopes for monitoring aircraft motion, and the method is not suitable for investigations remote from land unless large and expensive aircraft are available. Also unless more than one aircraft is used it is not possible to make the necessary simultaneous observations at different heights which are required if the detailed vertical structure of turbulence is being studied.

Tethered balloons have been used extensively over land to support turbulence-measuring instruments (eg. Jones and Butler 1958, Smith 1961, Thompson 1966) but reported uses over the sea are rare. Vasilchenko et al (1962) refer briefly to observations of "gust velocities" in a report which includes references to other Russian papers describing substantial series of measurements of (presumably) mean meteorological parameters using tethered balloons flown from ships. More recently a balloon-supported instrument for measuring turbulent fluctuations (Yokoyama 1969) has been used in a short series of observations at sea (Yokoyama et al 1969). The present writer is not aware however of any reports in the open literature describing the measurement of turbulent fluxes by balloon-supported equipment. Obviously there are difficulties in using this technique at sea but it is hoped that it will be made clear later in this thesis that these can be overcome to a large extent. The technique has the great advantage that it can be used as readily remote from land as in coastal waters.

The method of geostrophic-departure for determining momentum fluxes has been employed on a few occasions over the sea, the observations being made

from small islands. If accelerations are ignored, the equation of horizontal motion reduces to

$$f \underline{k} \times \underline{V} + (1/\rho) \nabla p = -(1/\rho) \partial \tau / \partial z \quad (13)$$

or

$$f \underline{k} \times (\underline{V} - \underline{V}_g) = -(1/\rho) \partial \tau / \partial z \quad (14)$$

Charnock et al (1956) used the special character of the vertical structure of the Trades (where a thermal wind is nearly opposed to the geostrophic wind) to arrive at surface values of the momentum flux as well as variations with height by integrating equation (14) numerically, but the technique was found to be much less useful in the deep baroclinic Westerlies (Sheppard et al 1952).

McAllister et al (1971) have reported measurements using a novel technique to determine the total vertical heat flux at the surface. It makes use of the sea's differing absorption coefficients for infra-red radiation of different wavelengths and hence the different depths which an infrared thermometer "sees" as the wavelength is varied : two suitable wavelengths are chosen which restrict the effective depths to the molecular conduction region at the sea surface, and hence the total flux is easily calculated from the measured temperatures. The system has been designed for use from aircraft but has to be deployed within a few tens of meters of the surface and breaks down therefore in strong winds not only because of disruption and foaming of the sea surface but also through spray contamination.

CHAPTER 3.

A TURBULENCE-MEASURING SYSTEM USING A TETHERED BALLOON TO SUPPORT THE SENSORS

3.1 INTRODUCTION

Experience has been gained within the Meteorological Office over a number of years in the measurement of atmospheric turbulence by instruments attached to large tethered balloons flown over land. However, when it was decided to extend the experimental programme to include measurements over the open sea, it was found necessary to redesign completely the overall system because clearly it was not feasible to use the existing tethered balloons (which were about 25 metres long) from research ships. The stimulus to expand the boundary-layer programme in this direction had come from a Royal Society sponsored air/sea interaction project which was planned to take place a few hundred kilometres west of Ireland and it was this project which determined partly the form in which the instrumentation was to be developed. For example it was anticipated that there would be upper wind-finding equipment on board the ships so no attempt was made to develop an airborne sensor which would measure wind direction.

The parameters which it was decided to measure were the horizontal and vertical wind speed, temperature and humidity which in principle would allow the calculation of vertical fluxes of momentum, heat and water vapour (the vertical flux of horizontal momentum is conventionally calculated as $\rho \langle uw \rangle$ whereas in the present case it was only possible to derive the approximation $\rho \langle U w \rangle$. Pasquill and Tyldesley (1967) showed that

$$\langle U w \rangle \simeq \langle u w \rangle (1 + \langle v^2 \rangle / 2 \langle u^2 \rangle)$$

The magnitude of the momentum flux at any level assuming alignment of the x-axis along the mean wind direction is given by

$$\rho (\langle u w \rangle^2 + \langle v w \rangle^2)^{1/2} \simeq \rho \langle u w \rangle (1 + \langle v w \rangle^2 / 2 \langle u w \rangle^2)$$

and the error in approximating this by $\rho \langle U w \rangle$ is therefore

$$\rho \langle u w \rangle (\langle v w \rangle^2 / 2 \langle u w \rangle^2 - \langle v^2 \rangle / 2 \langle u^2 \rangle)$$

McBean (1971) found $\langle v^2 \rangle / u_*^2 \approx 5$ in near-neutral conditions and using the rough approximation $u_*^2 \sim 10^3 \langle u^2 \rangle$ the second term inside the brackets is therefore negligible at least near the ground. Swinbank's (1970) analysis of the Leipzig wind profile 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 the 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° (Mendenhall 1967) compared with around 25° for the Leipzig profile and the approximation $\tau = \tau_{xz} = \rho \langle U w \rangle$ is probably accurate to a few percent at heights up to several hundred metres, but in trade wind areas where the veer can be as large as 30° over the sea (Charnock et al 1956) the errors may be substantially larger.

Initially it was considered likely that great difficulties would be experienced in applying the eddy-correlation method because of sensor motions introduced by wave-induced oscillations of the tethering point of the balloon cable. It was thought likely that it might be only feasible to estimate fluxes indirectly, by measuring certain terms in the balance equations for the turbulence parameters using data obtained at sufficiently high frequencies to be unaffected by ship motion. Since then further information from experiments over land (eg. Lenschow 1970) supports the conclusion that usually the divergence terms in the balance equations cannot be neglected in comparison with production terms : the former are triple correlations and therefore measured values are at least as sensitive to ship motion as the double correlations used to calculate the eddy fluxes and so the high-frequency method is of limited use (this point is taken up in more detail in section 7. 8).

The sensor motion is due in the main to periodic vertical movements of the balloon cable's tethering point at close to the ship's natural

frequency. The amplitude of the oscillations varies markedly with time due to differences between the frequency of the wave field which excites the motion, and that of the ship. The vertical displacements produce alternate tightening and slackening of the balloon cable which, unless the sensors are mounted close to the surface or the balloon, results in predominantly horizontal motion of the sensors more or less in the plane containing the wind vector. The spurious velocities which are thus introduced can be quite large. Figure 2 shows a record of vector wind speed obtained in the North-East Trades: the speed variations due to sensor motion have a period around 5 seconds and amplitude up to about $\pm 1 \text{ ms}^{-1}$. In this instance the balloon was about 30 m above the sensors but in cases where the latter were mounted nearer the midpoint of the cable even larger spurious velocities were measured. Coupled with the sensor motions are substantial nearly-horizontal accelerations which, if the device measuring vertical velocity is gravity-orientated by a pendulum, will produce very severe contamination of this velocity signal. Another effect of the periodic quasi-horizontal displacements is to vitiate to some extent the Taylor transformation ($\kappa = 2\pi n / \langle U \rangle$), chiefly in the frequency range around or above that of the sensor motion: then if the spectral density of the turbulent fluctuations varies rapidly over this frequency range the result will be distortion of the measured spectrum (the w, T and q spectra will be affected in this way even if the sensor motion is purely horizontal). The maximum errors from this cause occur at high frequencies and appear as a broadening over a frequency range $\pm \Delta n$ of any discrete feature in the spectrum at frequency n , where $\Delta n/n$ is equal to the maximum fractional variation of wind speed due to quasi-horizontal sensor motion.

Tethered balloons attached to a stationary mooring over land show a characteristic periodic migration across wind, initiated presumably by changes in wind direction. The effect of this motion on measured momentum fluxes was

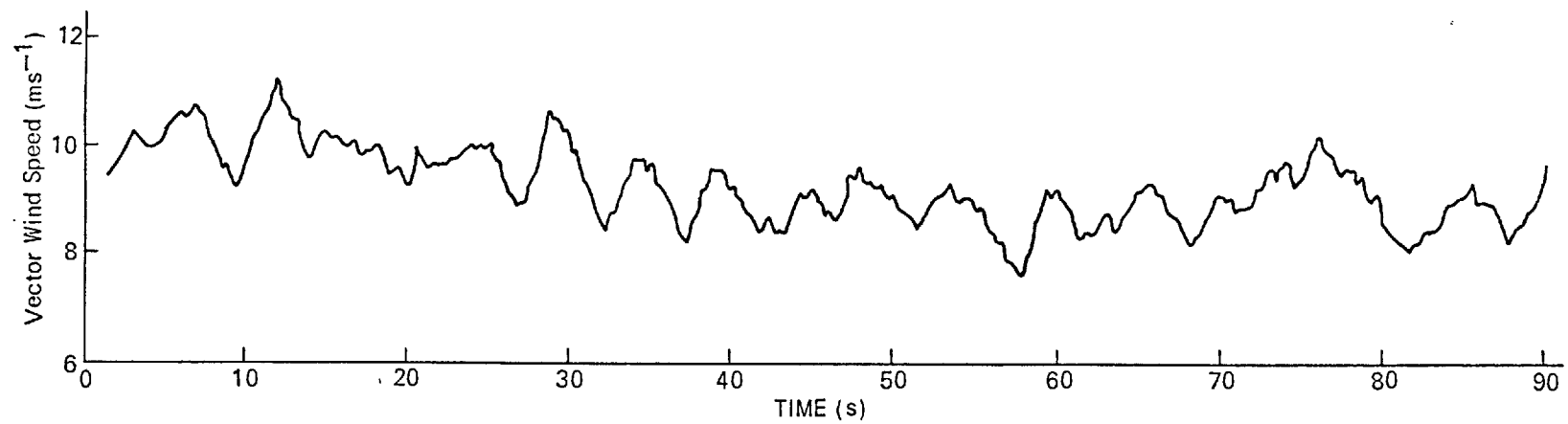


Figure 2 Vector wind speed, 180 m, 7/2/69 (~ 1300 Z) (Measured in NE'ly Trades at 10.8N, 36.1 W)

investigated by Thompson (1969, incorporated in this thesis as paper 2 in the Appendix) who used double-theodolites to track the motion of turbulence sensors attached to the balloon cable. In these experiments the sensors were at a height of about 600m (the balloon was 70m higher) and horizontal and vertical sensor velocities in excess of 1ms^{-1} and 0.2ms^{-1} respectively were measured frequently in convective conditions. The wind direction was not measured by the balloon-borne instruments and so only the vertical velocities could be corrected for sensor motion. However if it is assumed that the vertical displacements are largely uncorrelated with vertical air velocities then the effects of sensor motion on the measured fluxes may be removed even in the absence of horizontal velocity data by correcting the vertical speed only. This correction procedure was probably a reasonable first-order approximation because in the experiments the balloon cable had very little sag so that the majority of the vertical sensor motion was probably associated with the balloon swinging across wind. The results appear in Table 1. The correction is

Table 1 Influence of tethered-balloon motion over land on measured momentum flux

Date	Time (z)	Wind Speed (ms^{-1})	Momentum fluxes (Nm^{-2})	
			uncorrected	corrected for sensor motion
25/5/68	1451-1517	1.2	-0.032	-0.033
5/6/68	0945-1013	2.0	-0.092	-0.091
	1515-1557	5.1	0.095	0.112
21/8/68	0932-1019	4.6	0.187	0.237
	1417-1450	5.0	-0.004	0.064
22/8/68	1415-1440	2.7	0.253	0.279
27/8/68	1338-1415	9.8	-0.135	-0.062

seen to be small in winds less than 3ms^{-1} , but is significant at higher speeds, the average value in these cases being 0.05Nm^{-2} at a mean speed of 6ms^{-1} . The size of the correction decreases as the surface is approached

(if the turbulence probe's height is altered but the balloon's altitude is left unchanged), but above 200m or so will be of the same order as the true momentum flux (which would be typically around -0.2 Nm^{-2} at the surface over grassland ($z_0 \sim 3\text{cm}$) with a 10m wind of 6ms^{-1} : this follows from eq.(2) assuming near-neutral conditions).

Readings and Butler (1971) made simultaneous measurements over land of vertical momentum flux using two instruments, one attached at a height of 43m to the cable of a captive balloon, the other at the same height on an adjacent mast: a number of runs were done with the balloon flown at levels between 60m and 600m above the probe. Mean percentage differences between fluxes from the mast and the balloon cable are given in Table 1 (a). The approximate agreement between the measured fluxes when the balloon was at 600m was a direct consequence of the way in which the two instruments were levelled (the initial run each day was carried out with the balloon at 600m and after assuming that $\langle w \rangle = 0$ at the instrument level the measured fluxes were forced to agree by rotating the reference axis of the balloon-borne instrument: these axis orientations were used for the remainder of the runs on that day). It is seen that mean errors and scatter were small

Table 1(a) Percentage differences between momentum fluxes from
a mast and a tethered balloon at 43m

Balloon height above probe (m)	$\langle Uw \rangle$ mast	-	$\langle Uw \rangle$ balloon
			(%)
60	51	$\pm$	48
150	2	$\pm$	22
300	-6	$\pm$	11
600	3	$\pm$	10

with the balloon at the two largest heights, but at the lowest height the scatter was very large though there was a substantial systematic error on

average (in the same sense as found by Thompson (1969)).

Over the sea it is only possible to monitor sensor motion by devices such as accelerometers and gyroscopes attached to the instrument package but this introduces such complexity and weight penalty that it is not really practicable. However in some preliminary tests of a tethered balloon which were carried out by the author in convective conditions a few kilometers offshore from the coast of Southern Spain it was found that the balloon flew very stably with little evidence of the characteristic motions observed over land: subsequent studies in the North-East Trades and near OWS "J" have confirmed this behaviour. It appears then that apart from the direct influence of ship motion the tethered balloon technique when applied to turbulence measurements over the sea may be suitable for obtaining vertical turbulent fluxes.

Figures 3 and 4 show "universal" spectra and cospectra over water, based on the data of Miyake et al (1970a). Also marked on the diagram is the reduced frequency corresponding to the ship's characteristic period of motion: it has been assumed for this purpose that $z/\langle U \rangle = 20$, eg. the graphs refer to a height of 100 m in a 5 ms^{-1} wind. In practice the noise contribution due to sensor motion will spread over a finite range of frequencies and this extends well into regions with significant spectral density unless measurements are made far above the surface or in very light winds. In the case of the cospectra, even for the heat flux the influence of sensor motion will probably be confined to regions where genuine turbulent contributions are relatively small, and judicious smoothing may well be sufficient to eliminate much of the noise. "Universal" humidity spectra and q_w -cospectra obtained by Phelps and Pond (1971) are rather similar to u -spectra and uw -cospectra respectively, so similar conclusions may be made about the effects of ship motion on these also. It should be pointed out that reduced-frequency scaling eventually breaks down when the height above surface is

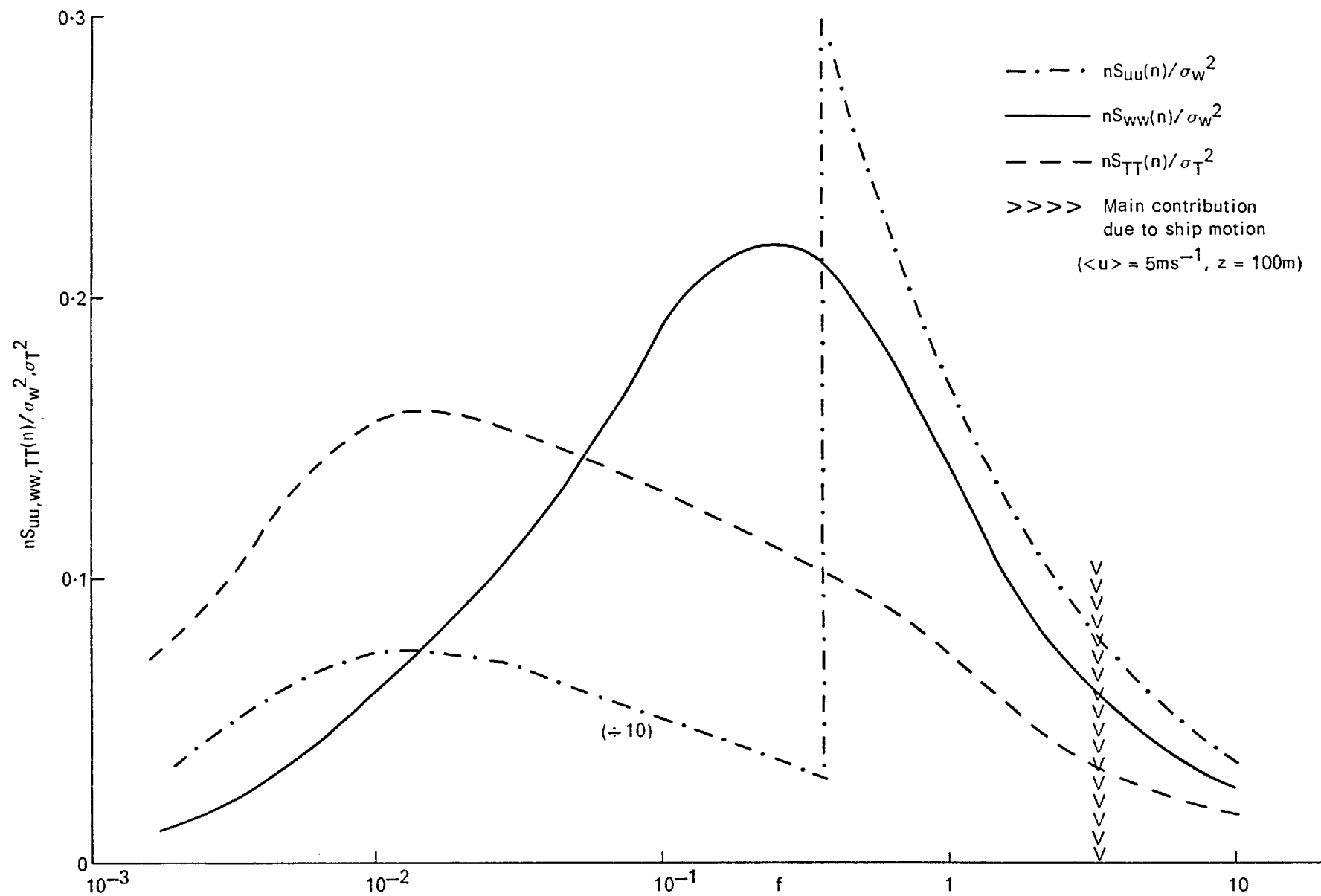


Figure 3 Universal spectra over the sea

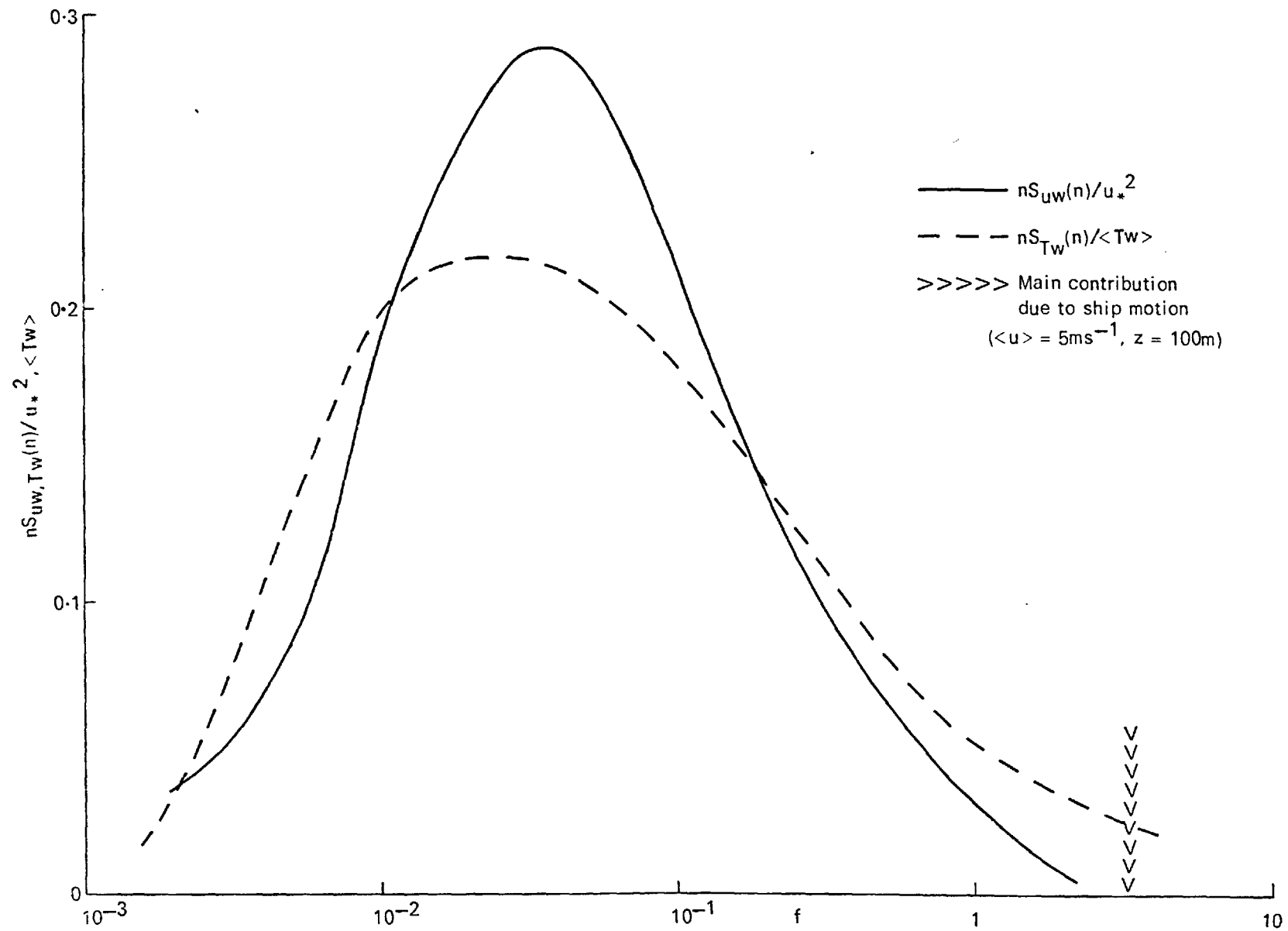


Figure 4 Universal cospectra over the sea

sufficiently large, though the scale-lengths of turbulence continue increasing up to heights of several hundred metres: in unstable conditions, i.e. those with the greatest meteorological significance, Busch and Panofsky (1968) found f increasing above about 50m height, with observations at Cedar Hill (Kaimal 1966) giving $f_{320m} / f_{46m} \sim 2$ at the peak and about 1.5 at the high frequency end of the vertical velocity spectrum. The conclusions above about the noise contributions from ship motion are substantially unaltered by this.

A final general point about the tethered balloon technique when used at sea concerns the lowest height at which turbulence measurements may be made without significant influences from the superstructure of the ship. For practical reasons it is usual to fly tethered balloons from the stern of ships, and with the superstructure usually extending up to about 20m above the surface even in the case of typical research vessels it appears unlikely that reliable measurements can be made as close to the surface as say 50m when headed into wind: in any event funnel smoke often spoils data at these low heights. When lying-to the effective height of the superstructure is rather less usually but even so there must be some doubt about the validity of data obtained at heights much less than say 100m.

3.2

BALLOON AND ANCILLARY EQUIPMENT

The tethered balloon technique is of limited value unless payloads of at least 10Kg can be carried but this leads to problems at sea because there are definite limitations here on the size of balloon which can be used. The factor controlling the balloon size is the unobstructed area available for inflation but suitable space is very restricted on most ships: fortunately the H-class survey vessels which have been used in the present studies have a large helicopter flight-deck at the stern and this has allowed the use of a balloon about 10m long. This was constructed from proofed-nylon fabric and was of conventional design with three air-inflated fins. It incorporated an air-filled chamber separated from the gas-filled cavity by a flexible ballonet, so that any loss of gas and hence change of shape was compensated automatically by intake of more air. The gas used was helium and the volume about 80m^{-3} giving a free lift around 50Kg. The tethering cable was steel of 1000Kg breaking strain weighing about 0.05 Kg m^{-1} . The system was designed to withstand winds up to at least 25ms^{-1} and has been used without damage in squalls of at least 20ms^{-1} . In light winds a 500 Kg cable could be used safely to achieve greater heights or allow a larger payload. With a payload of 5 Kg altitudes in excess of 400m have been reached easily using the heavier cable. The cable was controlled by a variable-speed reversible electric winch, giving speeds between about 0.1 and 1 ms^{-1} on both payout and recovery. For practical reasons the cable was led from the winch through a block mounted at the stern end of the flight deck. In order therefore to facilitate recovery of the balloon for topping-up a length of nylon rope about 30m long was attached to the balloon and normally tied at its lower end to the balloon cable: in recover this end was untied and the rope then used to pull the balloon forward to bring handling lines within reach. Launching of the balloon was always carried out without difficulty but recovery was potentially more hazardous because violent yawing sometimes occurred in winds in excess of a few metres per second when close to the deck, initiated presumably by ship-induced turbulence.

3.3

TURBULENCE-SENSING SYSTEM

If turbulence measurements are made no closer to the surface than a few tens of metres then the sensor response times required to resolve most of the higher-frequency contributions to the cospectra can be relatively long. The sensors can therefore be comparatively simple and robust: for example a cup-anemometer may be used for measuring wind speed. The present system (shown in figure 5 and diagrammatically in figure 6) did not measure the horizontal and vertical wind speeds directly but rather the magnitude of the wind vector and its inclination to the horizontal, from which the required components could be deduced. The inclinometer consisted of two heated platinum wires mounted in a double "V" configuration with equal and opposite inclinations to the horizontal (Jones 1961): by using the double rather than single "V" shape the sensitivity to azimuth fluctuations was reduced. The wire diameter was $13\mu\text{m}$ and the time-constant less than 0.015s in winds of a few metres per second (Taylor 1963). The wires were placed in series with fixed resistors (this compensated to some extent for sensitivity changes resulting from wind speed variations) and connected in the opposite arms of a simple Wheatstone bridge (figure 7(a)). Typical variations of the sensitivity with wind speed were about $\pm 3\%$ for the range 3 to 15 ms^{-1} . Outputs normally departed from linearity by less than $\pm 2\%$ for inclinations of $\pm 20^\circ$. The output sensitivity was adjustable to $\pm 30^\circ$ or $\pm 45^\circ$ full scale. The sensor was vertically orientated by mounting it on a short pendulum (period around 0.7s) whose oscillations were damped by an oil-filled dashpot.

The anemometer (based on a design by D Tribble of Imperial College) used a three-cup rotor running in jewelled bearings and mounted with its axis horizontal. The cups were pressed from plastic sheet and increased in thickness from tips to base to give additional strength. Wind tunnel tests showed a distance-constant of about 0.8m and a satisfactory performance at speeds up to at least 18ms^{-1} . The rate of rotation was sensed photoelectrically with a disc attached to the rotor periodically interrupting light shining on a phototransistor connected in a pulse circuit (figure 7(b)):

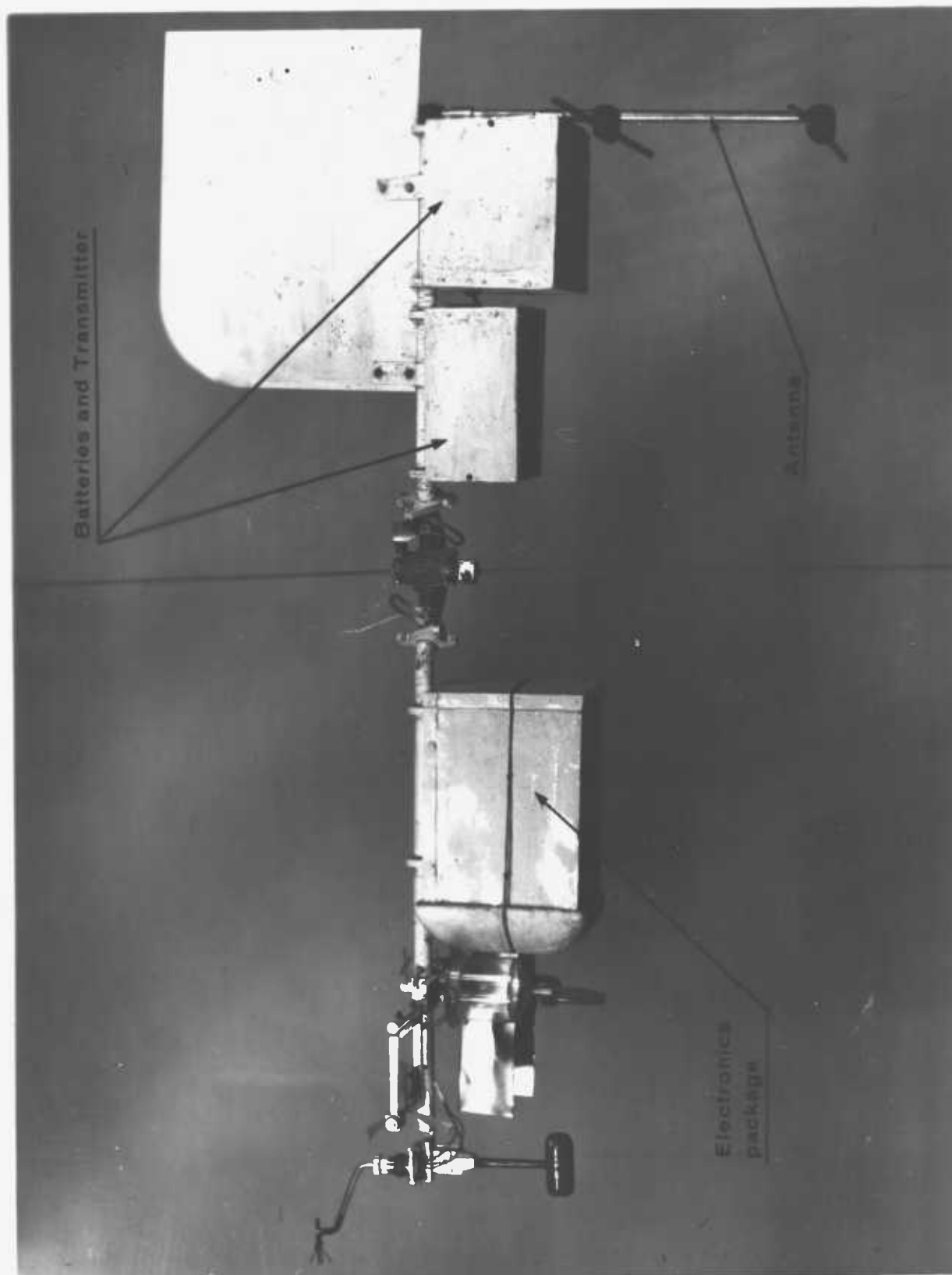


Figure 5 (a) Airborne turbulence - sensing package

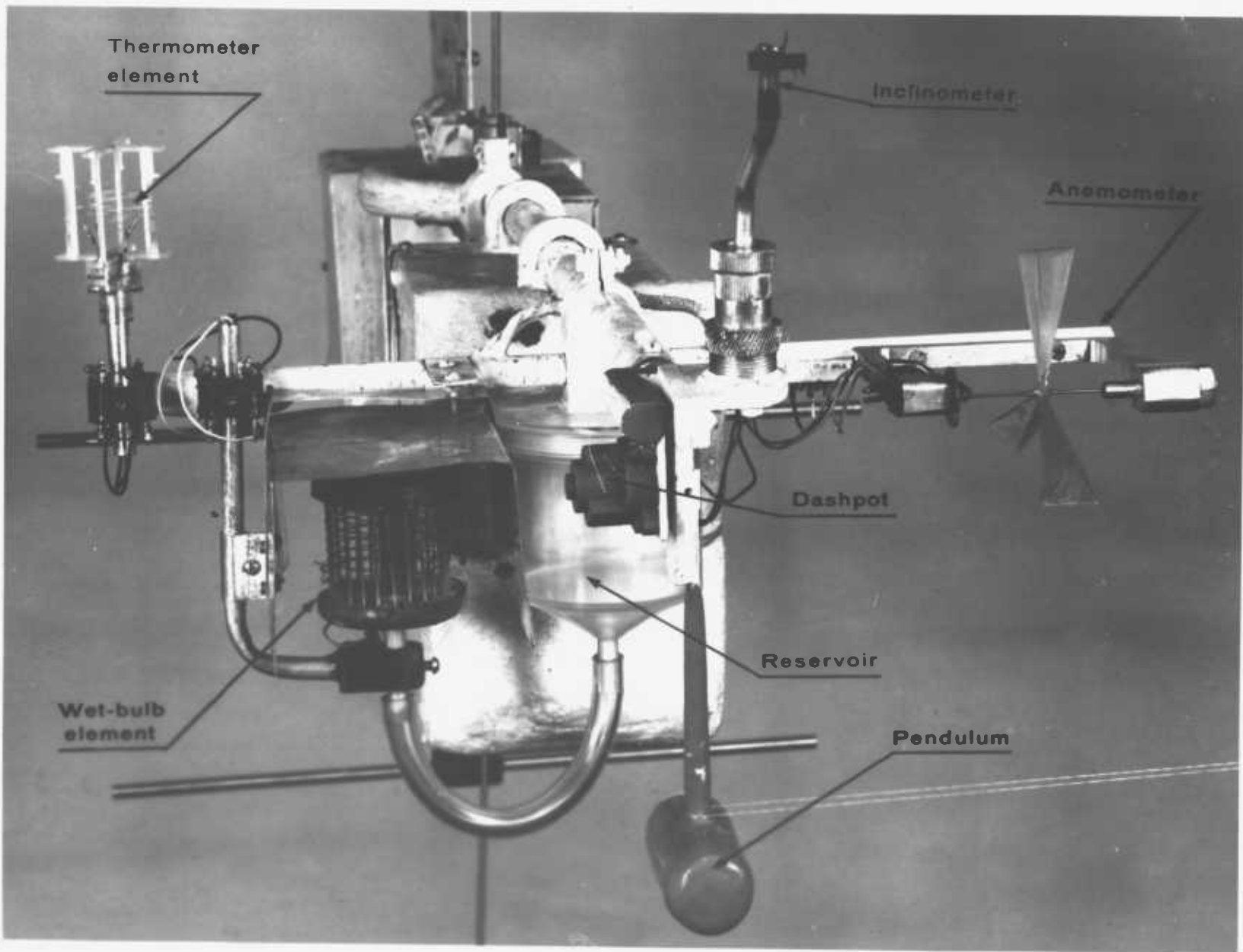


Figure 5 (b) Airborne turbulence - sensing package

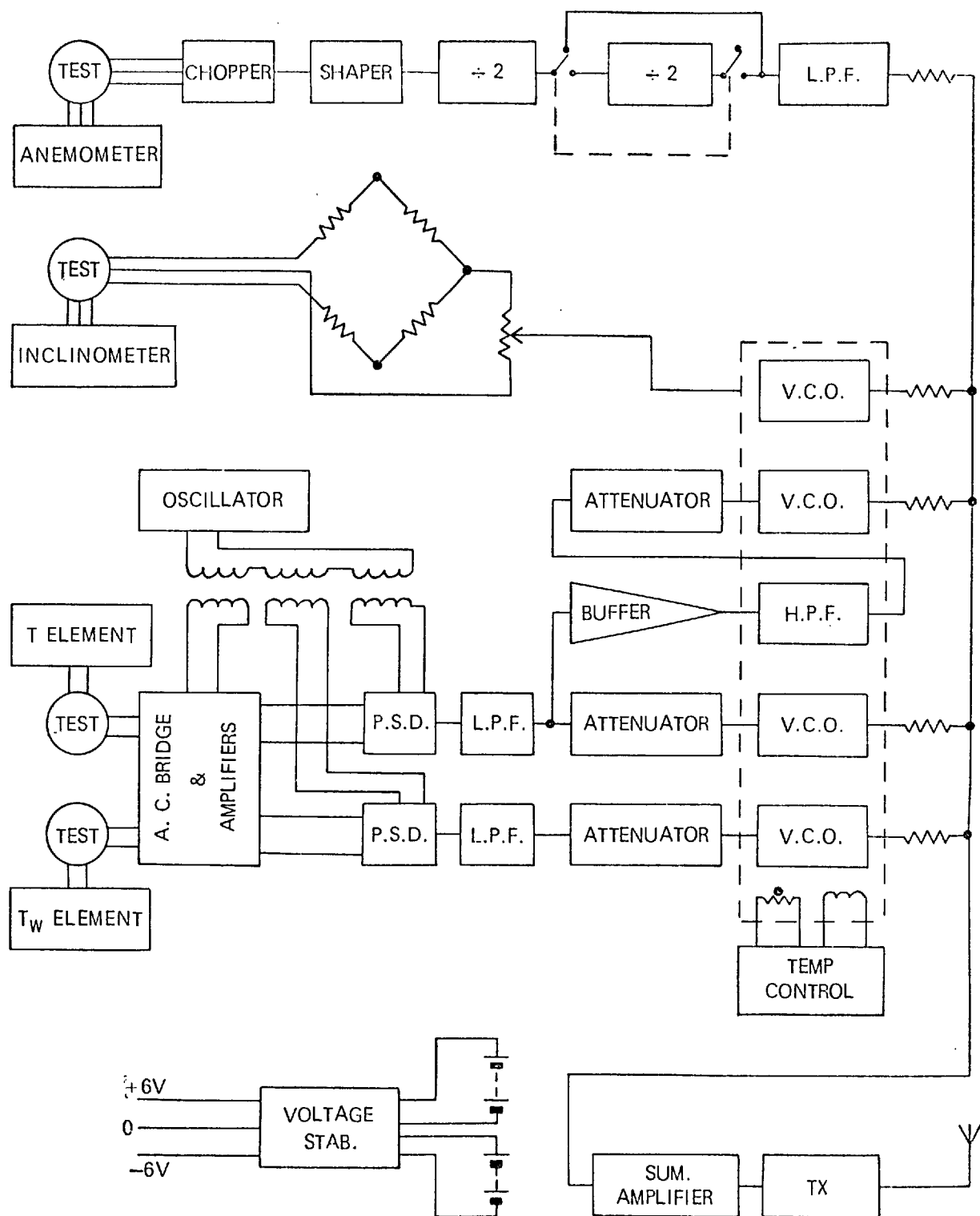


Figure 6 Sensors, signal conditioning and telemetry

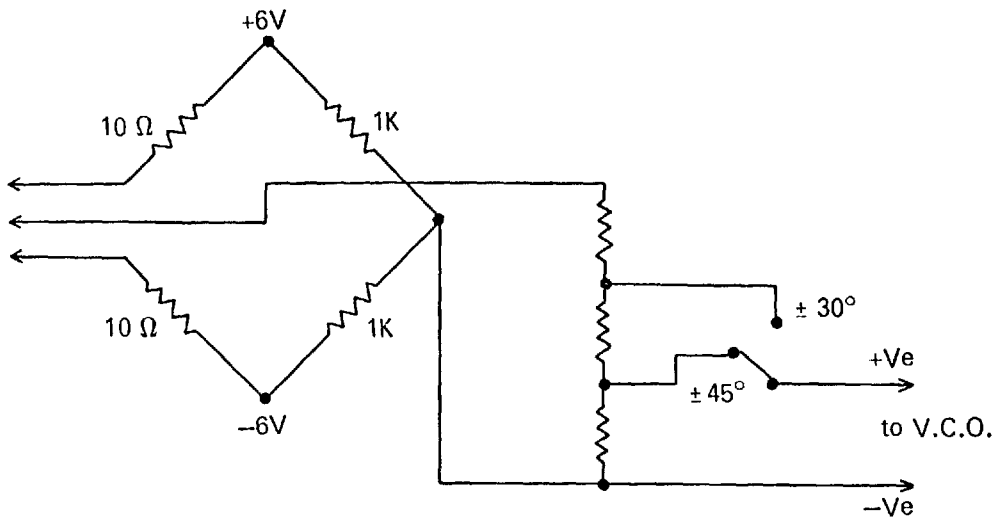


Figure 7(a) Inclinometer bridge

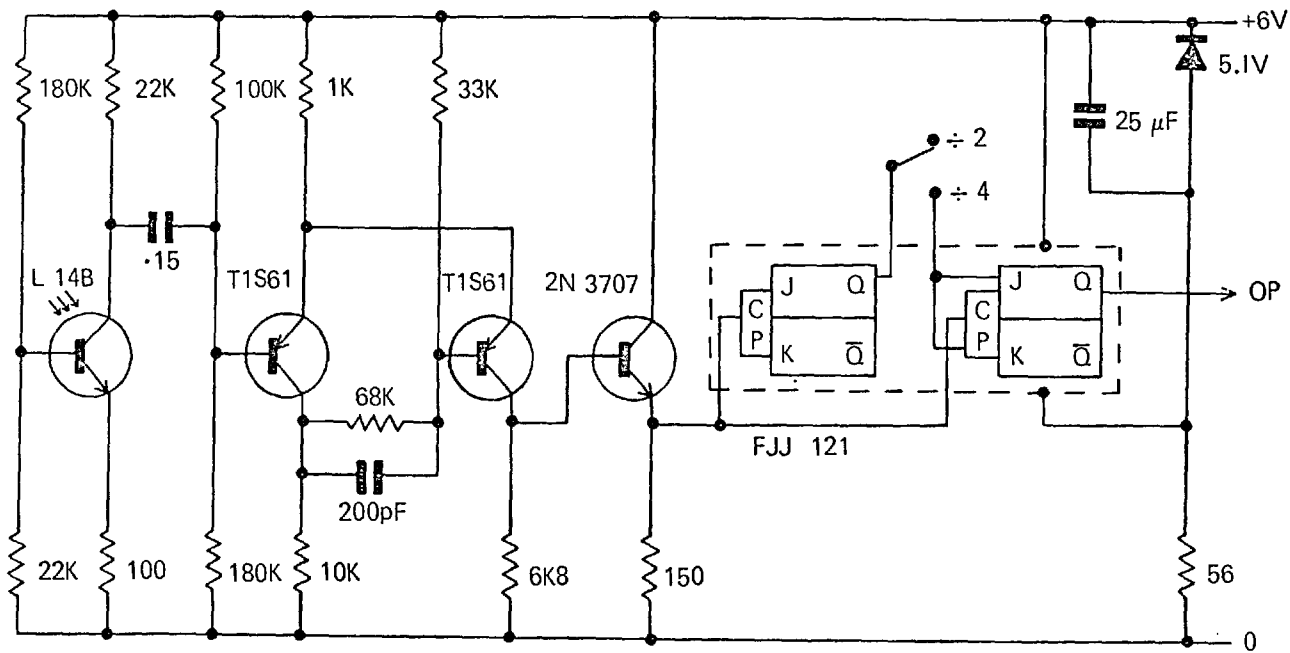


Figure 7(b) Anemometer circuits

the circuit included frequency-dividers to allow operation in light or strong winds. For compatibility with the telemetry and recording system the maximum allowable pulse rate was 650s^{-1} which corresponded to either 9 or 18ms^{-1} .

The thermometer consisted of a spiral of $25\mu\text{m}$ platinum wire wound non-inductively on a transparent plastic former. It was connected in an AC bridge (figure 7 (c)) operated at 2.9 KHz. The RMS current was restricted to about 1mA which gave a temperature rise due to self-heating of about 2×10^{-3} degrees C in a 5ms^{-1} wind (Taylor 1963), producing negligible fluctuations of indicated temperature due to turbulent fluctuations of wind. Heating due to absorption of solar radiation cannot be estimated very accurately but calculations similar to those for self-heating (above) based on incident radiation of 1KWm^{-2} and a reflectivity for the wire of 0.6 (the reflectivity of platinum for normal incidence radiation is about 0.6) gives a temperature rise of about 8×10^{-2} degrees C in a wind of 5ms^{-1} . However the effective reflectivity of the wire will be greater than that for a normal surface both because of curvature of the wire's surface and the incidence of radiation not being normal to the wire's length: the temperature rise will be smaller than calculated therefore, probably less than 5×10^{-2} degrees C. This is still significant but is not very sensitive to changes of wind speed, variations between 3 and 7ms^{-1} producing only about $\pm 1 \times 10^{-2}$ degrees C temperature change. A variation of radiational heating with wind speed will lead to a spurious contribution to the vertical heat flux which may be estimated as follows. The root-mean-square wind speed fluctuations may be estimated in near-neutral conditions from the approximate relations $\sigma_u / u_* \simeq 2.2$ (Mc Bean 1971) and $u_* \simeq 0.035 \langle U_o \rangle$ which follows from eq (3) with $C_D \simeq 1.3 \times 10^{-3}$ leading to $\sigma_u \simeq 0.4 \text{ms}^{-1}$ for $U_o = 5\text{ms}^{-1}$. Thus fluctuations of indicated temperature due to typical natural fluctuations of wind speed are around 2×10^{-3} degrees C. Associated

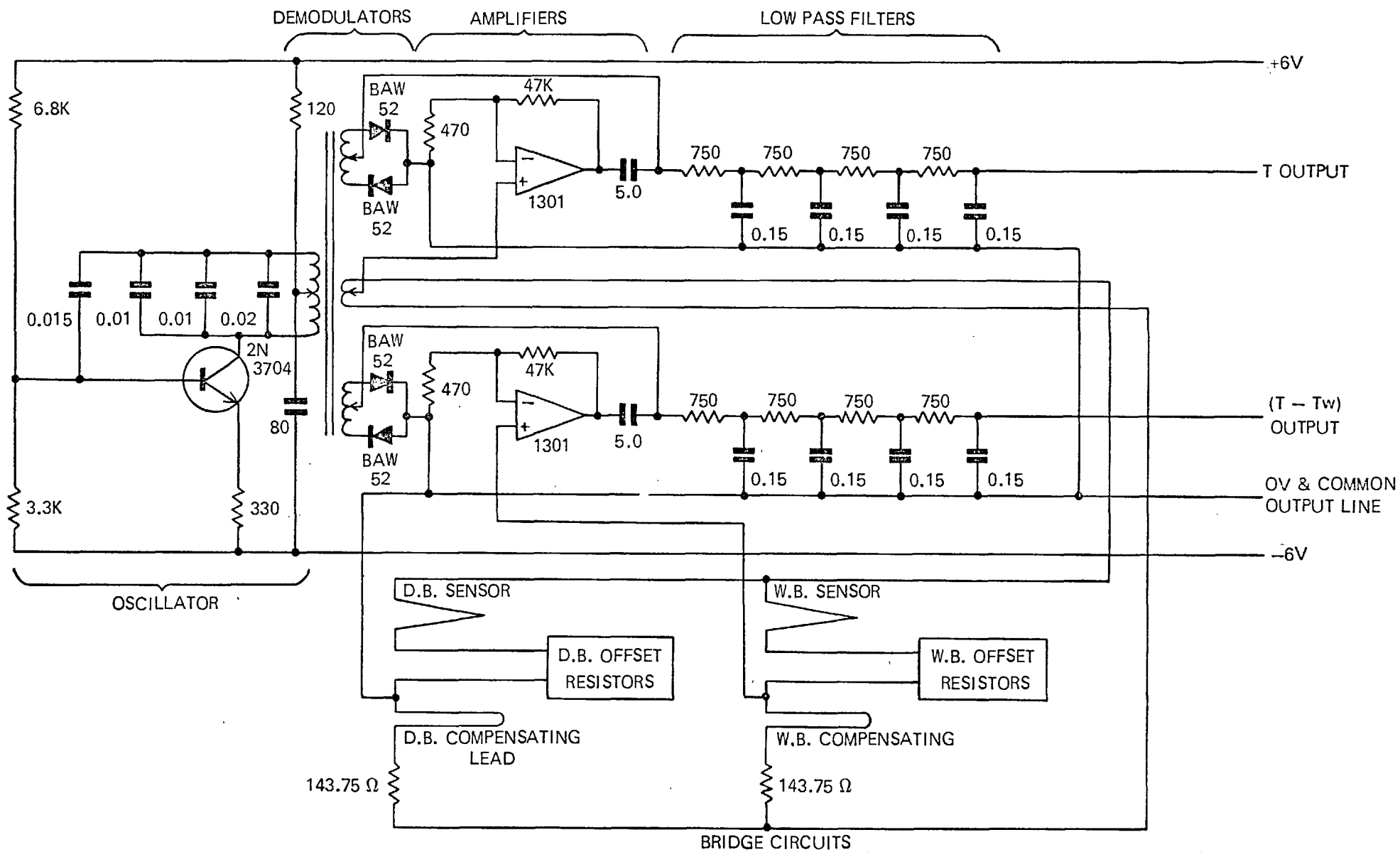


Figure 7(c) Temperature bridges

with the horizontal wind fluctuations are root-mean-square vertical velocity fluctuations around 0.25 ms^{-1} (since $\sigma_w / \sigma_u \approx 0.6$, Mc Bean 1971), and assuming a correlation coefficient of -0.3 between horizontal and vertical wind fluctuations (Haugen et al 1971) this leads to a spurious heat flux arising from lack of shielding from radiation of 0.2 W m^{-2} , which is negligible. The output signal from the AC bridge was amplified, demodulated and smoothed. Maximum sensitivity was ± 0.2 degrees C full scale. Switched binary-scaled resistors allowed the centre of the measured range to be set in steps of 0.5 degrees C between 0 and 31.5 degrees C.

Specific humidity was sensed indirectly, the relevant parameter which was measured being wet-bulb depression. The sensing-element was a $25 \mu\text{m}$ diameter platinum wire round which were wound two helices of very fine cotton (each being one of the three threads which are twisted to form "machine" cotton). The whole was twisted with machine cotton and wound spirally on vertical cotton wicks (in a similar way to that described by Dyer and Maher 1965) which were fed top and bottom by de-ionised water. The measured diameter of the element when wet was about 0.5 mm giving a time-constant around 1 s in a 5 ms^{-1} wind (Taylor 1963). Wind-tunnel tests gave variations of less than ± 0.02 degrees C in indicated wet-bulb temperatures over the speed range 2 to 15 ms^{-1} at an air temperature of 20 degrees C and wet-bulb depression of 4 degrees C. With a depression of 8 degrees C and an air temperature of 23 degrees C some drying out of the element was observed but such low humidities would not be expected usually in the lowest few-hundred metres over the sea except perhaps in the meteorologically-uninteresting case of strong subsidence and stable density stratification near the surface. The element was connected in the AC bridge system (figure 7 (c)).

Power was provided by lightweight rechargeable silver-zinc accumulators connected via a voltage-stabiliser giving outputs of $\pm 6 \text{ V}$ (figure 7 (d)):

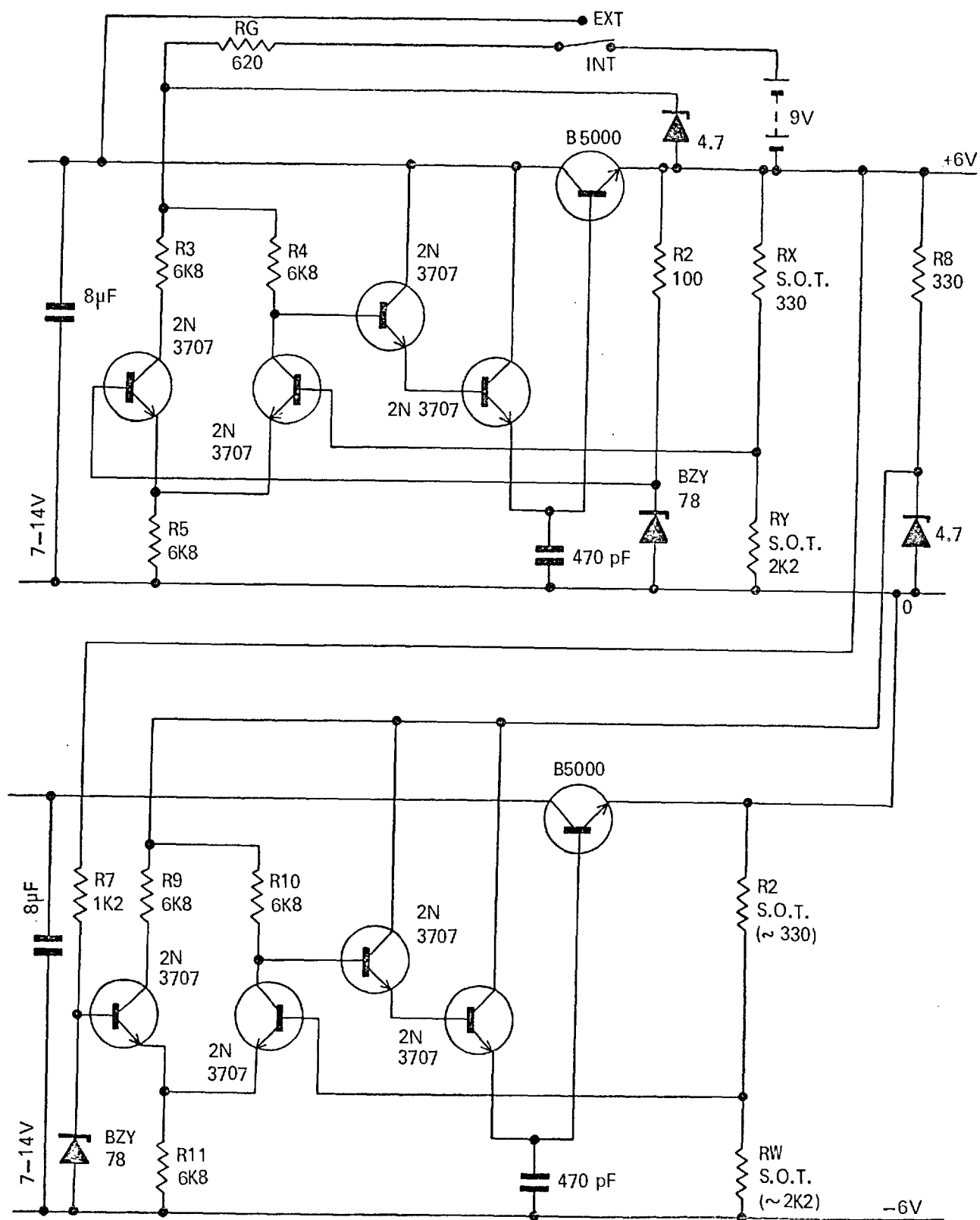


Figure 7(d) Voltage stabilizer

these last varied by less than $\pm 20\text{mV}$ for input voltage, output current and temperature changes between 7.5 and 14 V, 100 and 500 mA, and 0 and 30 degrees C respectively. Battery capacity was sufficient for about 5 hours operation before recharging was necessary.

The telemetry system was a commercially-available form using voltage-controlled oscillators (VCO's) and a frequency-modulated UHF transmitter. The two outputs from the temperature bridge and the inclinometer bridge output were fed to three VCO's with centre-frequencies around 4, 5 and 6 KHz and output-bandwidths of $\pm 250\text{Hz}$. The temperature signal was also high-pass filtered (figure 7 (e)) with cut-off at 0.1Hz and fed to a VCO with centre frequency near 7KHz. The output frequencies were mixed with the anemometer signal and the resulting frequency-multiplex modulated a 400 MHz transmitter feeding a half-wave dipole antenna. The VCO's were sensitive to temperature variations, typical drifts being about 0.4% of bandwidth per degree C, and were therefore kept in a simple temperature-controlled oven at about 10 degrees C above ambient temperature but it was not possible to achieve high stabilization because of heat conduction down the leads. Drifts in experiments were probably less than 1% of bandwidth (tests were carried out with the complete turbulence probe mounted in a wind tunnel and the oven set at 10 degrees C above ambient: VCO output frequency changes for speed changes from zero to 10ms^{-1} were around $\pm 2\text{ Hz}$, equivalent to temperature stabilization within about ± 1 degree C).

Sensors, electronics and batteries were all mounted on a vane about 1m long which pivotted about a bearing attached to the balloon cable and maintained the orientation of the sensors with respect to the wind. The vane also rotated about an axis along its length to allow for the plane of the balloon cable not always being in the plane of the instantaneous wind. An adjustment was included by which the vane tube could be levelled app-

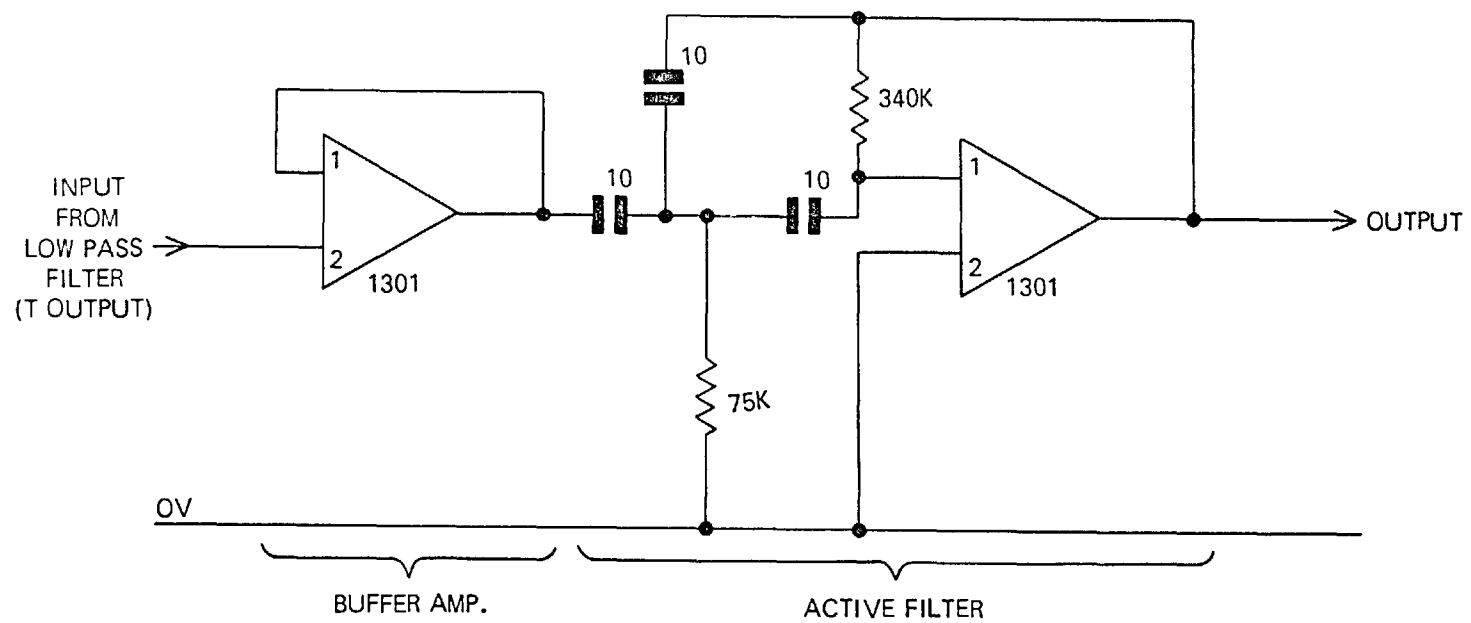


Figure 7(e) T High-pass filter

roximately for balloon-cable inclinations up to 40° from the vertical. The total weight was about 3.5 Kg.

3.4

RECORDING

The receiving aerial, positioned near the ship's stern, consisted of properly phased crossed-dipoles mounted above a ground plane. The signal was led to the recording equipment in the "dry" laboratory. Here the receiver output was mixed with a 12.5 KHz signal from a reference oscillator and directly recorded on one track of a 7-track tape recorder at 9.5 cm s^{-1} . The recorded information was monitored continuously using the replay heads and amplifiers, being led from these to band-pass filters and frequency-to-voltage converters whose outputs were displayed on multi-channel UV recorders (figure 8). The recorded reference frequency was used to compensate electronically for variations in voltage outputs resulting from tape-stretch or "flutter". Experience showed that continuous monitoring of all signals was essential as a check on both instrumental performance and meteorological variability.

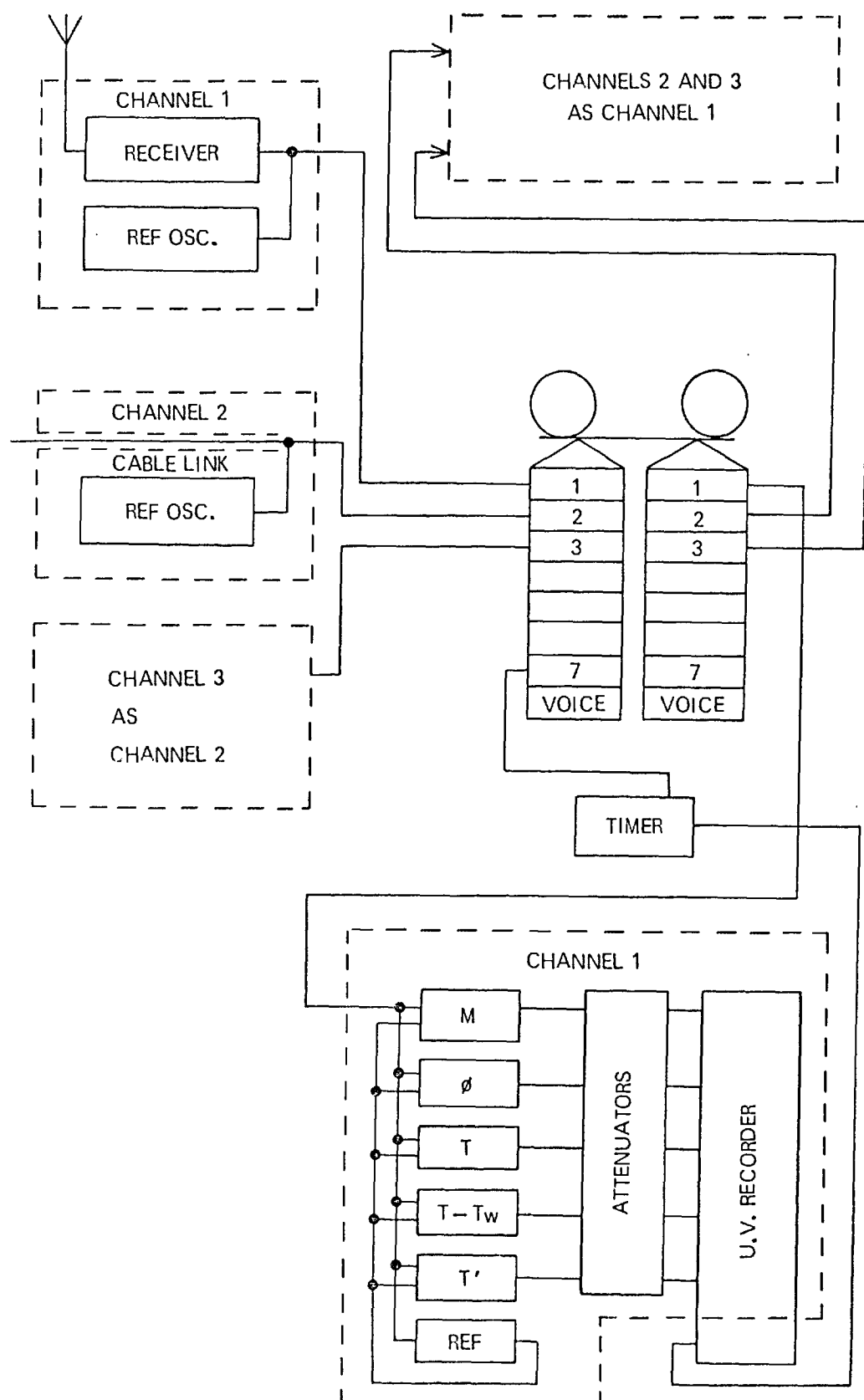


Figure 8 Ground equipment

3.5

CALIBRATION

Inclinometers were calibrated by rotating them in the vertical plane of the wind in a wind tunnel operated at 8ms^{-1} , outputs being measured at 10 degree steps between ± 40 degrees. The small change of output sensitivity with wind speed (less than $\pm 4\%$ for inclinations up to ± 30 degrees over the speed range 3 to 15ms^{-1}) was neglected. Anemometers were calibrated at several speeds between 2 and 18ms^{-1} using a pitot-static tube as the standard. Resistances of the wet and dry thermometers were measured when dry at several temperatures between 5 and 25 degrees C while inside an airfilled chamber immersed in a temperature - controlled water bath. Variable dummy sensors were used to measure the non-linearities of the airborne circuits and their associated VCO's. Final testing and calibration were done on board ship: here dummy sensors were used to adjust sensitivities and VCO centre-frequencies.

CHAPTER 4DATA

4.1

INTRODUCTION

Measurements of atmospheric turbulence were made during the Atlantic Trade Wind Expedition (ATEX) in February 1969 and the Joint Air/Sea Interaction Experiment (JASIN) in June 1970 (Royal Society Report, 1971). The ATEX observations (Thompson 1971b) were made using a prototype of the turbulence measuring and recording system described in §3 above and were of lower quality and are much more limited than those from JASIN because of difficulties with the instrumentation. They are however included here not for historical reasons but because they provided the first indications that the tethered-balloon technique could be used with some success at sea.

4.2

ATEX

This was a collaborative, mainly meteorological, experiment involving four ships spaced several hundred kilometers apart and drifting (lying-to) in the North-Easterly Trade wind belt. Turbulence measurements were made from HMS "HYDRA" at heights between about 50m and 400m using a first-version of the balloon-borne instruments which differed considerably from that described above. Numerous instrumental problems were encountered, stemming mainly from the late delivery of the telemetry system and the haste with which this had to be incorporated in the instrument, and out of about 20 experimental runs (which were typically between 40 and 90 minutes long) it was possible to analyse data from 5 only. Details of these are given in Table 2.

Table 2 Data from ATEX

Run No.	Date	Time	Position	Instrument Height (m)	Parameter Measured	Wind Speed at this height (ms^{-1})	Est Speed at 10m (ms^{-1})	$\Delta\theta$ (Surface -10m) (deg C)	Δq (Surface -10m) $\times 10^3$	L (m)
2	6/2/69	1839-1947	11.0N 36.7W	180	M, ϕ	7.9	6.9	0.2	4.2	-120
3	7/2/69	1300-1424	10.9N 37.1W	180	M, ϕ , T	9.1	7.8	0.1	4.0	-180
4	7/2/69	1657-1740	10.8N 37.2W	90	M, ϕ , T	8.5	7.5	0.1	4.2	-160
5	7/2/69	1907-2006	10.8N 37.2W	310	M, ϕ , T (dummy sensor)	8.2	6.9	0.1	4.1	-140
6	8/2/69	1307-1406	10.7N 37.5W	90	M, ϕ , T, T-T _w	9.5	8.4	0.3	5.0	-140

Over the observational period including these runs there was continuous cumulus development with the boundary layer varying in depth between about 1,000 and 2,000m. $\Delta\theta$ and Δq were obtained from a careful series of observ-

ations made from the ship at 90-minute intervals using an aspirated psychrometer at about 10m height and a bucket thermometer. Winds at 10m were estimated from those measured by the turbulence instrumentation by using the integrated form of eq (2) :

$$U_{10} = U_z - (u_* / \kappa) (\log (z / 10) + \psi_m (10/L) - \psi_m (z/L)) \quad (15)$$

: ψ was obtained from the estimates given by Dyer and Hicks (1970). For this purpose L was given with sufficient accuracy by

$$L = -2.4 U_e^2 / (\Delta \theta + 175 \Delta q) \quad (16)$$

(Thompson 1971 a): U_e was a rough estimate of the wind at 10m, assumed to be 90% of the measured wind (this also determined u_* using eq (3) with $C_D = 1.3 \times 10^{-3}$). The assumption that the constant-flux layer was deep enough to include all the levels at which the data were obtained was undoubtedly not correct and so the calculated 10m winds are only rough estimates. The listed values for L were obtained using these 10m winds in eq.(16).

4.3

JASIN

This was a joint oceanographic and meteorological expedition near OWS "Juliett". Its primary purpose was to test various techniques intended to be used during the Royal Society-sponsored air/sea interaction experiment planned for the same area in 1975. The turbulence data were obtained from HMS "Hecla" using the instrumentation described in § 3. A number of different groups had demands on ship time and this restricted considerably the data which were obtained. For example the ship's main transmitters produced enough power to affect adversely the performance of the turbulence-measuring equipment, thus preventing the acquisition of usable information for significant periods each day. A wide range of meteorological conditions was encountered during the 8-day period when opportunities occurred to make turbulence measurements. In the first three days winds were light and the near-surface density stratification usually stable, but the following days were more unsettled with stronger winds, significant convection and rain or showers at times. Two of these later days were wasted, one because of bad weather and the other (the final day of the expedition) when the balloon while being topped up was damaged by a squall of around 20ms^{-1} which temporarily turned the ship broadside-on. Most of the data of real meteorological interest were therefore obtained on three days only. Up to three turbulence-measuring instruments were flown at heights between about 50 and 200m, usually while the ship was making way along the line of the wind or else lying or hove-to. Radio-telemetry was available for only one of the instruments and if others were flown the additional frequency-multiplexes were fed to the recording system via electrical cables. A total of about 30 hours data was obtained during 11 periods each about three hours long usually. After discarding sections when ship-manoevres were taking place this left 19 more-or-less homogeneous records with lengths between 40 minutes and about 150 minutes.

The main supplementary data included observations from a meteorological buoy, and (obtained by other groups) a careful series of observations at 1-hourly intervals of air, wet-bulb and sea temperatures using a psychrometer at about 10m and bucket thermometer both deployed to minimise the influence of the ship, and occasional radio-sonde ascents.

Table 3 lists relevant general information for the 19 runs. The buoy data were obtained from a prototype self-recording system moored in 2600m of water at the centre of the experimental area: the maximum distance of the ship from the buoy was 17Km. A shielded but un aspirated thermometer system was used and there were probably significant radiation errors especially early in the exercise when winds were very light, but sea temperature (deduced from recorded values of air temperature and air-sea temperature difference) should have been reasonably reliable. In the quiet conditions that prevailed at first there was a significant warming of the surface-layer of the sea but this tended to be masked in the observations taken from the ship because of the stirring action of the ship: in these circumstances the buoy sea-temperature (at a depth of 0.7m) was often significantly higher than that measured from the ship in the afternoons, sometimes high enough to indicate unstable thermal stratification in the lower boundary-layer. The actual sea-surface temperature would probably have been a few tenths of a degree higher than the buoy sea temperature in these circumstances. In the unsettled spell that followed the ship data are preferred because wave-mixing then produced a well-mixed and nearly uniform upper ocean, and also because marked horizontal temperature gradients were found and so buoy measurements were not necessarily representative of those at the ship.

Winds at 10 metres were estimated using eq (15) and (16) in conjunction with the mean speed measured by the lowest balloon-borne instrument (corrected for ship's velocity). On occasions with stable density stratification Ψ_M was assumed to be 5.2 z/L (Webb 1970). The listed values for

Table 3 General data from JASIN

Run No.	Date (1970)	Start (Z)	Length (min)	Instru- ment heights (m)	(a) Wind Speed (ms^{-1})	(b) Buoy Wind Speed (ms^{-1})	Estim- ated 10m Speed (ms^{-1})	Air temp ($^{\circ}\text{C}$)	Wet- bulb temp ($^{\circ}\text{C}$)	Sea temp (bucket) ($^{\circ}\text{C}$)	Sea temp buoy ($^{\circ}\text{C}$)	$\Delta\theta$ (surface- 10m) ($^{\circ}\text{C}$)	$\Delta\rho \times 10^3$ (surface- 10m)	L (m)	Weather etc	Notes
1	14/6	1330	90	45												Not analysed o/o Cu, Sc
15	18/6	1745	60	95	10.6	7.7	8.5	14.5	12.5	14.1	14.5	-0.5	1.6	+750	Dry,	" 0.6ms ⁻¹
16	18/6	2020	49	95	14.3	8.9	11.6	14.3	12.5	13.9	14.5	-0.5	1.5	+1600	Dry, 3/8 Cu, Sc	Headed into wind at 1.6ms ⁻¹
17	20/6	1021	86	95	8.3	6.3	7.3	11.9	8.3	12.3	12.3	0.3	3.4	-150	Dry apart from light shower at end	Drifting acrosswind at 0.3ms ⁻¹ 5/8 Cu, Cb
18	20/6	1830	77	95	8.7	5.9	7.8	11.0	8.1	12.6	13.1	1.5	3.4	-70	Dry, 5/8 Cu, Sc	Hove to
19	20/6	2005	92	95	9.0	7.3	8.1	10.5	8.3	12.3	13.1	1.7	2.8	-70		

Notes (a) From turbulence instrument, corrected for ship's speed
 (b) At 2m
 (c) Using buoy sea temperature

L were obtained using equations (3), (8a) and (9a) with transfer coefficients corrected for departure from neutral stability (the neutral values for all three coefficients were assumed to be equal at 1.3×10^{-3}).

CHAPTER 5DATA PROCESSING

A similar scheme was adopted for the computer-processing of the turbulence data from both ATEX and JASIN. On playback the recorded frequency-multiplex was fed to band-pass filters which isolated the individual VCO outputs which were then led to discriminators (frequency-to-voltage converters). Fluctuations in the recorded reference frequency were monitored by a reference discriminator whose output was used to adjust the other discriminator outputs electronically and hence compensate for tape-speed variations. The voltage outputs were passed through low-pass filters to reduce aliasing and then sampled by an analogue-to-digital converter with punched paper tape output. Sampling and filter details are listed in Table 4. All data were

Table 4 Analogue-to-digital conversion

Parameter	Filter Type		Sampling Rate	
	ATEX	JASIN	ATEX	JASIN
U	4-stage RC, -3db at Nyquist freq.	4-stage RC, -12db at Nyquist frequency	1 scan per second	1 scan per second for all data: also 5 scans per second for parts of runs 11, 12 and 13.
T, ϕ	4-stage Butterworth -3db at Nyquist frequency			
T-T _w	4-stage RC, -3db at Nyquist freq.			

sampled once per second but additionally ten equally-spaced sections each about 200s long were selected from JASIN runs 11, 12 and 13 and sampled at five times per second. The filtering used for the JASIN data appeared to reduce aliasing to negligible amounts (this will be demonstrated in § 7) and the filters themselves were properly matched, but a less-satisfactory system had to be used for ATEX, with some mismatch (ie. unequal phase-shifts) and a smaller reduction of aliasing in the case of U and T-T_w. However the ATEX data were of lower quality and deficiencies in the filtering technique were of less importance compared to the shortcomings in the data. In any case the

differences in phase-shifts at frequencies of real interest, that is those lower than the band including the bulk of the ship-induced sensor motions - say less than 0.1Hz - were negligible.

Figure 9 gives a brief summary of the way the data were processed by the digital-computer (an English Electric KDF9). The data were transferred from paper tape to "archive" digital magnetic tape before any substantial calculations were done; at the same time they were converted from the binary-coded decimal numbers punched on the paper tape to floating-point numbers which were scaled to meteorological parameters. In this way the data were available for subsequent calculations in a form which could be quickly read into the computer's high-speed store, and the unsatisfactory process of repeatedly handling paper tape was avoided. The error check (figure 9(a)) detected obvious mispunchings eg. characters other than digits or legitimate separators, and substituted data from the previous scan where necessary.

Linear functions were used to scale the A-D converter outputs in the case of the ATEX data but for JASIN cubic functions fitted to the calibration data by least-squares were used to simulate the non-linearities of the sensors and associated electronics.

The temperature and wet-bulb depression data were used to obtain specific humidity by means of the following expressions.

$$q = 0.622e/(p - 0.378e) \quad (17)$$

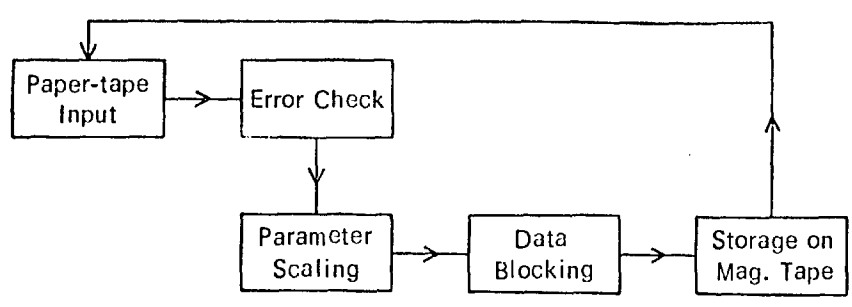
$$= e_{sw} - Ap (T - T_w) \quad (18)$$

where $A = 6.60 \times 10^{-4} (1 + 1.15 \times 10^{-3} T_w)$ for p in millibars (Smithsonian Meteorological Tables). e_{sw} (the saturation vapour-pressure at the wet-bulb temperature) was calculated using the approximation given by Langlois (1967).

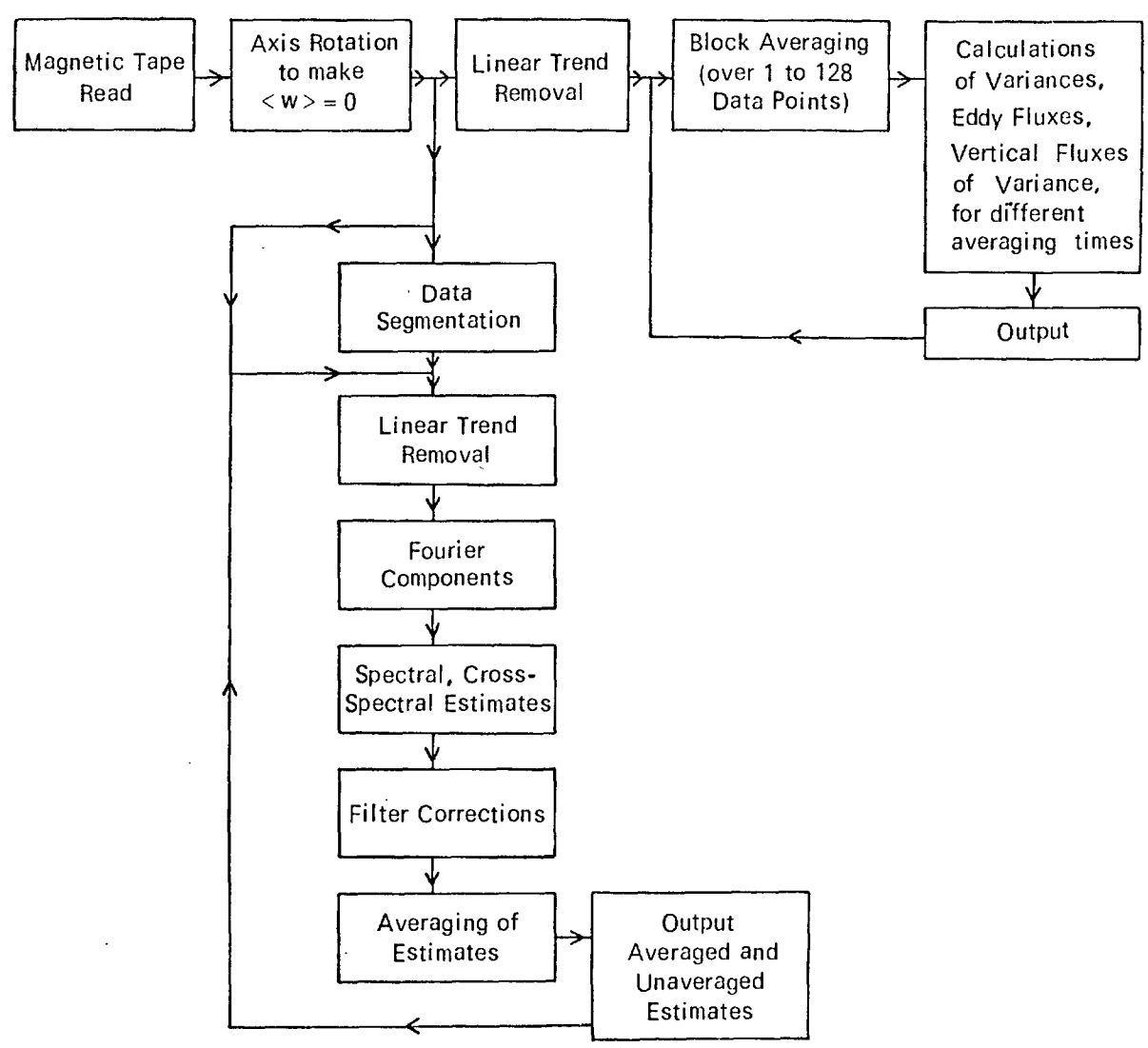
$$= (T_w^2 - 488.56 T_w + 60009.3) / (0.0361622 T_w^2 - 24.209 T_w + 4104.45) \quad (19)$$

: this was accurate to better than 0.25%.

Data from each channel were assembled in separate areas in the core-store



(a) Data storage



(b) Calculations on stored data

Figure 9 Data processing

and written to an "archive" magnetic tape as blocks of 512 numbers.

Conventional procedures were employed for adding data from further runs to the archive: thus the existing information on the tape was copied to a new tape and the new data then added, so preserving the original tape for use again if problems were found in writing the new data.

The main calculations were the computation of variances, eddy fluxes, spectra and cospectra. Because of the contributions due to ship's motion the usual practice of normalising the spectral estimates by variances or quantities such as u_* or T_* was not relevant and partly for this reason the spectral and cospectral calculations were done separately. Figure 9 (b) gives an outline of the computing procedures. It was assumed for all the calculations that the mean vertical velocity for each run was zero and while this must be true for measurements close to the surface over a sufficiently long period, away from the surface there may be mean vertical motions resulting from either too short a sampling period so that long-period eddies are not adequately sampled, or because of synoptic-scale ascent or descent. At heights of a few hundred metres the velocities of these latter organised motions are typically around 0.01 ms^{-1} and in a 5 ms^{-1} wind this produces a "tilt" error of about 0.1 degrees, or around 1% in the calculated momentum flux and 0.2% in the velocity variances (Rayment and Readings 1971). Random tilts produced by too short a sampling period probably would have negligible effect on measured fluxes unless sampling periods were so short that the calculated fluxes would be in any case considerably underestimated due to loss of low-frequency contributions.

When the computer programmes for data-processing were written it was anticipated that most of the contributions to variances and fluxes due to sensor motion would be confined to a reasonably well-defined band of frequencies. It was thought therefore that attempts to eliminate the affects of sensor motion could be carried out along two lines, one involving only

calculated variances and fluxes, the other spectra and co-spectra also. In the first case the variances etc were calculated from the turbulence data to which moving averages of varying lengths were applied, thus simulating low-pass filtering with variable cut-off frequency. Linear trends and means were first removed by fitting linear functions to the data by a least-squares method. Running averages over 2^m data points ($m = 0$ to 7) were then applied and the following quantities evaluated for each average:

$$\begin{array}{cccc} \langle U'^2 \rangle & \langle w^2 \rangle & \langle T'^2 \rangle & \langle q'^2 \rangle \\ \langle wU'^2 \rangle & \langle w^3 \rangle & \langle wT'^2 \rangle & \langle wq'^2 \rangle \\ e \langle Uw \rangle & e c_p \langle Tw \rangle & e \langle qw \rangle & \end{array}$$

that is, the variances, vertical flux of variance and vertical turbulent fluxes. If the number of data points over which the averaging was carried out was large then the attenuation F produced by the numerical filter in the case of a turbulent fluctuation of frequency n was

$$F = (\sin \pi ns) / \pi ns \quad (20)$$

where s was the averaging time (Pasquill 1962): the corresponding cut-off frequency was $0.45/s$. For short averaging times (ie for very small values of n) the filter attenuation may diverge significantly from that given by eq (20) because of the increasingly important influence of aliased high frequencies. However the effects would not have been very important in the present cases because of the electrical pre-filtering of the data. Most of the effects of ship motion on the measured turbulence occurred at frequencies above about 0.1Hz , and an averaging time of 8 seconds (corresponding to a cut-off at about 0.06 Hz) was suitable for drastically attenuating these without significantly decreasing the energetic genuine turbulent fluctuations in many cases. It will be shown later though that the technique had limitations when there was very large contamination of the turbulence signals by sensor motion, or when measurements were made relatively close to the surface in stronger winds.

Spectra and cospectra were evaluated using the Fast Fourier Transform

(FFT) (Cooley and Tukey 1965), a computationally-efficient method for harmonic analysis. If it is assumed that the length of record T is made up of N equally spaced samples X_j ($j=0, 1, \dots, N-1$) so that

$$T = N \Delta T \quad (21)$$

then following the treatment of Hinich and Clay (1968) the length of record can be represented by the discrete Fourier series

$$X_j = \sum_{k=0}^{N-1} (a_k \cos(2\pi k j / N) + b_k \sin(2\pi k j / N)) \quad (22)$$

or in complex notation

$$X_j = \sum_{k=0}^{N-1} A_k \exp(-2\pi i k j / N) \quad (23)$$

where

$$A_k = (1/N) \sum_{j=0}^{N-1} X_j \exp(2\pi i k j / N) \quad (24)$$

It follows from equation (24) that $A_K = A_{N-K}^*$ (the asterisk denotes the complex conjugate): hence $|A_K|^2 = |A_{N-K}|^2$. $|A_0|^2, \dots, |A_{N-K}|^2$ are the periodogram ordinates at the discrete frequencies $n_k = k/T$ ($k=0, 1, \dots, N/2$).

If n is finite the value of $|A_k|^2$ would be proportional to the power in the frequency band of width $\sim 1/T$ near $n_k = k/T$ if the spectrum is flat in this region. However because of the finite length of record some "leakage" from adjacent estimates will occur unless the spectrum consists only of lines at frequency $0, 1/T, 2/T$ etc. Atmospheric turbulence is composed of a continuum of frequencies and if its spectrum changes very rapidly with frequency in any region then the $|A_k|^2$ will not be good estimators. The actual spectral window has the shape $(\sin \pi T n / \pi T n)^2$ so that a particular frequency in the original record appears not as a single spectral line but as a series spaced $1/T$ apart, with amplitude falling off by a factor of 4 for each octave. Leakage can be reduced by "hanning" which is equivalent to tapering the length of record to zero at each end but this has the undesirable results of reducing the total variance: in any case quasi-

stationary atmospheric turbulence spectra usually have no very sharp lines and so leakage is of little importance if T is reasonably large.

Owing to sampling fluctuations the standard deviation of $|A_n|^2$ is very large (equal to the size of the mean) and to obtain meaningful power-spectral estimates it is necessary to trade off some of the resolution of the periodogram, either by averaging across frequency or else by breaking up the record into non-overlapping pieces and then computing average values for corresponding periodogram ordinates. For example, if 2 estimates were used for obtaining each average value there would be an 80% chance of the ratio (averaged estimate)/(averaged estimate from an infinite number of samples of length N) lying between 0.33 and 9.5, but by using 20 estimates the range is reduced to 0.7 to 1.7 (Mercer 1968).

Aliasing is the result of a sampling rate too low to resolve the highest component frequencies of the record. By Shannon's theorem the maximum frequency that can be sampled without aliasing is given by the Nyquist frequency,

$$n_N = 1/2\Delta T \quad (25)$$

Variance contributions at higher frequencies are folded back across n_N to contaminate the calculated estimates.

The conventional discrete Fourier transform (DFT) applied to a series of N samples where $N = 2^p$ requires about $2N^2$ arithmetic operations to evaluate the Fourier coefficients, in contrast to the FFT which uses about $2pN$ operations, a very large saving for typical values of p around 10. The FFT takes advantage of the fact that the DFT of a sequence of N numbers can be reduced to computing the DFT of two sample sequences of length $N/2$ which are usually odd and even-numbered sequences from the original data, followed by a simple recombination of these DFT's: the sample sequence of length $N/2$ can in turn be subdivided in the same way so that the DFT's are

carried out ultimately on a number of short sequences (Cochran et al 1967).

An FFT subroutine written by Rayment (1970) for KDF 9 was used. Real values for X were used so that $a_{N-k} = a_k$ and $b_{N-k} = -b_k$: by selecting even values for N , $b_0 = b_{N/2} = 0$ so that only N unique non-zero Fourier coefficients resulted from each FFT, leading to simplification of their subsequent storage in the computer system because they could be overwritten on the data points from which they were produced. The power-spectrum estimates were then calculated as

$$S_{xx}(\eta_k) = 2T(a_k^2 + b_k^2) \quad (26)$$

at frequencies k/T ($k = 0, 1, \dots, N/2$). If the FFT was applied to two simultaneous time-series X and Y , yielding Fourier coefficients a_k, b_k and c_k, d_k , then the cospectral estimates were

$$S_{xy}(\eta_k) = 2T(a_k c_k + b_k d_k) \quad (27)$$

(Rayment 1970).

The spectral and cospectral calculations were done in two ways (figure 9 (b)). In the first, after the axis rotation to make $\langle w \rangle = 0$ the lengths of data N were subdivided into m segments each of length N/m where $m = N/1024$ (rounded up): linear trend corrections were applied to each section and zeros added to bring them up to 1024. The Fourier coefficients were then calculated for each segment and spectral and cospectral estimates then computed, a correction being applied for the reduction of variance due to the adding of zeros. Means of the m estimates for each frequency were then calculated and output after applying corrections for the low-pass filtering used when the data were initially digitized. The means were also averaged across frequency in non-overlapping blocks of 8 and 16 numbers and then were output both with and without corrections for filtering.

The second method of calculation (used only with the JASIN data) block-averaged the full data-lengths to reduce them to 1024 data points after which similar calculations to those described above were done.

CHAPTER 6

ATEX: RESULTS AND DISCUSSION

The data from ATEX were severely contaminated by pick-up of radio signals from the turbulence package's transmitter which produced frequency offsets in the outputs of the VCO's. Unfortunately these offsets were not constant but varied periodically with changes in relative orientation of sensor platform and electronics box (these were mounted separately on the balloon cable about a metre apart rather than joined together as in JASIN). A preliminary analysis on a short-length of data was carried out at the Institute of Sound and Vibration Research at the University of Southampton using a high-speed A-D convertor interfaced directly with a computer and major peaks were found in spectra and cospectra at around 1Hz, close to the natural frequency of the vane and undoubtedly due to variations in RF pickup. However it was thought that the spurious contributions would be confined mainly to frequencies around and above those of the characteristic ship motion and therefore probably would not intrude seriously on the main energy-containing region of the turbulence spectra and cospectra.

A typical record of wind-vector magnitude (M), inclination (ϕ) and temperature is shown in figure 10. The main spurious contribution due to sensor motion caused by wave-induced ship's motion are clearly seen as fluctuations in both M and ϕ with periods around 5 second. By contrast the temperature record appears to be free of such periodicities. All three signals show significant noise around 1Hz due to variations of RF pickup as explained above. Obviously the velocity signal fluctuations due to sensor motion are very large especially in the case of vertical velocity ($M \sin \phi$) where the genuine turbulent fluctuations appear small by comparison: although there is some phase difference between the spurious contributions to M and ϕ it is very unlikely that U_w correlations due to sensor motion will cancel out and it must be anticipated therefore that velocity spectra and U_w cospectra will all have very substantial contributions above 0.1Hz. This

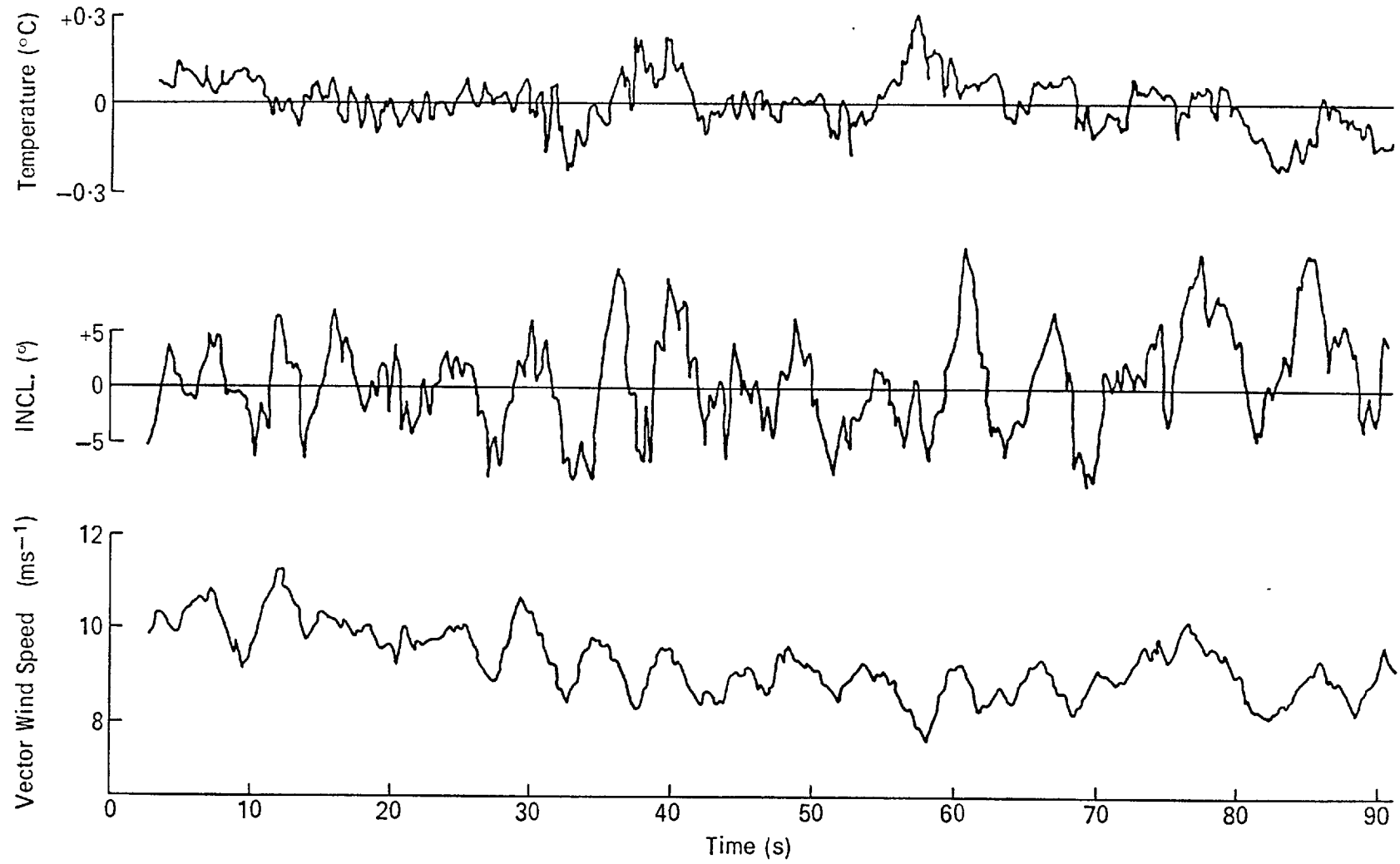


Figure 10 Section of record from ATEX (Run 3, 180m)

Table 5 Calculated eddy-fluxes
deviations for AT&X

Run number	Averaging time()	Equivalent cut-off frequency (Hz)	Momentum flux (Nm ⁻²)	Heat flux (Wm ⁻²)	moisture (mgm ⁻¹)
2 (180m)	0	-	- 0.0369		
	2	0.22	- 0.0243		
	4	0.11	- 0.0151		
	8	0.055	- 0.0191		
	16	0.028	- 0.0174		
	32	0.014	- 0.0164		
	64	0.007	- 0.0140		
	128	0.0035	- 0.0099		
3 (180m)	0	-	- 0.0043	- 7.21	
	2	0.22	- 0.0023	- 7.44	
	4	0.11	- 0.0267	- 8.05	
	8	0.056	- 0.0471	- 8.48	
	16	0.027	- 0.0513	- 8.45	
	32	0.014	- 0.0539	- 7.87	
	64	0.007	- 0.0504	- 7.11	
	128	0.0035	- 0.0407	- 6.16	
4 (90m)	0	-	- 0.0111	- 2.01	
	2	0.22	+ 0.0032	- 1.59	
	4	0.11	- 0.0188	- 2.01	
	8	0.055	- 0.0361	- 2.73	
	16	0.027	- 0.0366	- 2.73	
	32	0.014	- 0.0340	- 2.58	
	64	0.007	- 0.0267	- 2.58	
	128	0.0035	- 0.0166	- 2.24	
5 (310m)	0	-	- 0.0002	1.13	
	2	0.22	- 0.0087	1.47	
	4	0.11	- 0.0330	1.86	
	8	0.056	- 0.0505	2.06	
	16	0.027	- 0.0525	1.93	
	32	0.014	- 0.0573	1.96	
	64	0.007	- 0.0605	1.81	
	128	0.0035	- 0.0571	1.37	
6 (90m)	0	-	- 0.1275	1.48	66.
	2	0.22	- 0.0931	1.33	64.
	4	0.11	- 0.0636	1.01	60.
	8	0.056	- 0.0726	0.79	54.
	16	0.027	- 0.0680	0.67	51.
	32	0.014	- 0.0633	0.56	45.
	64	0.007	- 0.0538	0.43	38.
	128	0.0035	- 0.0411	0.31	29.

is confirmed by Table 5 which lists the fluxes and standard deviations calculated from the unaveraged data, and also for averaging times from 2 to 128 seconds. In general the measured velocity variances and momentum fluxes change rapidly as the averaging time is increased to 8 seconds but remain more nearly constant for further increases. However the momentum fluxes show complex variations as the averaging time is altered and it is clear that any attempt to eliminate the effects of sensor motion without the help of the plotted Uw cospectra will not be very convincing: these cospectra are shown in figures 11(a), (b). The major contributions from sensor motion are clearly seen at frequencies above about 0.1Hz (in all cases at values of f greater than 1) and can be readily smoothed out without significantly attenuating expected contributions from turbulent fluctuations (cf figure 4). The form of the "noise" differs somewhat from run to run, presumably because of differences of sensor height and sea-wave structure but broadly it consists of a positive-going peak between 0.1 and 0.2Hz, with a negative peak beyond. The reason for the bi-modal structure is not clear, though it may be the result of a combination of sensor displacements and variations in rf pickup. What is more important is the positive-going region shown by all the cospectra at frequencies between 0.01 and 0.05 Hz; in most cases this is at reduced frequencies well below 1 which, if the anomaly is a spurious effect, prevents an objective smoothing-out without the risk of reduction of genuine turbulent contributions. This phenomenon was completely unexpected since noise in this region would have been anticipated only as a result of tethered-balloon motions independent of ship's motion, and as pointed out earlier the balloon appeared to fly very stably. On the other hand a plot of balloon height (about 30m greater than sensor height) against frequency of the effect (figure 12) suggests a significant correlation between the two as would be expected if it were due to balloon motion.

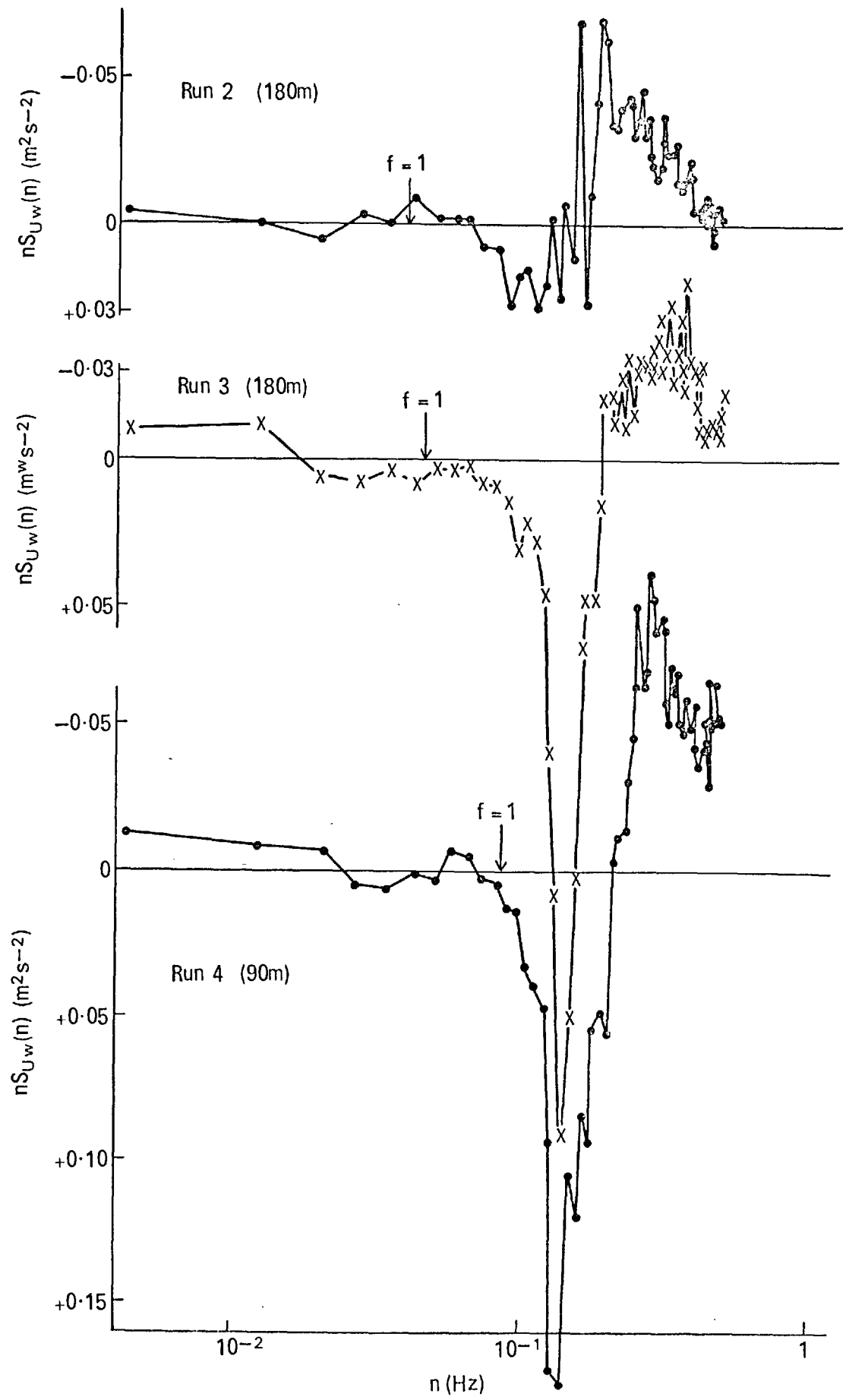


Figure 11(a) Uw cospectra, ATEX Runs 2, 3 and 4

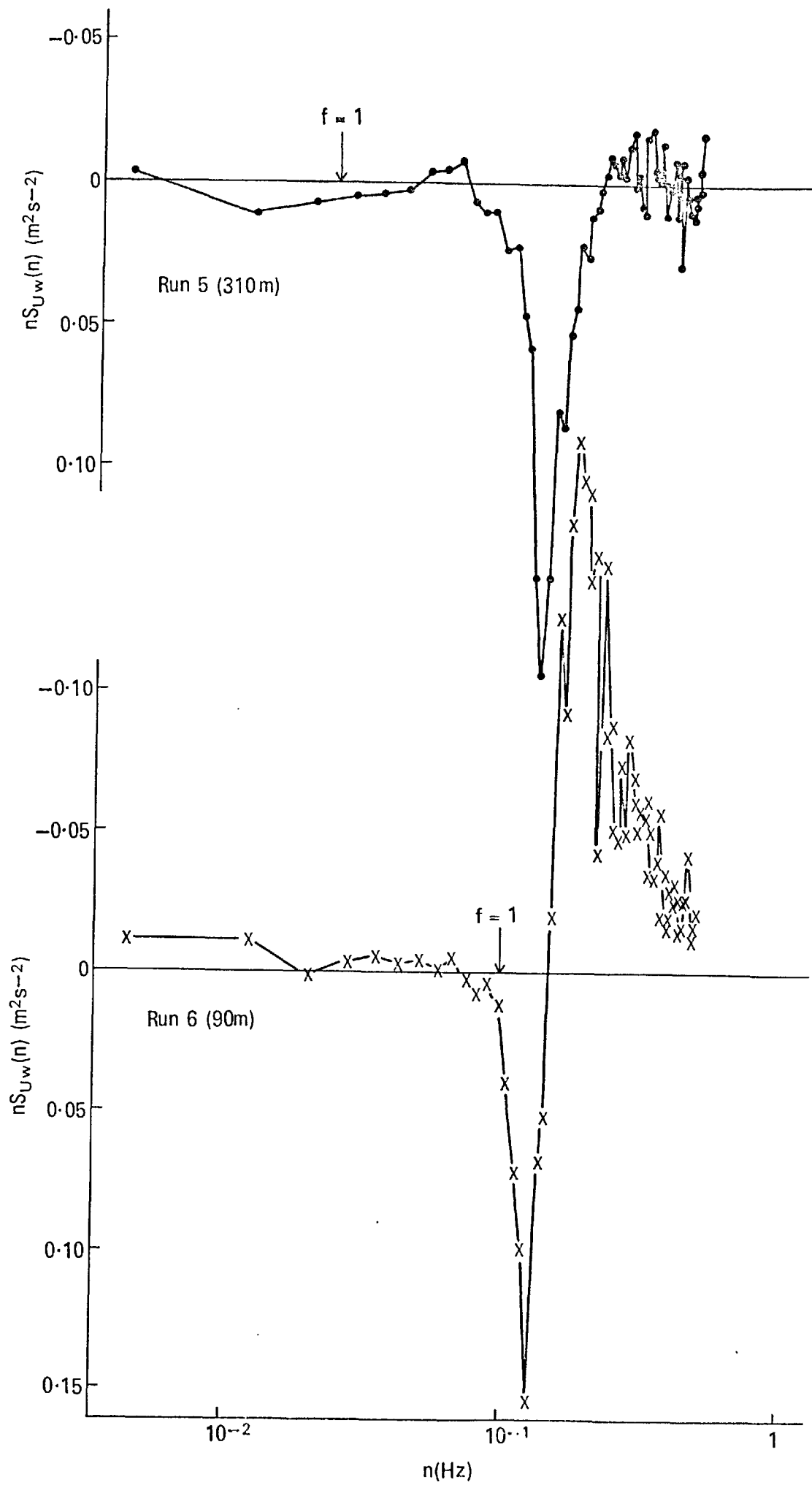


Figure 11(b) U-w cospectra, ATEX Runs 5 and 6

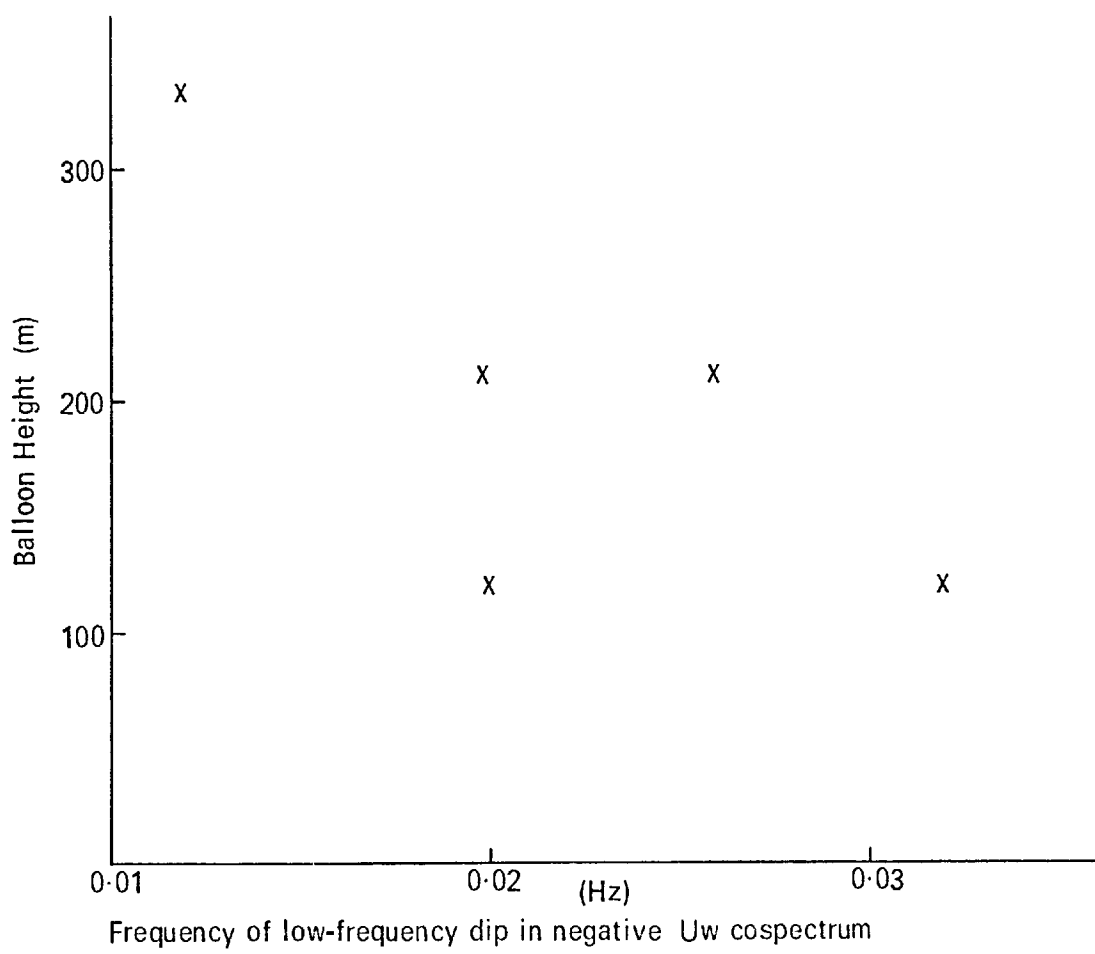


Figure 12 Relation between balloon height and spurious low-frequency contribution to momentum flux

Table 6 gives values for the vertical fluxes of momentum (a) as calculated from the raw (unaveraged) data (b) using 8 second averaged data and (c) obtained by subjectively smoothing the cospectra to zero at and above $f=1$ and then correcting the raw flux estimates by subtracting the contribution lost by the smoothing (this was found by planimetry of the cospectra plotted without corrections for the attenuation by the low-pass filtering used during the initial digitization of data since the raw flux calculations also used this filtered data: filter attenuation at frequencies below 0.1Hz was negligible and so it was not necessary to apply a correction for smoothing of genuine turbulence fluctuations by the filters).

Table 6 Vertical momentum fluxes from ATEX

Run No	Height (m)	Momentum fluxes (Nm^{-2})		
		(a) from unaveraged data	(b) 8s averaged data	(c) cospectra smoothed to zero at $f=1$
2	180	-0.037	-0.019	-0.020
3	180	-0.004	-0.047	-0.045
4	90	-0.011	-0.036	-0.038
5	310	-0.001	-0.051	-0.053
6	90	-0.128	-0.073	-0.074

Estimates from methods (b) and (c) are seen to agree fairly closely: the latter will be used in the following discussions. There is no obvious correlation between size of flux and height above surface but this may be because wind speeds differ between the runs. A better comparison can be made by normalising the fluxes to the mean speed of all the runs by multiplying each by the ratio (mean wind speed at 10m / wind speed at 10m)² (Table 7). However even then the scatter remains large enough to obscure any systematic variation with height. The average from all the runs was -0.045 Nm^{-2} at a mean height of 170m. The corresponding surface flux was estimated as -0.087 Nm^{-2} , using eq (3) with C_D corrected for stability. Thus the downward flux appears to be halved in going from surface to 170m.

Table 7 Normalised momentum fluxes, ATEX

Run No	Height (m)	Normalised flux (Nm^{-2}): $\langle U_{10} \rangle = 7.4 \text{ms}^{-1}$
2	180	- 0.023
3	180	- 0.041
4	90	- 0.037
5	310	- 0.061
6	90	- 0.065

The plausibility of the apparent flux variation with height may be investigated with the help of eq (14). Close to the surface this may be rewritten as

$$\partial \tau_{xz} / \partial z = \rho f V_g \sin \alpha \quad (28)$$

Values for α found by Mendenhall (1967) over the sea appeared to increase with decreasing latitude but he had no data below 17 degrees N where α was apparently around 10° in near-neutral conditions. At 52N (OWS "J") α was around 5° , so there seems to be a very approximate inverse relation between α and latitude: on this basis at 11 N (the latitude of the present experiments) the variation of τ_{xz} with height would be $9 \times 10^{-5} \text{ Nm}^{-3}$, a change of 0.015 Nm^{-2} up to 170m if $\partial \tau / \partial z$ was independent of height (it has been assumed here that $V_g = 10 \text{ms}^{-1}$).

The preceding calculation has ignored the veer of wind with increasing height and baroclinicity, and also it uses a value for α which is little more than speculative, and the result can only be considered a very rough estimate. What is believed to be a useful improvement can be made along the following lines. First an alternative (hopefully better) value for α may be obtained from the measured mean surface wind (055°) and the surface geostrophic wind direction from climatological charts (087°) and this would imply $\partial \tau_{xz} / \partial z = 18 \times 10^{-5} \text{ Nm}^{-3}$ at the surface. In the shallow North-Easterly Trades baroclinicity in the boundary layer may be assumed to result

from horizontal temperature gradients produced by space-variations of sea-surface temperature. The variation of T with height in the lowest few hundred metres probably does not change significantly in the horizontal which allows eq (14) to be re-written:

$$\partial \tau_{xz} / \partial z = \rho f (v_g - v + (zg/fT) \partial T / \partial x) \quad (29)$$

$$\partial \tau_{yz} / \partial z = \rho f (u - u_g + (zg/fT) \partial T / \partial y) \quad (30)$$

The horizontal temperature gradient (from climatological charts) was $0.29^\circ\text{C} / 100\text{Km}$ in the direction 187° , with the resulting thermal wind from 277° , increasing by $3.3 \times 10^{-3} \text{ ms}^{-1}$ per metre height change, or by 0.56 ms^{-1} between surface and 170m. Referring to eq (29) this reduced the component of geostrophic wind perpendicular to the surface wind direction by about 7% in going from surface to 170m. No observations of the variation of the horizontal wind vector with height were made from HYDRA during ATEX, the nearest upper-wind measurements being carried out by the German research ship METEOR about 250Km to the South-East. Here the windfinding was done using a stabilised radar system but detailed data were not usually obtained at levels below about 800m because tracking could only be carried out beyond a minimum range of a few hundred metres. 18 wind finding ascents were made during the period covered by runs 2 to 6 and on average the surface wind was backed 11° from that at 1000m. If it is assumed that the veer of the actual wind from the surface wind increased linearly with height then this would have implied a reduction of 6% in $\partial \tau_{xz} / \partial z$ in going from surface to 170m: this and the effect of baroclinicity would give a variation of τ_{xz} of around 0.029 Nm^{-2} between surface and 170m. The change in τ would have been reduced by the compensating contribution from $\partial \tau_{yz} / \partial z$. However this last was even more difficult to calculate with any certainty because it depended chiefly on the differences of two quantities u_g and u which were of similar magnitude. Its surface (10m) value was around

$-4 \times 10^{-5} \text{ Nm}^{-3}$ using the magnitude of V_g , α and $\langle U_{10} \rangle$ as given above but an increase of V_g of only 2 ms^{-1} (to 12 ms^{-1}) changed it to $-9 \times 10^{-5} \text{ Nm}^{-3}$ (Charnock et al (1956) found the ratio U_{10}/V_g to be around 0.5 and applying this to the ATEX data would give $V_g \sim 15 \text{ ms}^{-1}$). Table 7 summarizes the calculations of $\partial \tau_{xz} / \partial z$, $\partial \tau_{yz} / \partial z$ for two values of V_g , and for $\alpha = 32^\circ$ and 20° : the wind used at 170m was the mean of those measured during the runs and was assumed veered 2° from the surface wind. The implied fluxes at 170m are all greater than the means of those which were observed

Table 8 Calculated variation of momentum flux with height, ATEX

	$\partial \tau_{xz} / \partial z \times 10^5 (\text{Nm}^{-3})$			$\partial \tau_{yz} / \partial z \times 10^5 (\text{Nm}^{-3})$			$\tau_{170m} (\text{Nm}^{-2})$
	Surface	170m	Mean	Surface	170m	Mean	
$V_g = 10 \text{ ms}^{-1}$ $\alpha = 32^\circ$	18	16	17	-4	+2	-1	-0.058
$\alpha = 20^\circ$	11	9	10	-7	-1	-4	-0.070
$V_g = 5 \text{ ms}^{-1}$ $\alpha = 32^\circ$	26	24	25	-18	-12	-15	-0.051
$\alpha = 20^\circ$	17	15	16	-23	-17	-20	-0.069

and for $\alpha = 20^\circ$ the differences are significant at about the 5% level. However, in view of the uncertainties in the values of V_g and α which should be used the results from Table 8 offer no more than very slender evidence for a systematic over-estimation of the measured flux by the tethered balloon technique. On the other hand the observed dip in the negative cospectra between 0.01Hz and 0.05Hz is contrary to all expectations and was due almost certainly to shortcomings of the balloon-borne system: better flux estimates would be likely to result from smoothing it out and doing so decreased the normalised means to -0.057 Nm^{-2} , a value not significantly different from any of those in the table.

Some indication of the accuracy of the heat fluxes is given by the results from run 5. Here a dummy temperature element (a metal oxide resistor unshielded from solar radiation) was used and the residual "flux" calculated from the output of this sensor and the vertical velocity was found to be very small, around 2 Wm^{-2} . Table 5 shows that the majority of the contribution

to this came from frequencies below 0.003 Hz, or almost certainly at lower frequencies than those of the characteristic tethered balloon motions. An explanation for the "flux" being non-zero is that changes of wind speed produced variations in the temperature excess of the resistor caused by absorption of solar radiation: the resistor had a small temperature coefficient and hence there were resultant resistance variations correlated with horizontal and therefore vertical wind speed in the correct sense to give a positive "flux". With a proper sensor the noise limit in the measured heat flux would probably have been smaller than 2 Wm^{-2} . The results from run 6 are not thought to be reliable because of significant cross-talk between temperature and humidity channels on this occasion: corrections were applied for this but the computed fluxes were very small and even the sign must be open to doubt. Tw-cospectra for runs 3 and 4 are plotted in figure 13. They show considerable scatter at higher frequencies with substantial contributions between about 0.05 and 0.3 Hz due presumably to the effect of ship motion. On the other hand the calculated fluxes were less dependent on averaging time than those for momentum (see Table 5) because of the small effect of sensor motion on the temperature signal compared to the spurious contribution to the measured horizontal speed. The fluxes obtained from 8-second averaged data (-8.5 and -2.7 Wm^{-2} respectively) would have included insignificant contributions from sensor motion but were probably underestimates by a few percent because averaging would have filtered out genuine turbulent contributions just below $f=4$. The measured air-sea temperature differences were the same for each run, implying similar (small) surface fluxes in each case, and it appears therefore that there was an increasing downward flux with increasing height. This is entirely consistent with a simple boundary-layer model described in § 7.9. It is perhaps interesting to note that the apparent flux convergence in the lowest 180m (about 10 Wm^{-2}) would have been sufficient to raise the temperature by about 5 degrees

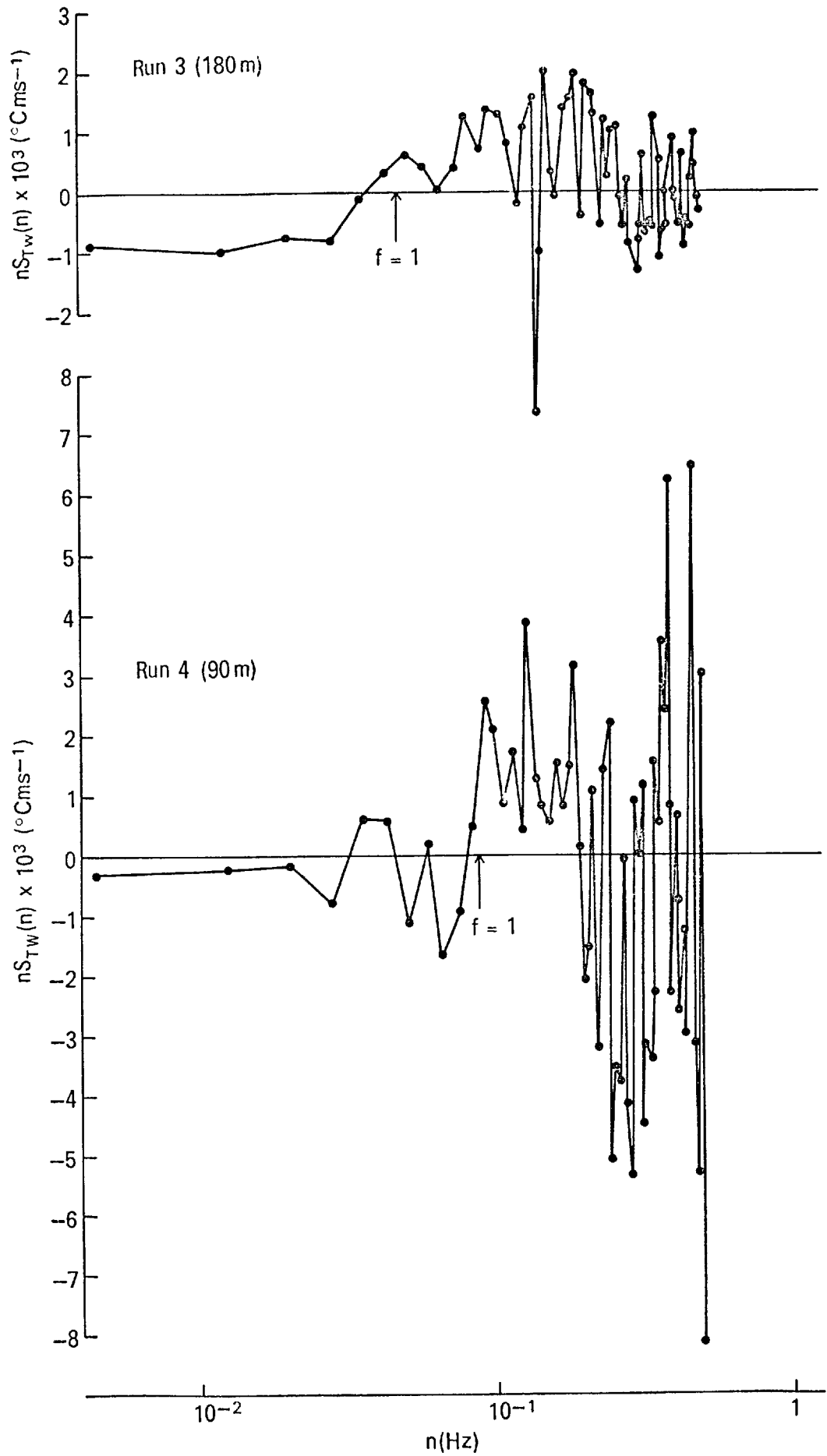


Figure 13 Tw cospectra ATEX Runs 3 and 4

C per day in the absence of radiational cooling: the net warming rate would have been about 4 degrees C per day which would have soon produced stable density stratification at the surface since the likely rate of increase of sea-surface temperature along the expected trajectory was around 1 degree C per day. This suggests that the measured flux for run 3 was unrepresentatively high, perhaps as a result of local unsteadiness or because of sampling errors or alternatively the surface heat flux was directed downwards. In this latter case this could have resulted from a cool "skin" at the sea surface due to loss of heat by evaporation, the skin temperature being perhaps 0.3 degrees C lower than the bulk water temperature (Hasse 1971). Using a surface flux calculated from the corresponding adjusted air-sea temperature difference reduces the implied rate of heating of the air above to around 2 degrees C per day (probably less at night due to cooling of the sea surface and hence increased downward heat flux).

The single water-vapour flux cospectrum (run 6) is shown in figure 14. This is not thought to be very reliable as the water supply to the wet bulb did not appear to be correctly adjusted. The cospectrum shows comparatively small obviously spurious contributions at reduced frequencies below $f=1$ compared with the T_w -cospectra and in these circumstances the 8-second averaged data should have given a representative moisture flux in the absence of unreliability of the wet-bulb data. If the measured flux was assumed to be representative of that at the surface then using the observed air-sea specific humidity difference and calculated 10m wind in eq (9a) leads to a value of 1.1×10^{-3} for C_E , close to that expected.

Wind spectra (figures 15 (a), 15 (b)) show clearly the major contributions from sensor motion at frequencies above about 0.07Hz, but there are obviously spurious contributions in the case of both U and w at frequencies

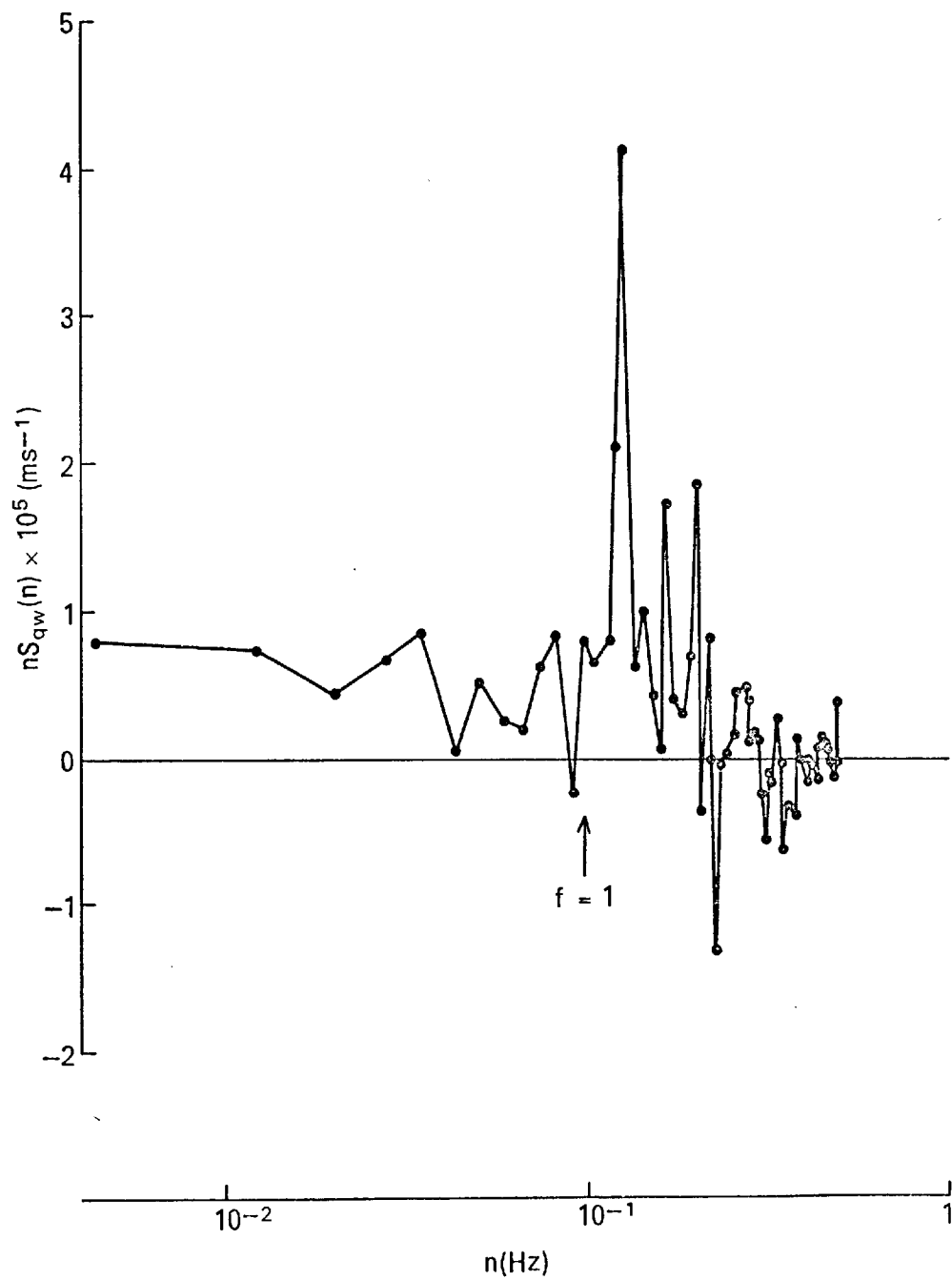


Figure 14 qw cospectrum Run 6 ATEX

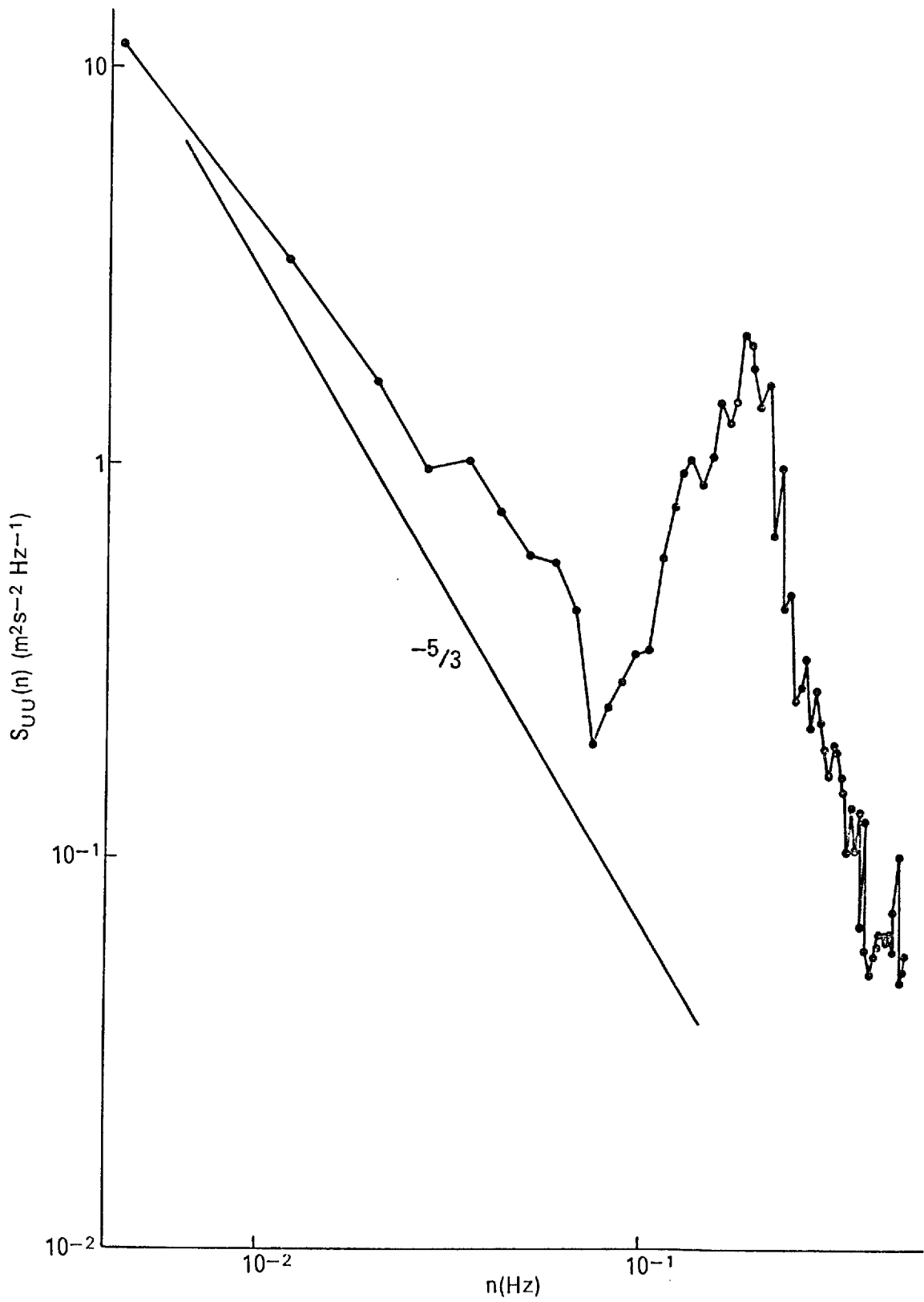


Figure 15(a) U – spectrum Run 6 ATEX

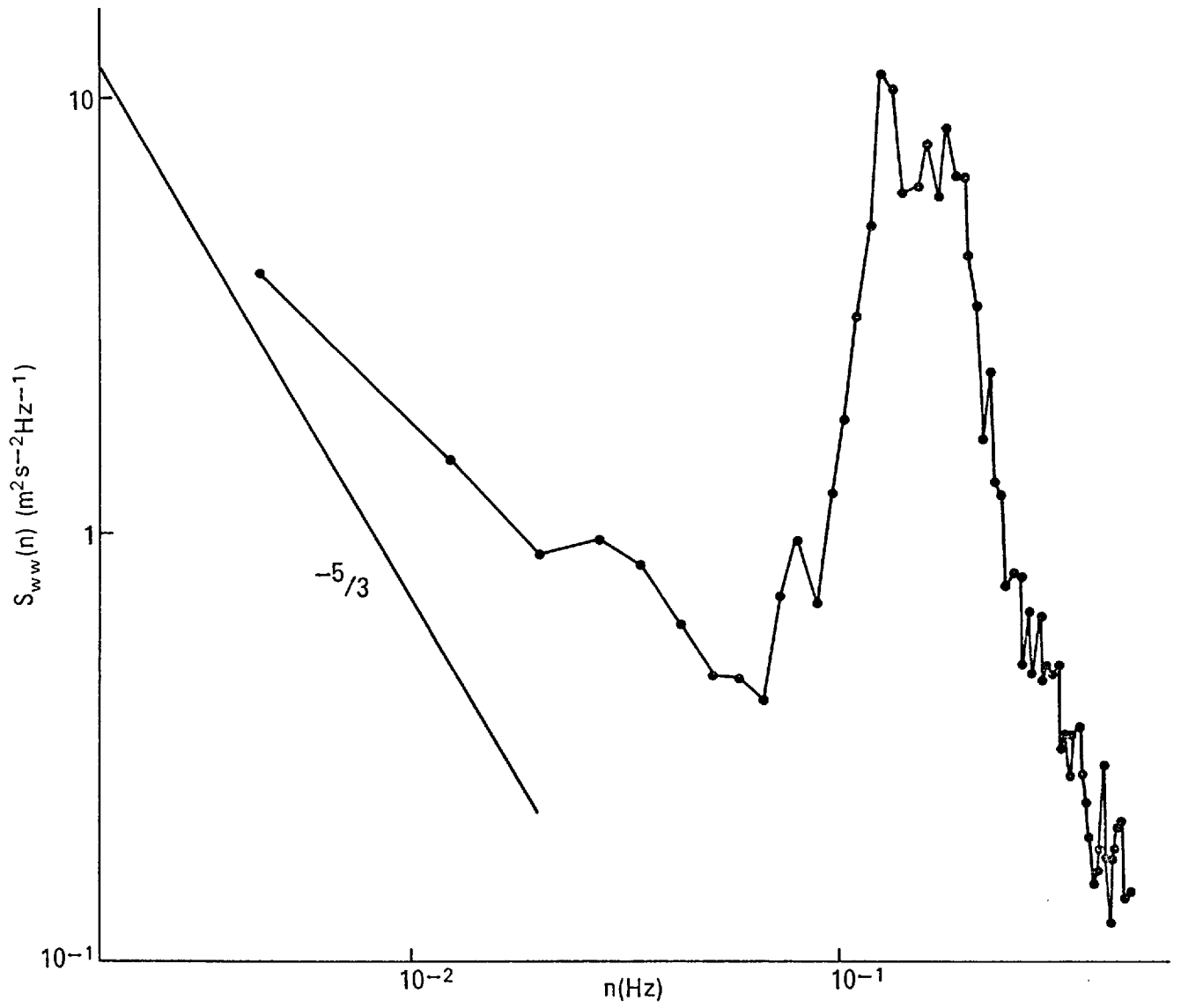


Figure 15(b) W – spectrum Run 6 ATEX

between about 0.02 and 0.07 Hz: it is these last which presumably lead to the observed dip in the negative cospectra for momentum flux. The contamination of the w-spectra is great enough to obscure the expected $-5/3$ slope of the inertial subrange, though it still appears at sufficiently low frequencies in the U-spectra. The standard deviations of the velocity fluctuations may be represented approximately by values calculated from 8-second averaged data, with the genuine contributions lost by the filtering being compensated to some extent by the spurious contributions below 0.07Hz. Table 9 gives values for the ratios σ_u/u_* and σ_w/u_* using these standard deviations in conjunction with estimates for u_* obtained after smoothing out the dip in the negative cospectra. Apart from the entry for σ_w/u_* for run

Table 9 Scaled standard deviations of turbulent velocities from ATEX

Run No	σ_u / u_*	σ_w / u_*
2	2.6	2.3
3	2.5	1.0
4	2.3	1.2
5	1.9	1.0
6	2.7	1.2
averages	2.4	1.3

2 the scatter is small and the averages are close to those found in the constant-flux layer in near-neutral conditions (McBean 1971, Pond et al 1971, Miyake et al 1970a, b: it should be noted that the mean value of 2.97 for σ_u / u_* given by Pond et al was an overestimate because of spurious contributions to σ_u from motion of the instrument platform), suggesting that the calculated momentum fluxes were realistic in value though of course u_* is less sensitive to errors than τ .

Despite the instrumental problems which were encountered the results from ATEX in general showed a promising approach to consistency and suggested

that the tethered balloon technique even in prototype form was capable of providing reasonably convincing estimates of vertical eddy fluxes in winds up to at least 8 ms^{-1} .

CHAPTER 7

JASIN: RESULTS AND DISCUSSION7.1 Records of turbulence

Low-pass filtered records of turbulence data obtained simultaneously at heights of 90m and 185m on a convective occasion are shown in figure 16. An important feature of the wind vector and inclination traces is that the easily-recognisable spurious contributions due to sensor motion (which are large enough to mask almost completely the genuine turbulent fluctuations) are significantly greater at the lower height. This demonstrates that sensor motion results primarily from alternate straightening and relaxing of the balloon cable catenary (the balloon was about 30m above the higher instrument and the lower instrument was close therefore to the midpoint of the cable). Consequently in multi-level measurements there will be a decrease in accuracy of measurement of turbulence parameters as sensors are moved closer to the middle of the balloon cable. The frequency of the spurious fluctuations is typically greater than 0.1Hz but there are clearly significant departures from this (particularly for inclination) and significant changes in amplitude also: The variations occur because the ship's motion depends on both its natural period and the period and amplitude of the forcing function - the wave field. As expected the temperature and wet-bulb depression records demonstrate a much better signal-to-noise ratio. However the temperature fluctuations are very small and this signal therefore suffers more from higher-frequency noise introduced by the telemetry and recording equipment. A striking feature is the apparent lack of correlation of, for example, the wet-bulb depression records at the two heights. It is interesting to note that at 185m the warmer sections of temperature records marked A and C are associated with humidities lower than average and a colder section (B) with a higher humidity: making the reasonable assumption that the dryer air originated from above the level and the wetter from below this would imply that the vertical heat and moisture

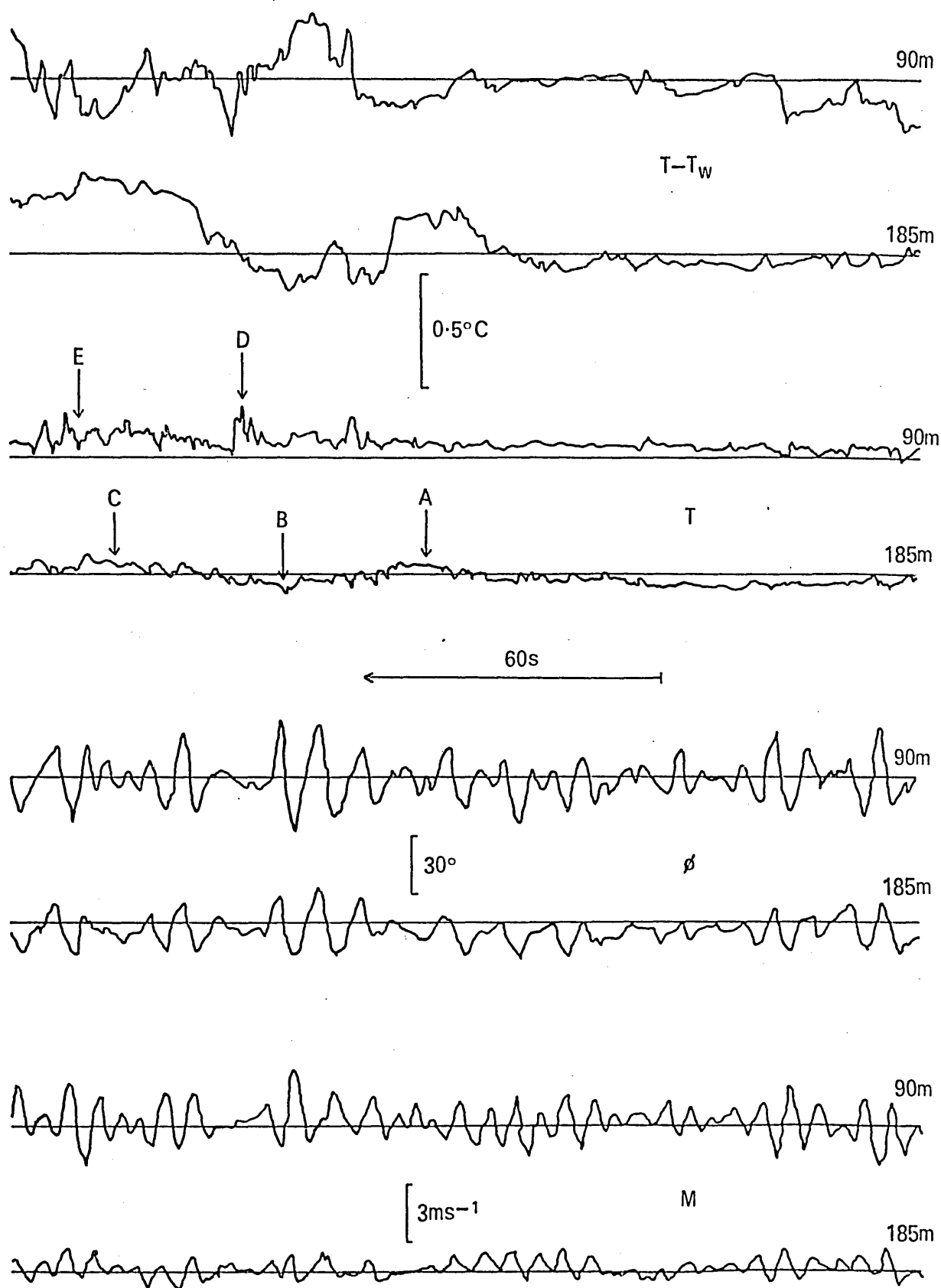


Figure 16 Turbulence records at 90m and 185m at 2300Z, 17.6.70 (JASIN). $\langle U_{10} \rangle \approx 5.7\text{ms}^{-1}$

fluxes were of opposite sign, at least for the short period under consideration. In contrast the record at 90m is more consistent with the fluxes of both quantities being directed upwards: thus at D and E the air is both warmer and moister than on average.

7.2

SPECTRA

The main feature of the measured spectra are illustrated by figures 17-20: these show spectra of U , w , T and q obtained simultaneously at heights of 90m and 185m in a convective situation. Dots are averages of the individual spectral estimates obtained with the data subdivided into lengths of about 15 minutes. These averages were themselves smoothed across frequency in blocks of either 16 (between 0.1 and 0.5Hz) or 8 (at lower frequencies) before being plotted. 90% confidence limits for an arbitrary size of estimate are shown on figure 17 by the vertical bars together with the frequency regions to which they apply. Crosses were averages of either 8 or 16 estimates obtained using the full length of record without subdivision, as explained in § 5. Here the confidence limits were necessarily larger.

Figures 17 and 18 show clearly the overwhelming contribution due to sensor motion at frequencies above about 0.1Hz. This contribution was significantly less at the greater height, demonstrating that the sensor motion was predominantly horizontal as explained earlier. In the present case the " $n S(n)$ versus $\log n$ " form of plotting tends to conceal some of the finer detail at lower frequencies though of course it gives a readier impression of the contribution of noise to the measured variance than does a plot of $S(n)$: in spite of this a comparison with figures 15 (a) and 15 (b) reveals many similar features (and gives some confirmation of the validity of the ATEX data). In each case the major peaks occurred at similar frequencies for U and w , with the w -peaks the larger and broader. Probably of more importance is that identifiable noise appears in both sets of data at frequencies substantially less than the main peak. This is shown best in the JASIN data by figure 18 where lower frequency estimates have been plotted on an enlarged scale. There are obviously peaks at

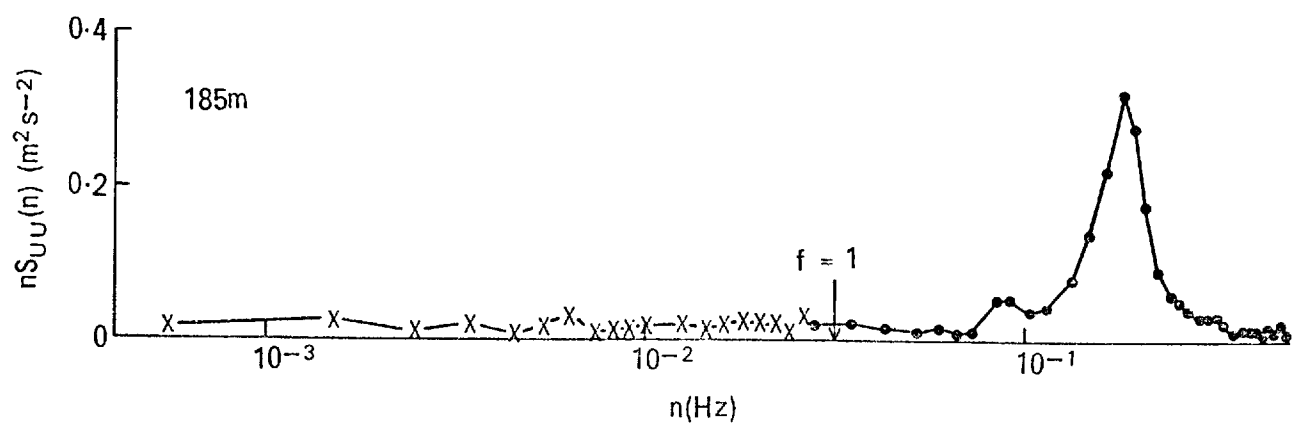
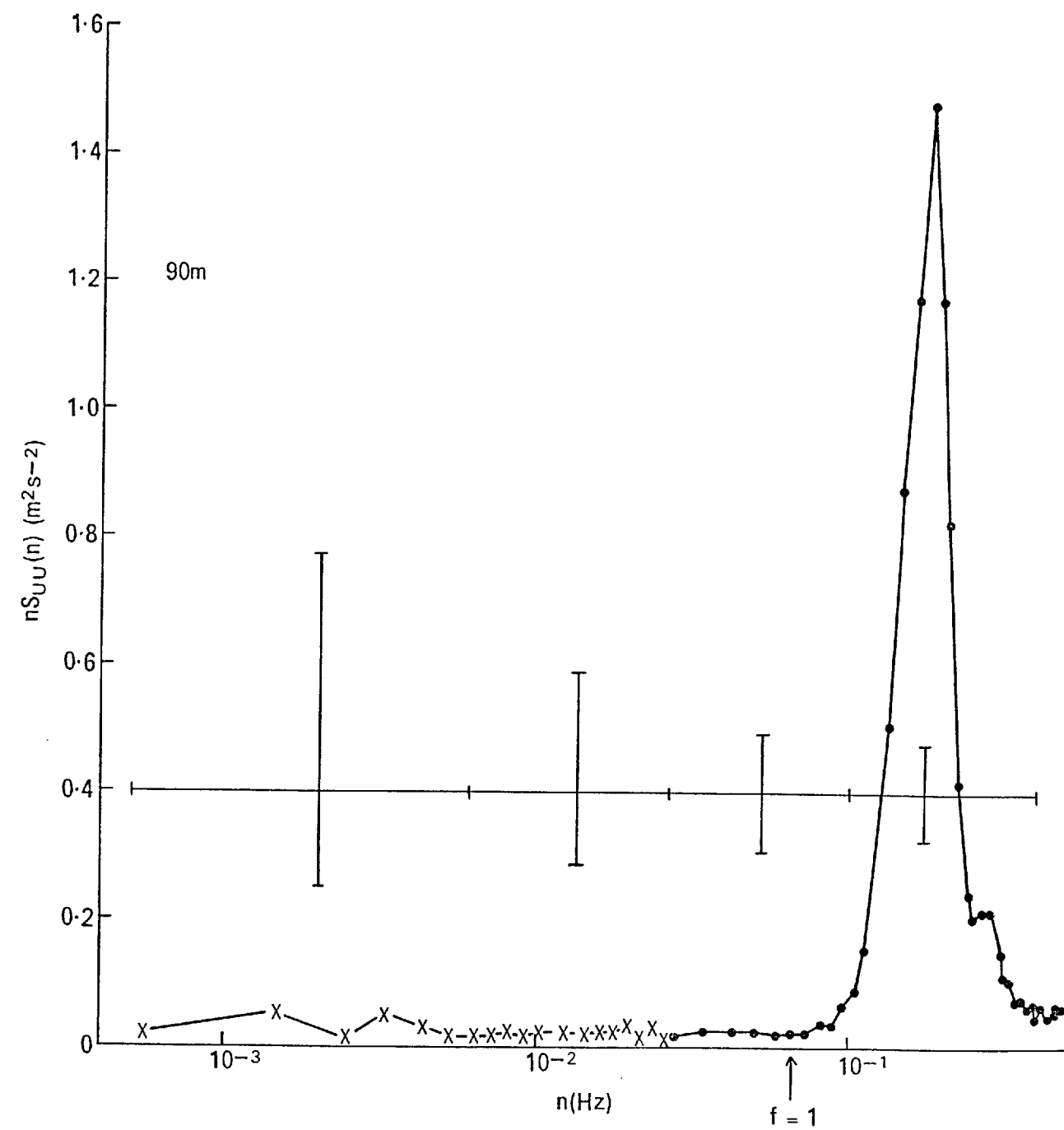


Figure 17 Composite U – spectra, 2143 –0005 Z, 17.6.70 (JASIN Run 13)

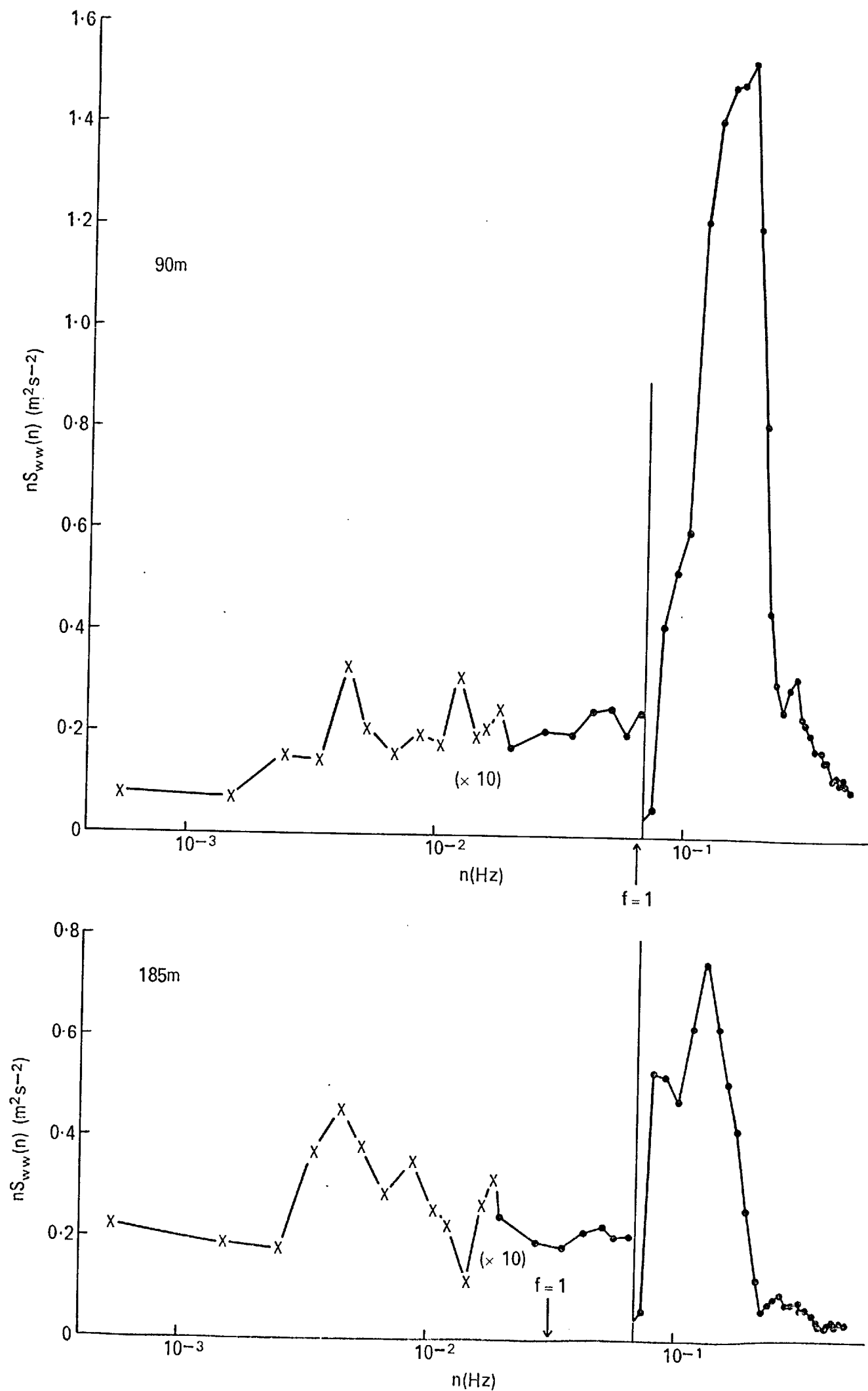


Figure 18 Composite W-spectra, 2143 -0005 Z, 17.6.70 (JASIN Run 13)

around 0.09Hz and 0.05Hz, frequencies close to the subharmonics of the main peak (there are probably peaks at 0.02 and 0.03 Hz at 90m also). It is possible therefore that the explanation given in § 6 for the peaks (characteristic tethered-balloon motions) is incorrect and they are in fact due to ship motion - their low frequency would be explained if they were the result of occasional optimum phasing between the ship's natural oscillations and the sea waves leading to much lower-frequency periodicities in the data.

Temperature spectra (figure 19) are almost flat between reduced frequencies of about 1 and 10^{-2} , a feature found by Miyake et al (1970a) from measurements at a height of about 2m above a water surface. There are clear indications of noise due to ship motion at frequencies above about 0.06Hz. It was caused by vertical sensor displacements in an environment with a temperature gradient close to dry adiabatic (-0.01 degrees C m^{-1}). Not unexpectedly, since the sensor motion was predominantly horizontal the noise was relatively small: it is seen most clearly at frequencies around 0.09Hz. As anticipated there was considerable telemetry and recording noise at the highest frequencies. This resulted from using a large telemetry bandwidth (about ± 1.5 degrees C) to allow for both errors in predicting before each experiment the mean temperature at measurement level as well as turbulent fluctuations: excursions outside this bandwidth had to be avoided to prevent significant cross-talk with other channels in the telemetry system. The standard deviation of the fluctuations was typically around 0.1 degrees C and so the signal/noise ratio was inevitably unfavourable. It was for this reason that a high-pass filtered signal was transmitted also: here a bandwidth of about ± 0.5 degrees C could be used, giving a substantial improvement.

Humidity spectra (figure 20) showed considerably less influence from ship motion. This follows from similarity arguments using the following

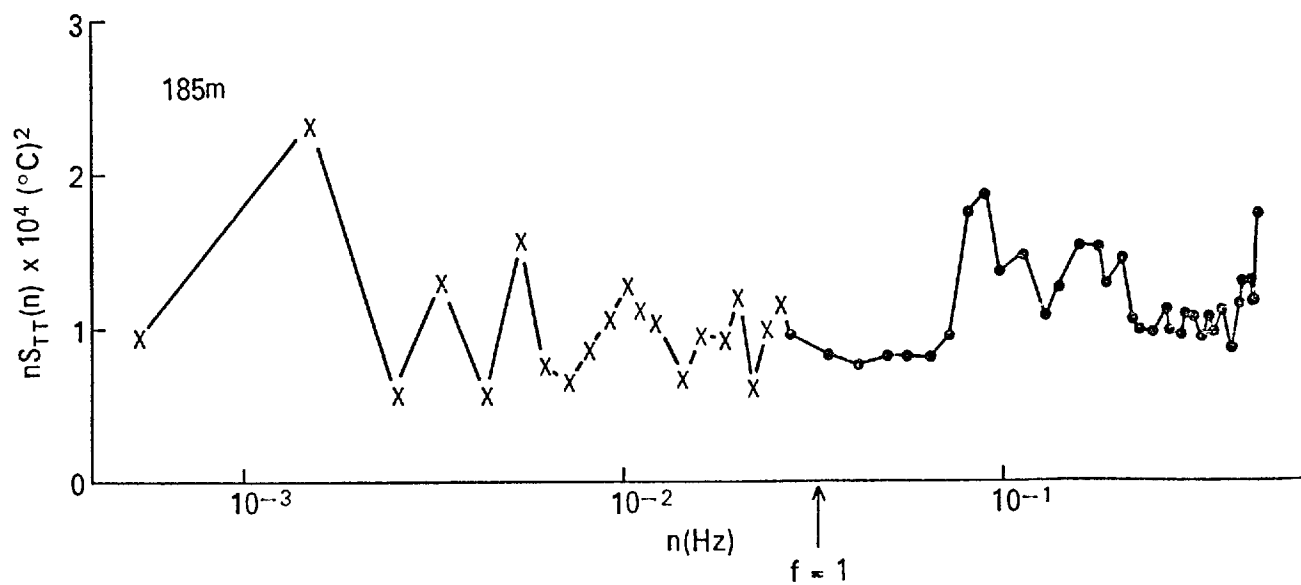
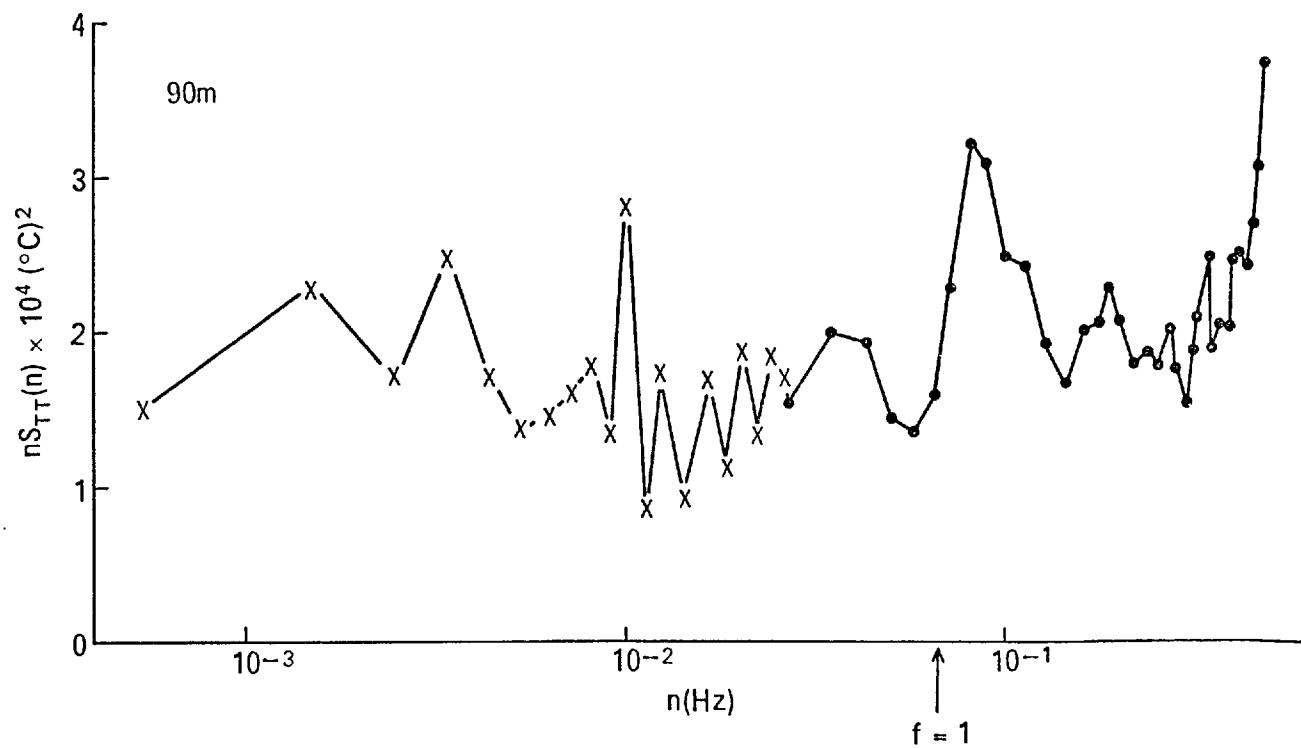


Figure 19 Composite T-spectra, 2143 - 0005 Z, 17.6.70 (JASIN Run 13)

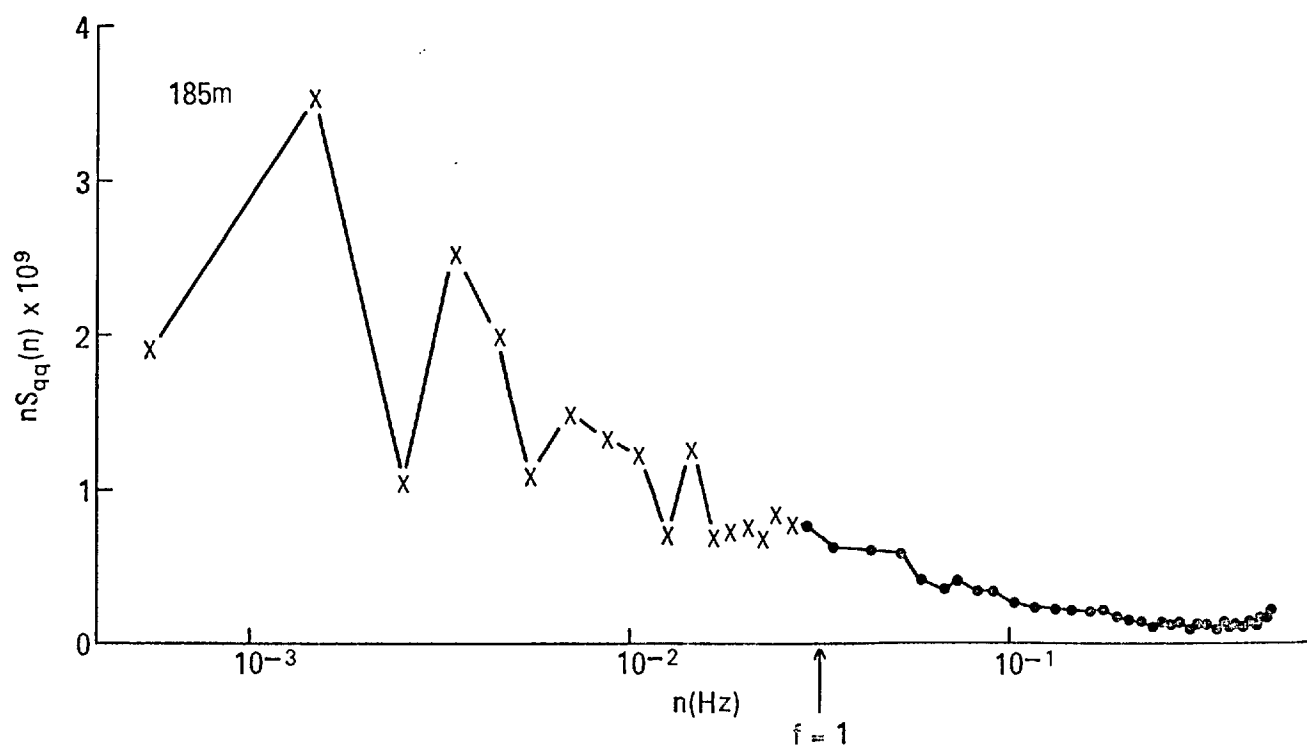
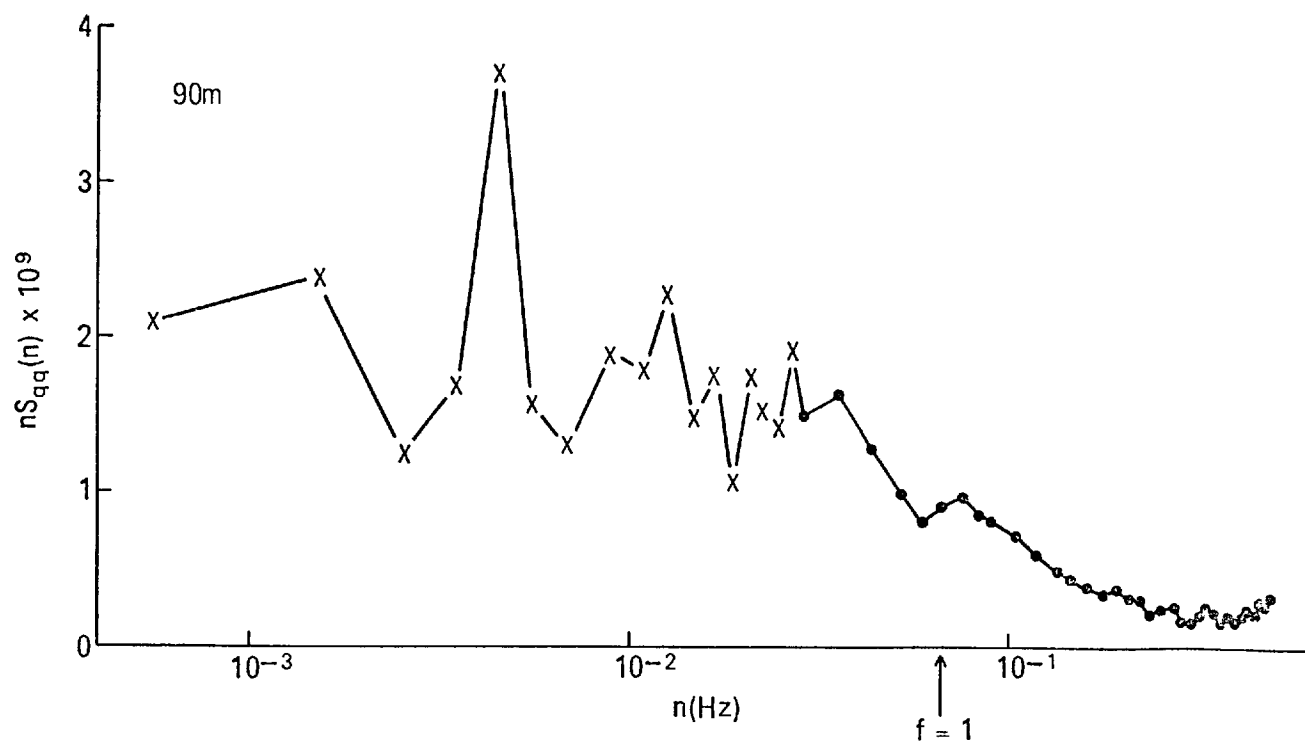


Figure 20 Composite q-spectra, 2143 -0005Z, 17.6.70 (JASIN Run 13)

approximation

$$\sigma_T / T_* \sim \sigma_q / q_*$$

; hence

$$\sigma_T / \sigma_q \sim (\partial\theta/\partial z) / (\partial q/\partial z) \quad (31)$$

Vertical displacements of the sensors produced fluctuations of T and q proportional to $\partial T/\partial z$ and $\partial q/\partial z$ respectively and since $|\partial T/\partial z| \gg |\partial\theta/\partial z|$ (lapse rates were usually close to dry adiabatic) the resulting spurious fluctuations of temperature were a larger fraction of the true variances than in the case of humidity. There was some telemetry and recording noise at the highest frequencies but apart from this the ultimate slope (on a log-log plot) was less than the $-2/3$ expected for the inertial subrange: this was because of fall-off of the wet-bulb sensor response at higher frequencies. There is no indication in the diagram of a decreasing contribution to the variance with decreasing frequency even at the lowest plotted frequency (f less than 10^{-2}). If the Taylor transformation is valid at such low frequencies this last corresponds to a horizontal scale greater than 10Km. The spectra are broadly similar to those found by Miyake et al (1970b) at a height of 18m using an instrumented aircraft.

Earlier investigators (eg. Swinbank 1951, Taylor 1963) have commented on the high correlation between temperature and humidity fluctuations measured close to the surface over land but figure 16, and a comparison of figure 19 and 20 do not support this conclusion at greater heights over the sea. Thus the variation of humidity variance with height is clearly not large, but in contrast the temperature fluctuations are obviously smaller at the greater height. A possible explanation is that specific humidity in these conditions is a well-behaved function of height, decreasing from surface upwards through the whole boundary layer, whereas potential temperature shows a more complex

structure with a layer near the surface obeying similarity theory, a region above this with nearly constant Θ , and higher up an increase of Θ with height: a rising parcel of air originating at the surface would be both warmer and moister than its surroundings at first and the humidity difference would tend to be maintained during its upward motion because of the humidity lapse but in contrast the temperature excess would decrease once above the region of constant Θ even in the absence of mixing.

U and w-spectral estimates multiplied by frequency and normalised by u_*^2 are shown plotted against reduced frequency in Figures 21 and 22: each includes 11 spectra, all obtained during the latter part of JASIN on occasions when u_* could be determined with some confidence (the calculation of u_* is discussed in detail in § 7.4). The spectra are not extended to natural frequencies above about 0.1Hz to eliminate the major contributions from sensor motions. The considerable variability displayed by the U - spectrum at low frequencies is in broad agreement with the results of, for example, Miyake et al (1970 a), and the w-spectrum shows the expected fall-off in this region although spectral densities for values of f between 10^{-1} and 10^{-2} are somewhat larger than data obtained at lower heights (eg. Miyake et al 1970a). At high frequencies noise due to sensor motion dominates the spectra.

It is customary to normalise spectral estimates for temperature by σ_T^2 when plotting "universal" T-spectra but in the present case because of sensor motion which could not be filtered out convincingly a normalisation by T_*^2 was used. The heat fluxes are discussed in §7 (e) where it is pointed out that rapid variations in H with height (sometimes even change of sign) can occur. In these circumstances T_*^2 also may not be a satisfactory normaliser especially when heat fluxes are very small. For this reason figure 23 includes only 4 spectra, obtained on occasions with the largest

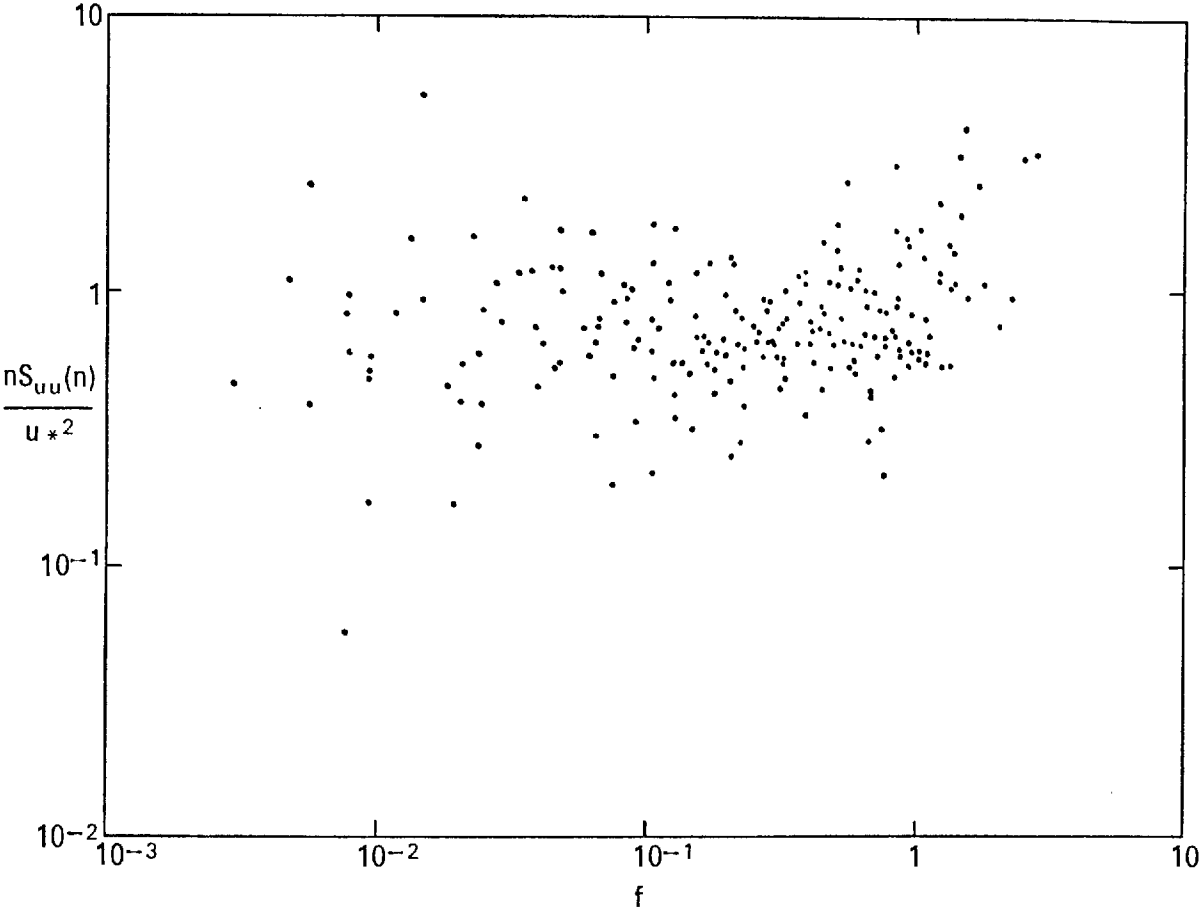


Figure 21 Normalised U – spectra

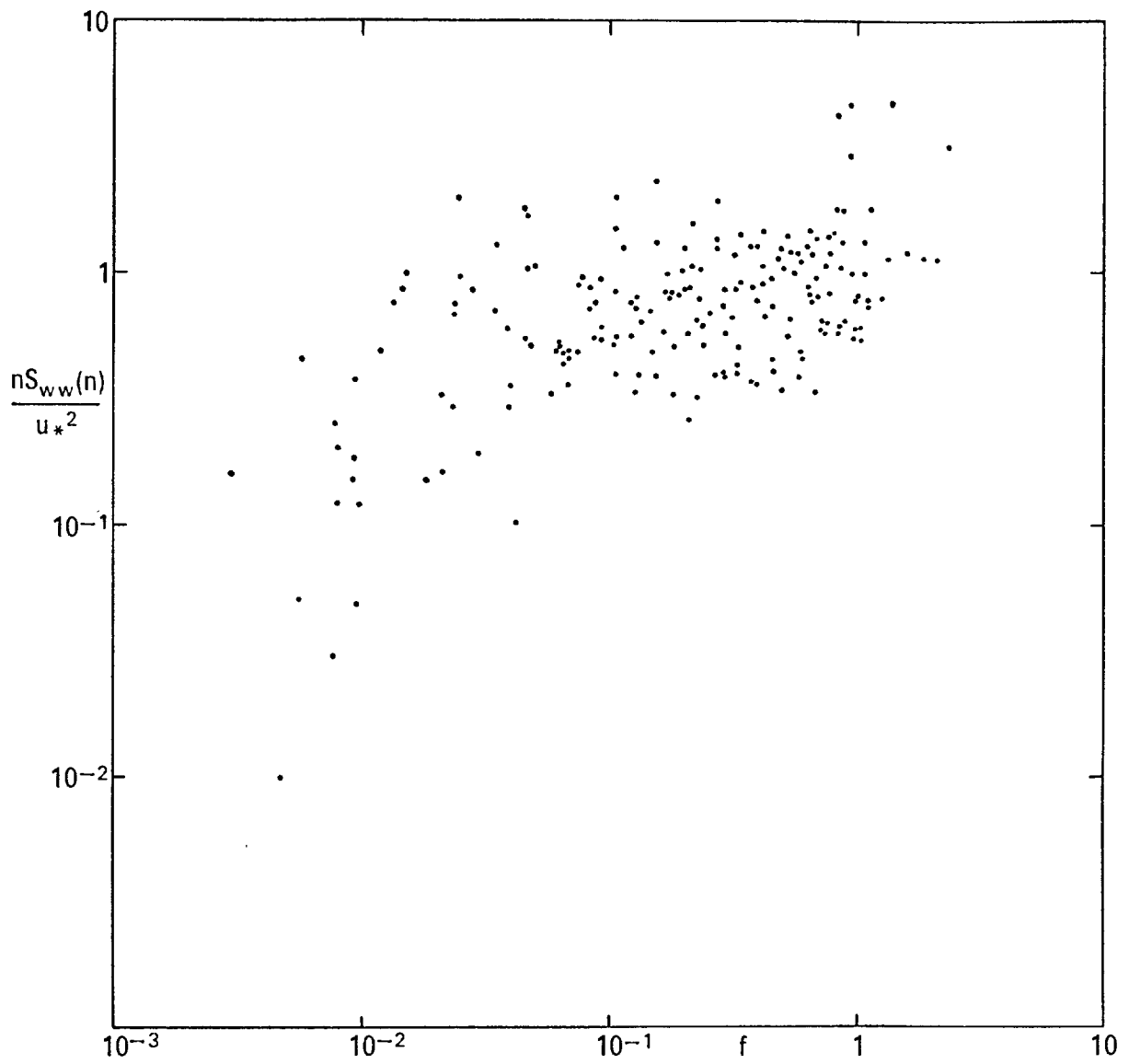


Figure 22 Normalised W – spectra

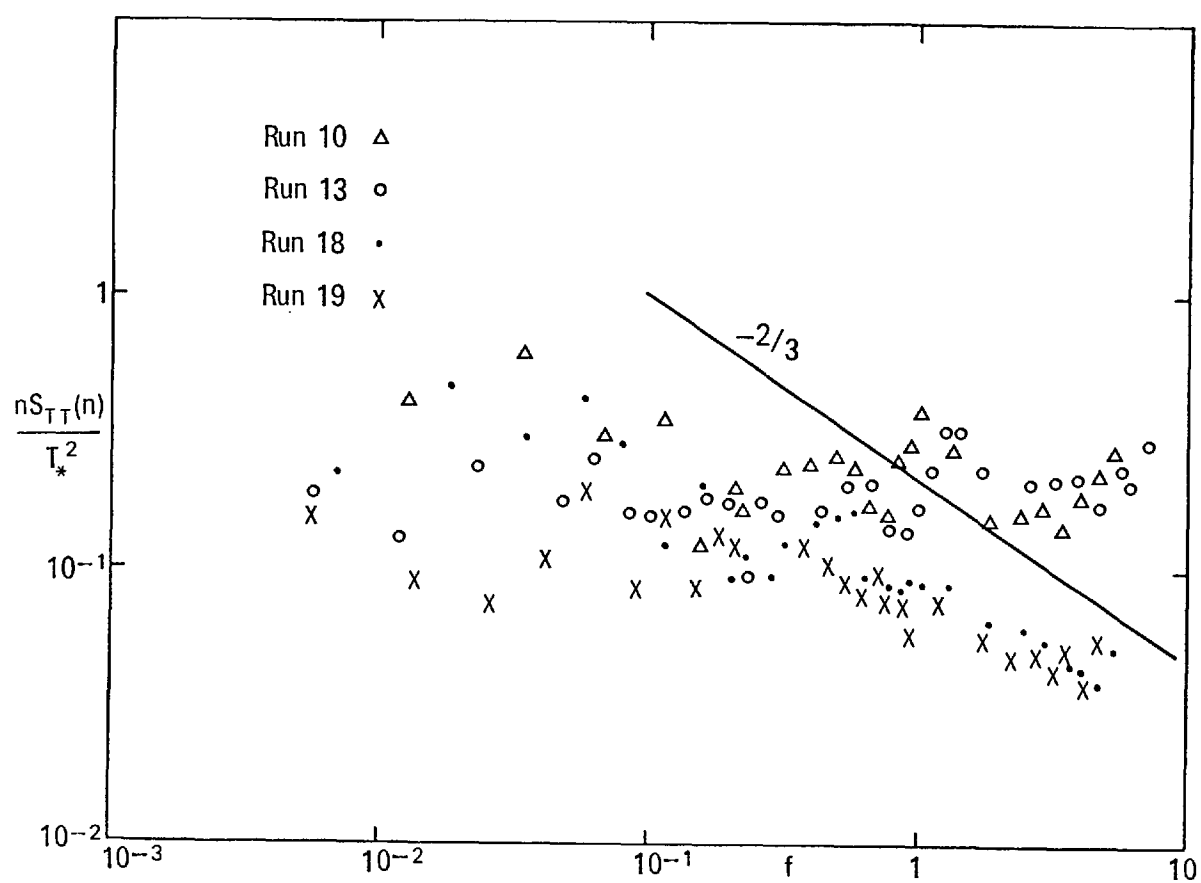


Figure 23 Normalised T – spectra

measured heat fluxes. In two cases (runs 18 and 19) u_* could not be determined from the data with any reliability and here the values used to calculate T_* were based on the estimated surface value with a correction for instrumental height. All four spectra show considerable scatter and overlap at low frequencies but there are systematic differences at high frequencies. Here two of the spectra have slopes approaching the minus two-thirds expected in the inertial subrange and, using the approximation $\sigma_T^2 \sim T_*^2$ also have spectral levels similar to those shown in figure 3. The much-higher normalised values for the other two spectra in this region are due, at least in part, to less-favourable signal-to-noise ratios since in each case the values of the unnormalised estimates were only about half of those for runs 18 and 19.

Humidity spectra normalised by σ_q^2 are shown in Figure 24 for the latter part of JASIN: 9 spectra have been plotted but data from runs 11 and 12 have not been included because there was evidence here for spurious temperature (and hence humidity) fluctuations being produced by raindrops occasionally shorting the windings of the temperature element (data from run 16 when sensor motion was very violent are also not plotted). The large fluctuating contributions at low frequencies reflect presumably meso-scale variability in the humidity field and probably because of this the normalised spectra show a large scatter at high frequencies. The spectra all have significant noise contributions at the highest frequencies but apart from this the ultimate slopes are rather steeper than the $-2/3$ expected for the inertial subrange, presumably because of the wet-bulb sensor's finite time constant.

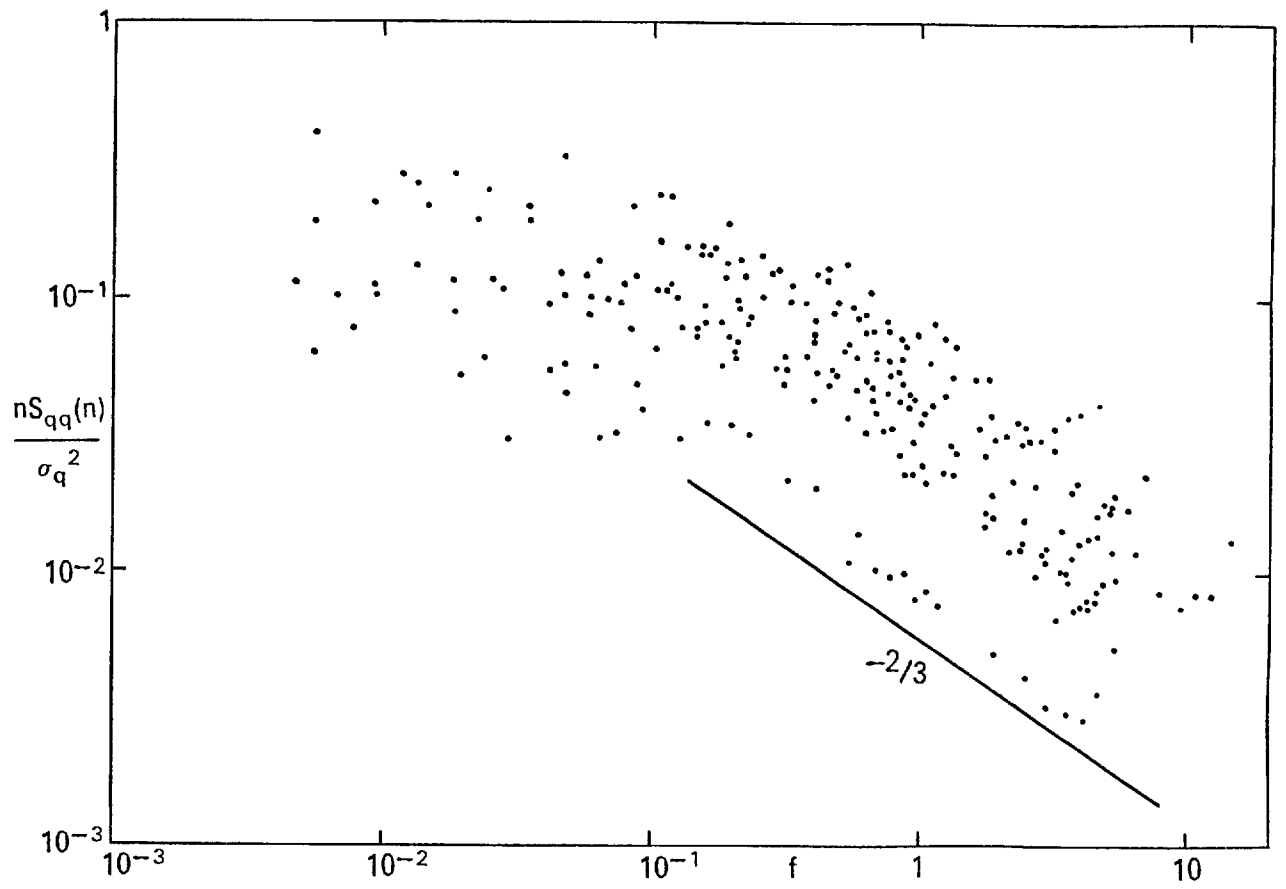


Figure 24 Normalised q -spectra

7.3

COSPECTRA

Before discussing measured cospectra in detail it is instructive to consider the validity of the "universal" curves shown in figure 4. It was pointed out in § 3 that reduced-frequency scaling of spectra (and presumably cospectra) breaks down above about 50m with the value of f at the peak of the vertical velocity spectrum probably increasing by a factor of about 1.5 in going from close to the surface to about 200m, but with a somewhat smaller shift at higher frequencies. For the present purpose this may be considered negligible when discussing cospectral cut-offs at high frequencies. On the other hand to use figure 4 to decide likely low-frequency limits for cospectra if measurements are made hundreds of metres above the surface will almost certainly lead to difficulties. If it is assumed that cospectral estimates are negligible beyond $f = 10^{-3}$ then it is easy to calculate the length of record required for reasonable statistical reliability of the calculated estimates at low frequencies. Thus at a height of 200m in a wind of 5ms^{-1} the record length would have to be at least 1300 minutes to resolve $f = 10^{-3}$ and several times longer (say in excess of 5,000 minutes) for acceptable statistical significance! On the other hand a simple physical argument suggests that the low-frequency cut-off may be at much-higher reduced frequencies at these greater heights. Thus in convective conditions estimates will be reliable if a sufficiently large number of cumulus clouds pass over the instrumentation during the sampling period. If the clouds are randomly distributed and spaced say 2Km apart (the spacing would of course vary with the clouds' vertical development) then a sampling time of say 2 hours would probably be sufficient in a wind of 5ms^{-1} . This is about one and a half orders less than the earlier estimate which thus appears unduly pessimistic (practically of course cumulus clouds are often larger and more widely spaced, or else

organised into systems such as cloud streets and in these circumstances a somewhat longer sampling period will be necessary to reduce statistical variability). It seems plausible therefore that the "universal" curves of figure 4 will become grossly in error on the low-frequency side above heights of a few tens of metres, with a general tendency for the cospectra to be telescoped towards higher frequencies.

Uw-cospectra (figure 25) show very large contributions from sensor motions as anticipated from the component spectra but it is seen that the great majority of the noise appears in regions where the contributions from genuine fluctuations are small. A comparison with figure 11 (a) shows much greater noise at the upper level than during ATEX at a similar height, implying considerably more ship motion during JASIN: it is seen also that there are positive cospectral contributions at frequencies around 2×10^{-2} Hz in the case of both ATEX and JASIN data at around 180m - possible explanations for these have already been given.

The much smaller size of the spurious contributions to the T-spectrum than to the U-spectrum, the fact that tethered-balloon motion produces Tw-correlations almost certainly smaller than those for Uw, and the larger correlation coefficients for turbulent fluctuations of temperature and vertical wind speed than for U and w (Haugen et al 1971) all contribute to the much smaller influence of sensor motion on the Tw-cospectra (figure 26). If noise-free cospectra have effective cut-offs at around $f = 4$ it is clear that subjective smoothing of cospectral estimates, rather than electrical filtering, is needed in order to minimise the contributions due to sensor motion, but in spite of this it is reasonable to expect errors of only a few percent in the measured heat fluxes if these have been derived from smoothed cospectra. Figure 27 demonstrates most effectively how reduced-frequency scaling can break down, in this case for quite different reasons

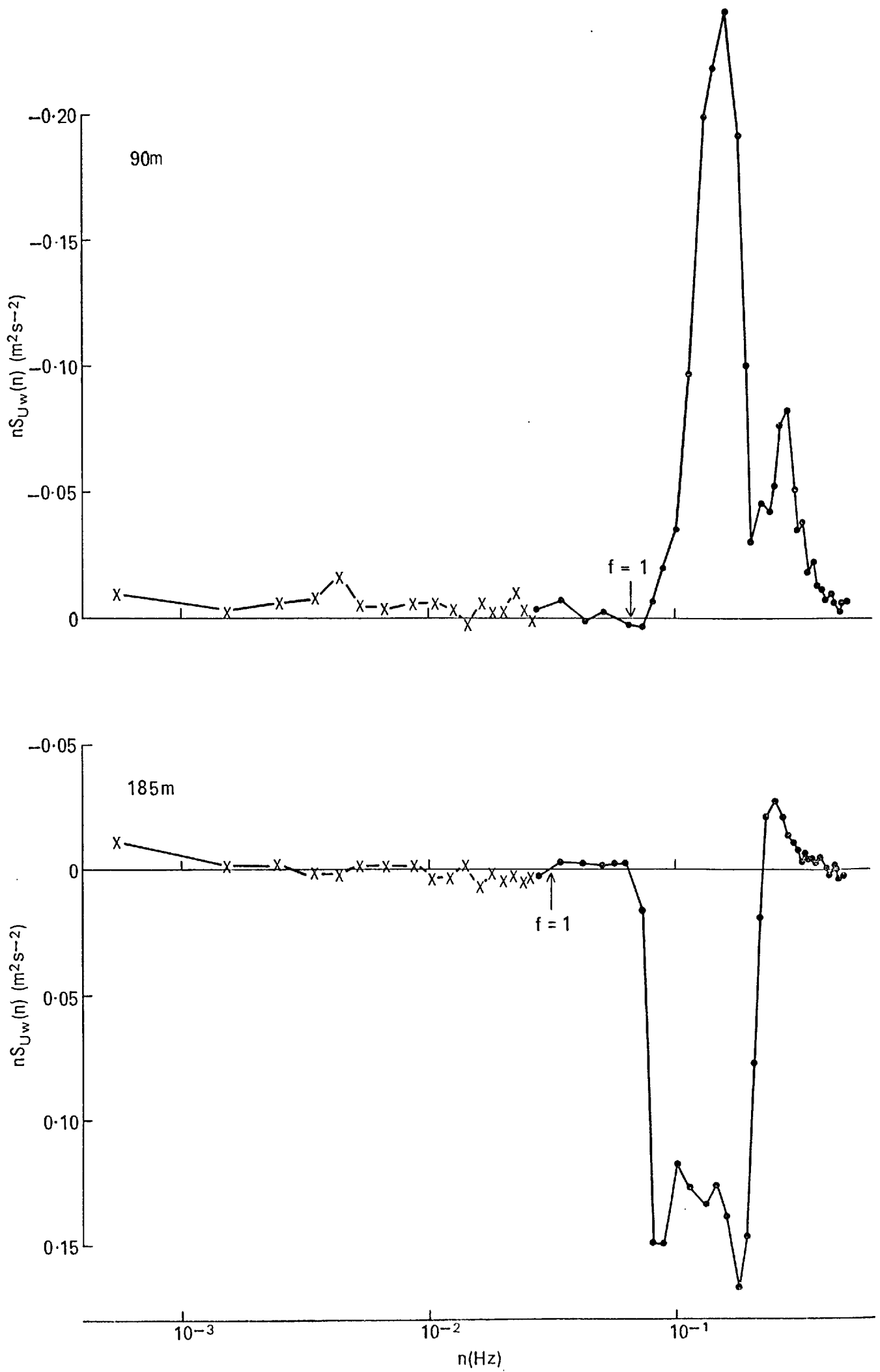


Figure 25 Composite Uw – cospectra, 2143 – 00052, 17.6.70 (JASIN Run 13)

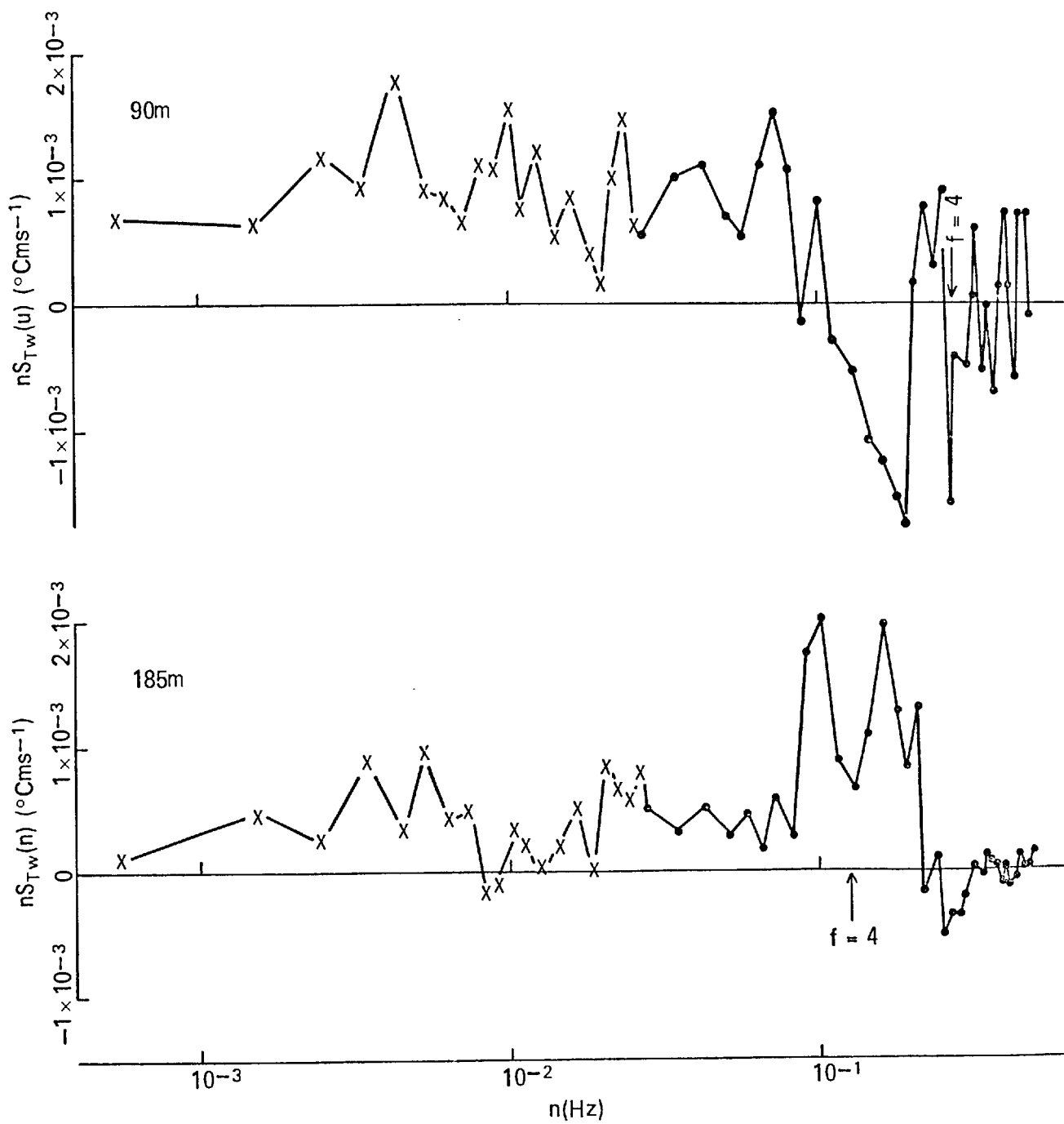


Figure 26 Composite T_w - cospectra, 2143 -0005Z, 17.6.70 (JASIN Run 13)

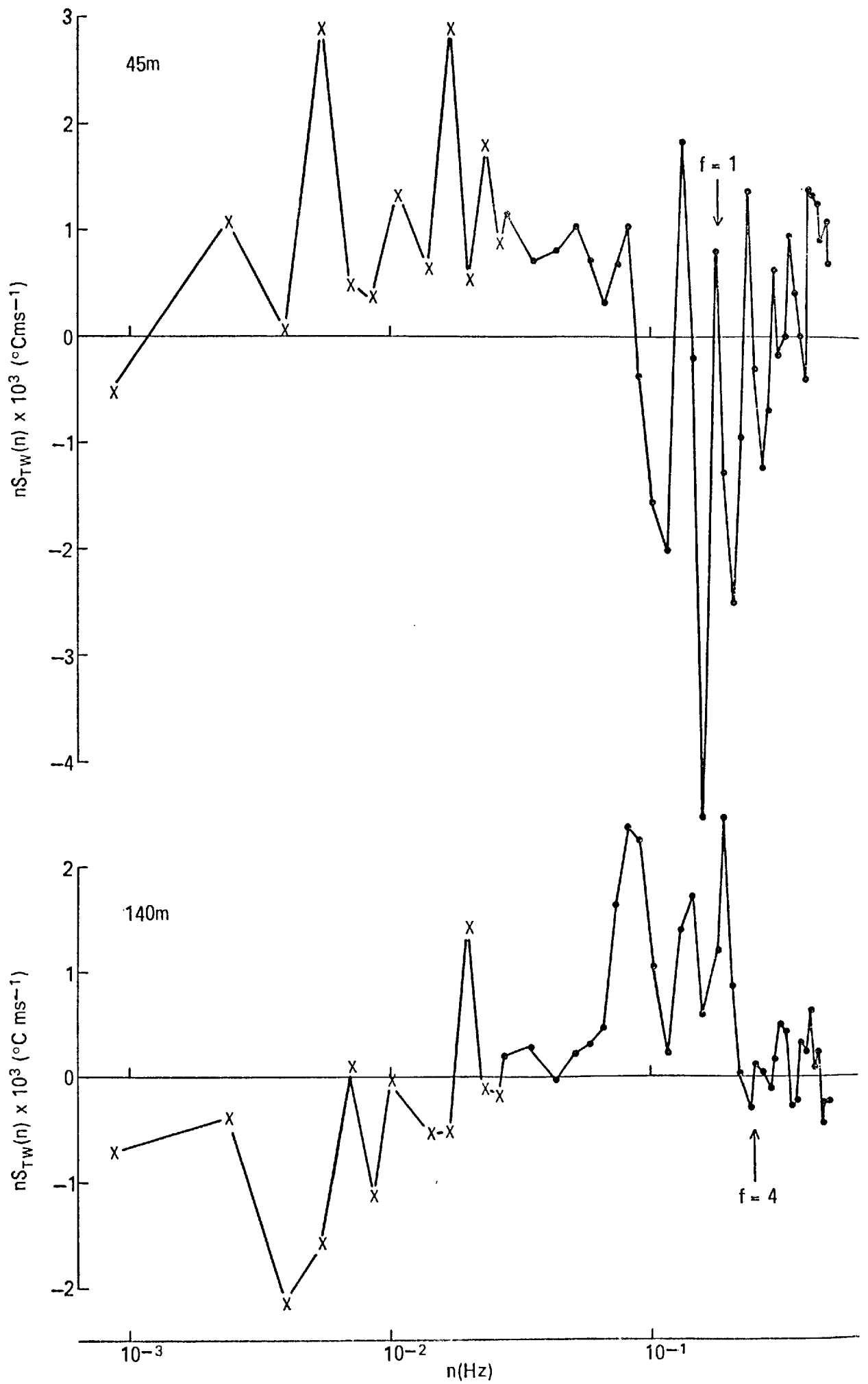


Figure 27 Composite T_w - cospectra, 1610 -1738 Z, 17.6.70 (JASIN Run 11)

than a fall-off of the rate of increase of characteristic turbulence scale with increasing height. On this occasion (JASIN run 11) apart from obviously identifiable noise the cospectral estimates were generally positive at 45m and negative at 140m, demonstrating a reversal of heat flux at the greater height (it will be remembered that figure 16 also suggested that for significant periods the heat flux could be downward at a height of about 200m while upwards at the lower level). It might be noted too that the lowest-frequency plotted estimate at 45m is negative also, suggesting that warm downdrafts were penetrating to this level also.

qw-cospectra (figure 28) alone showed almost insignificant influence from ship motion, as might be expected from lack of noise in the q-spectrum (figure 20).

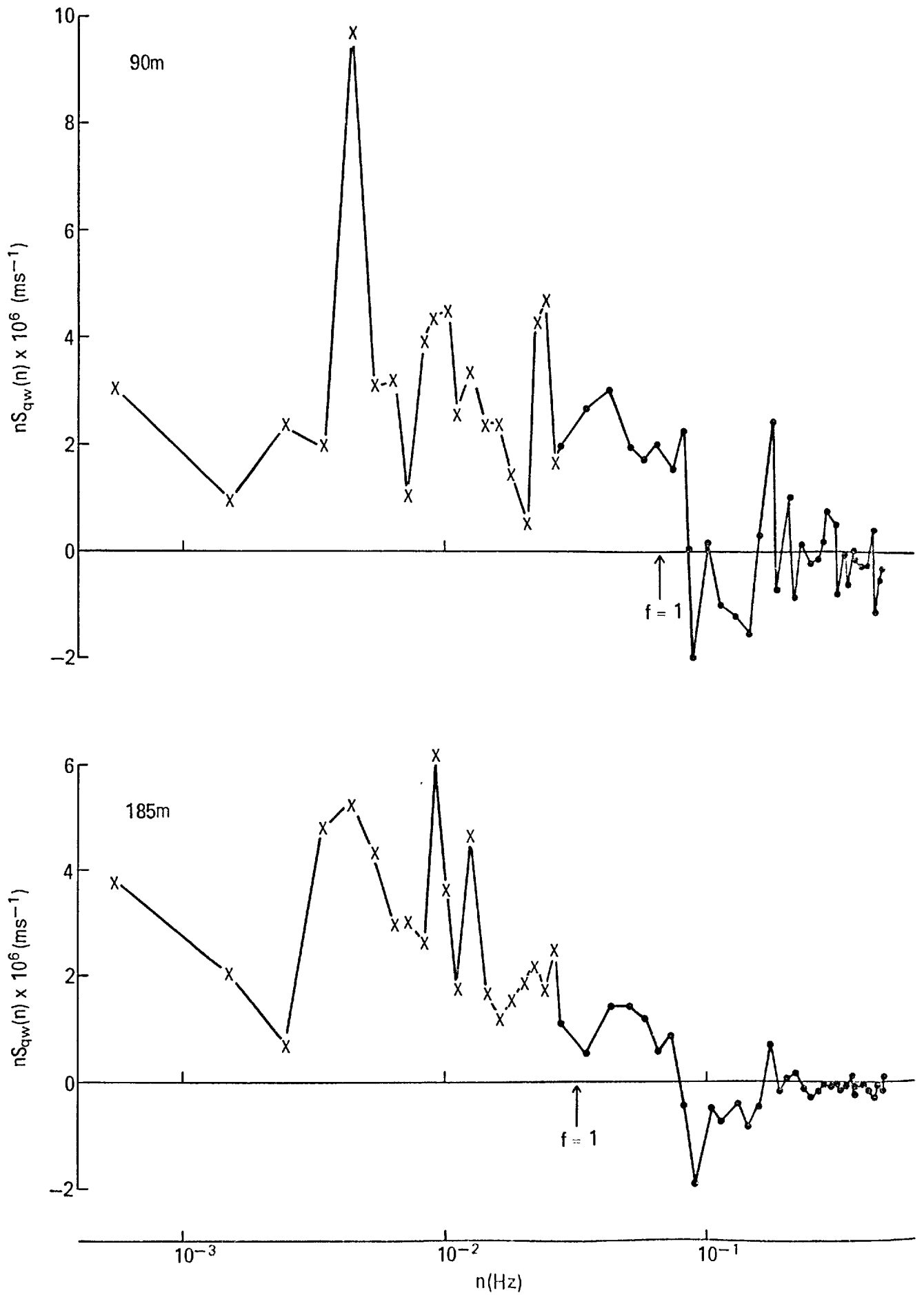


Figure 28 Composite qw - cospectra, 2143 -0005 Z, 17.6.70 (JASIN Run 13)

7.4

MOMENTUM FLUXES

These were calculated for all runs except the first, where measured winds were too light for satisfactory functioning of the anemometer, and run 16 where the ship motion was sufficiently violent to prevent satisfactory data being obtained: here the balloon-cable oscillations caused large lateral fluctuations of the vane supporting the sensors. The results are presented in Table 10 in the same way as those obtained from the ATEX data (Table 6). The Table also includes estimates for the surface flux using a value of 1.3×10^{-3} for the drag coefficient in neutral conditions with a correction for stability, and a final column giving the ratios of the measured fluxes to these surface estimates during the unsettled latter part of JASIN. The results for runs 18 and 19 were unreliable because of an anemometer fault which resulted in "clipping" of the highest wind speeds and although this probably produced comparatively small errors in vertical velocity, the horizontal speeds were significantly underestimated: this in turn resulted in spurious positive contributions to the fluxes. Discounting these runs the mean ratio was 0.74, with the turbulence measurements made at an average height of 100m. The mean 10m wind was about 7.3 ms^{-1} and the likely magnitude of the 900 m wind was therefore 8.3 ms^{-1} (Findlater et al 1966). If baroclinicity is ignored this latter wind corresponds closely to the geostrophic and then using equation (3) in conjunction with a geostrophic veer of 5 degrees (Mendenhall 1967) leads to a change of momentum flux between surface and 100 m of 0.010 Nm^{-2} , about 12% of the estimated surface flux. The measured fluxes can therefore be interpreted as implying a drag coefficient of 1.1×10^{-3} or alternatively as an underestimate by about 20% on average if the drag coefficient is assumed to be 1.3×10^{-3} . In the light of our present knowledge either interpretation is plausible but because of the positive

Table 10 Vertical Momentum fluxes measured during JASIN

Run Number	Instrument heights(m)	$\langle U_{10} \rangle$ (ms^{-1})	(a) Unaveraged data	FLUXES (Nm^{-2}) (b)8-sec average data	(c)Cospec smoothed to zero at $f=1$	(d)Estimated Surface flux	L (m)	Measured flux Surface flux
2	185	~ 2	0.036	0.010	0.010	-0.007	~ -15	
3	185	~ 2	0.016	0.013	0.010	-0.008	~ -15	
4	185	~ 2	0.046	0.008	0.006	-0.008	~ -15	
5	90	2.6	0.005	-0.010	-0.010	-0.009	+55	
	185		0.032	0.004	0.003			
6	90	5.3	-0.049	-0.015	-0.015	-0.042	-250	
	185		0.047	0.015	0.016			
7	90	2.3	0.025	0.007	0.006	-0.008	-130	
	185		0.054	0.023	0.025			
8	90	~ 1	-0.032	-0.0003	-0.003	-0.002	~ -30	
	185		-0.009	0.005	0.006			
9	95	8.1	0.072	-0.120	-0.150	-0.103	-80	1.47
10	95	8.1	0.167	-0.043	-0.053	-0.099	-130	0.54
11	45	7.7	-0.196	-0.080	-0.080	-0.090	-220	0.89
	140		0.181	-0.033	-0.059			0.66
12	45	6.9	-0.138	-0.034	-0.039	-0.073	-145	0.53
	140		0.114	-0.049	-0.056			0.77
13	90	5.7	-0.152	-0.039	-0.039	-0.052	-55	0.75
	185		0.101	-0.015	-0.023			0.44
14	95	7.7	0.007	-0.042	-0.045	-0.087	+490	0.52
15	95	8.5	-0.017	-0.050	-0.064	-0.106	+750	0.60
17	95	7.3	0.061	-0.084	-0.078	-0.081	-150	0.96
18	95	7.8	0.132	0.013	0.005	-0.095	-70	-0.05
19	95	8.1	0.102	-0.010	-0.013	-0.103	-70	0.13

contributions to the measured Uw -cospectra which are usually observed it appears more likely that the measured fluxes are slight underestimates. This is confirmed by figure 29 where Uw -cospectra from the latter part of JASIN (but excluding runs 16, 18 and 19), normalised by $\langle Uw \rangle$, have been plotted against reduced frequency: only cospectral estimates at natural frequencies less than about 0.1 have been plotted so that the main contribution due to sensor motion does not appear. Despite the considerable scatter (which at lower frequencies must be due in part at least to sampling variations) it is seen that on average contributions to the momentum fluxes at reduced frequencies between about 1 and 0.1 are smaller than expected (cf Figure 4).

In view of the considerable distortions of the U and w spectra and especially the cospectrum produced by sensor motion it is unremarkable that the fluxes appear to be erroneous, but the comparatively small size of the apparent average error is noteworthy: this suggests that even rather crude servostabilisation of the tethering point of the balloon cable (say reducing the amplitude of the motion by a factor of three) would suffice to allow momentum fluxes to be measured by the present technique with accuracies of a few percent.

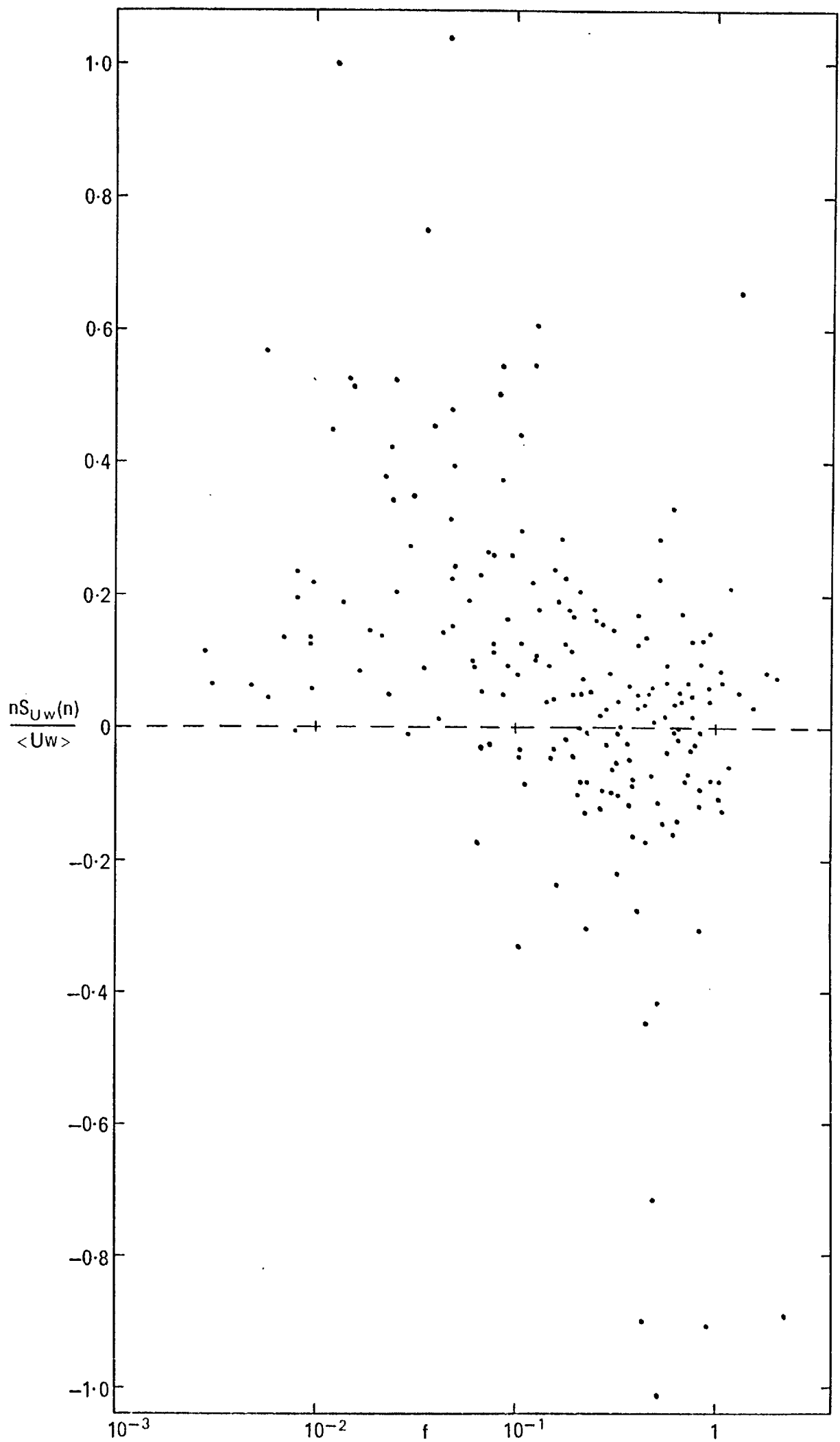


Figure 29 Normalised Uw - cospectra

7.5

HEAT FLUXES

These are given in Table 11. Here the preferred estimates are those obtained by smoothing the cospectra to zero at $f = 4$, but these differ by little from those calculated using the unaveraged data (column (a)), thus demonstrating the much-smaller influence of sensor motion than in the case of the momentum flux. The surface fluxes were obtained using eq (8a) but must be considered very approximate because of the difficulty of measuring the sea-air temperature difference accurately. In most cases the measured fluxes were less than the surface estimates, suggesting a flux convergence in the lower boundary layer, and this is confirmed by runs 11, 12 and 13 where measured fluxes either decreased considerably or changed sign in going from the lower to the upper instrument level (it will be recalled that the ATEX results supported a similar behaviour). In several cases the cospectra indicated that there were comparatively large contributions to the measured fluxes at low frequencies and this was thought to be of sufficient importance to be investigated in more detail using data from the latter part of JASIN: the relevant results are given in Table 12. Listed in column (b) are flux estimates which were obtained by applying an arbitrary low-frequency cut-off to the cospectra equivalent to shortening the data lengths by a factor of eight. The final column gives the corresponding (reduced) cut-off frequency. Typically this lay near 2×10^{-2} and it would be anticipated therefore that there would be significant contributions to the fluxes beyond this cut-off. On the other hand it is seen that on the occasions with the smallest upward surface fluxes (runs 11, 12 and 17), the flux at instrument level is increased by this high-pass filtering. As mentioned earlier it is likely that the upward heat transfer was due primarily to warm updraughts originating from near the surface and presumably associated chiefly with smaller-scale turbulence because of this. In contrast

Table 11 Vertical Heat fluxes during JASIN

Run Number	Instrument heights(m)	$\langle U_{10} \rangle$ (ms^{-1})	FLUXES (W m^{-2})				Measured flux Surface flux
			(a) Unaveraged data	(b) 8-sec averaged data	(c) Cospec smoothed to zero at $f = 4$	(d) Estimated surface flux	
2	185	~ 2	-2.2	-2.1	-2.3	1	
3	185	~ 2	0.2	0.02	0.1	3	
4	185	~ 2	-0.8	-0.9	-0.9	3	
5	90	2.6	-0.9	-1.1	-1.3	2	
	185		-1.5	-1.1	-1.2		
6	90	5.3	-3.8	-3.8	-3.8	0	
	185		-3.3	-3.6	-3.8		
7	90	2.3	-2.7	-2.7	-2.9	-0.5	
	185		-3.9	-3.9	-4.1		
8	90	~ 1	0.3	-0.1	-0.2	-0.2	
	185		0.2	-0.03	-0.05		
9	95	8.1	5.4	3.0	3.7	20	0.19
10	95	8.1	8.5	5.3	7.7	10	0.77
11	45	7.7	-0.8	-0.6	1.2	5	
	140		-6.3	-7.7	-7.6		
12	45	6.9	5.2	5.1	7.5	5	
	140		-8.4	-9.0	-8.9		
13	90	5.7	5.8	5.5	6.8	10	0.68
	185		3.4	2.0	1.9		
14	95	7.7	-3.0	-3.5	-3.6	-7	
15	95	8.5	4.1	2.8	3.9	-7	
17	95	7.3	-0.2	-3.5	-0.7	4	
18	95	7.8	28.1	24.1	27.4	20	1.37
19	95	8.1	26.1	21.9	25.0	23	1.09

Table 12 Low-frequency contributions to heat fluxes
obtained for the latter part of JASIN

Run Number	Instrument Heights (m)	FLUXES (Wm^{-2})			
		(a) Cospec smoothed to zero at $f=4$	(b) As (a) but with low-frequency cut-off also	(c) Estimated surface flux	f at low- frequency cut-off
9	95	3.7	4.5	20	0.030
10	95	7.7	6.0	10	0.046
11	45	1.2	5.2	5	0.009
	140	-7.6	-1.9		0.027
12	45	7.5	8.3	5	0.010
	140	-8.9	4.8		0.030
13	90	6.8	5.4	10	0.015
	185	1.9	1.5		0.032
14	95	-3.6	-2.8	-7	0.012
15	95	3.9	2.3	-7	0.028
17	95	-0.7	12.2	4	0.019
18	95	27.4	26.5	20	0.024
19	95	25.0	21.5	23	0.017

the major downdraughts were likely to occur in conjunction with larger-scale features such as cumulus clouds, and would tend to transfer heat downwards because of the sub-adiabatic lapse rate at the higher levels: the greater the height above the surface the more chance there would be of these downdraughts penetrating to instrument level, thus resulting in a decrease of upward flux with height. In cases with large upward surface fluxes the contributions to fluxes at greater heights and low frequencies from warm updraughts while relatively small may still be large enough to override the downward transfer associated with the larger-scale features, but where upward surface fluxes are small the net contribution from low frequencies is more likely to be negative, thus offering an explanation for the differences between columns (a) and (b) in the Table for most of the runs.

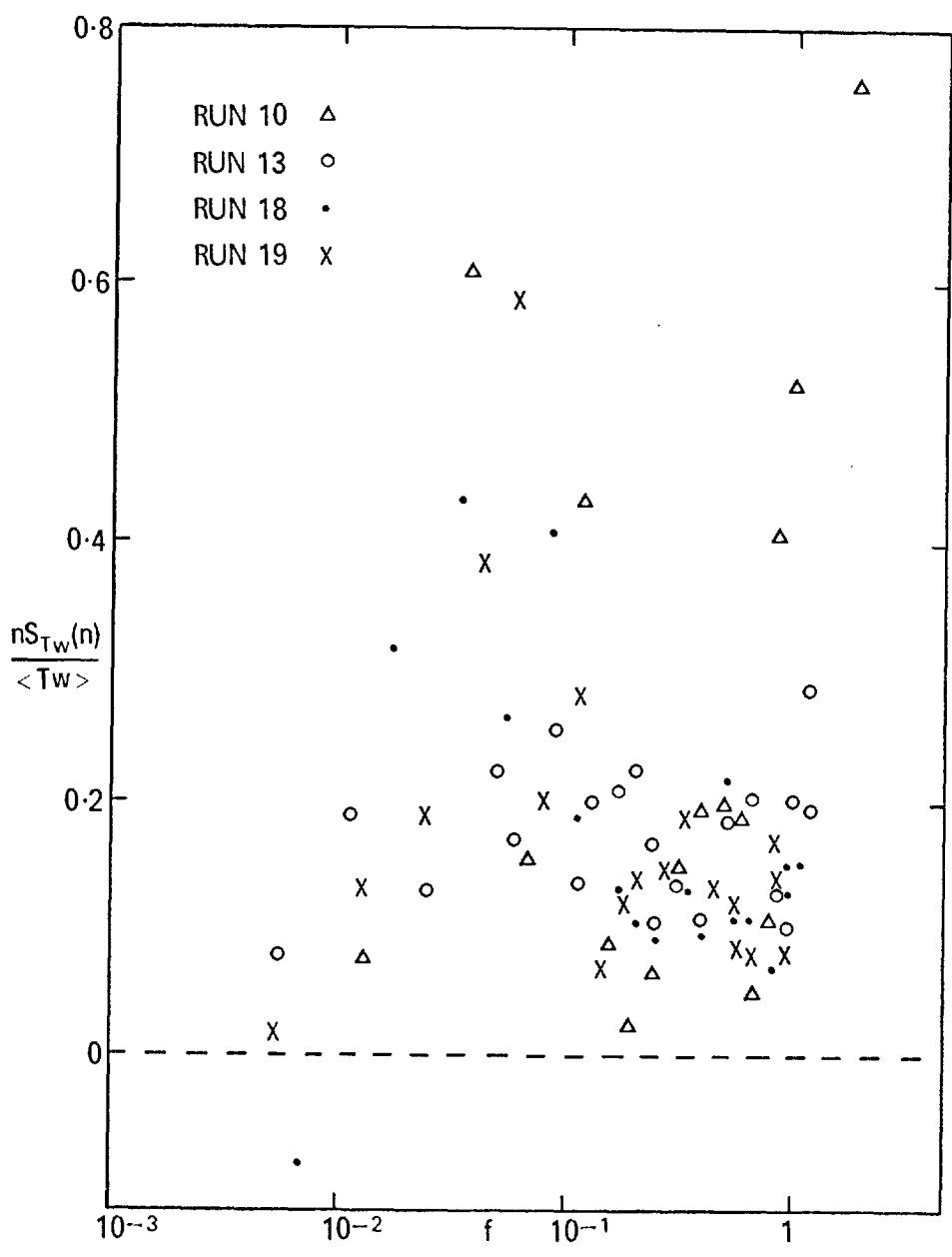
There is the possibility that instrumental faults such as electronic drift could have caused spurious correlations at low frequencies which might have contributed to the apparent change of flux with height. It is certainly true that changes in the apparent mean inclination of the airflow to the horizontal, due presumably to ageing of the wires of the inclination sensor were detected when data from consecutive runs were compared, the typical values being around 5×10^{-3} radians. If these changes were correlated with apparent temperature changes of the right magnitude then spurious low-frequency contributions to the heat flux of the right size would have resulted. Run 12 at 140m gave a large negative heat flux at low frequencies (about -14 Wm^{-2} , equivalent to $\langle T_w \rangle \sim -0.012 \text{ degrees Cms}^{-1}$). On this occasion the mean inclination changed by 1.5×10^{-2} radians from the previous period (an apparent change of 0.12 ms^{-1} in $\langle w \rangle$) and it appears therefore that spurious low-frequency temperature fluctuations of about 0.1 degrees C if suitably phased with the

apparent change in w would have sufficed to produce the flux. However such large temperature fluctuations were not anticipated as a result of instrumental drift and in any case would not have correlated perfectly with spurious inclination fluctuations. Also the trend corrections applied in the data analysis would have reduced further the correlation.

Sampling errors might have resulted in unexpected contributions to the fluxes, but the sign of the contributions would have been random, in this case certainly not observed. It appears therefore that very considerable vertical variations of vertical heat flux will be found in slightly unstable conditions in the lower boundary layer over the sea and it will be demonstrated later that this has important implications in the parameterization of the marine boundary layer.

Normalised cospectra for the four runs with the largest upward fluxes and similar instrumental heights are shown in Figure 30. The scatter at low frequencies is considerable, as in Figure 29, 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 recognised 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 (1970a) 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.

The measured fluxes (obtained after smoothing the cospectra to zero at $f = 4$) were compared with the estimated surface values on five occasions when the sea-air temperature difference was around 1 degree C or greater (see Table 11): the average ratio was about 0.8 (alternatively if surface fluxes were assumed to be equal to those measured aloft on these occasions then the resulting average value for C_H in eq (8a) was around 1.1×10^{-3}).



Because of the decrease of upward flux with height these results offer satisfactory support for the validity of the present technique for measuring vertical heat flux.

7.6

WATER-VAPOUR FLUXES

The fluctuations of specific humidity were obtained from the measured air temperature and wet-bulb depression as described in § 5 but neither of these parameters was known very accurately, partly because the shipboard calibration system used in JASIN was designed primarily to check output sensitivities of the sensor circuits rather than absolute calibrations but also because uninsulated wire was used in the wet-bulb elements (which were calibrated in the laboratory when dry) and due to the shunting effect the measured wet-bulb temperature was an underestimate by about 1 degree C. More reliable estimates for mean specific humidity were obtained using the measured fluxes in a relation applicable to the constant-flux layer

$$q_z = q_{10} - E (\log(z/10) - \psi_E(z/L) + \psi_E(10/L)) / \rho k u_* \quad (32)$$

: ψ was obtained from the data of Dyer and Hicks, and Webb. These revised values for q were used in conjunction with temperatures estimated by making the reasonable assumption of an adiabatic lapse rate between 10m (the observational height on the ship) and instrument level to adjust the measured water-vapour fluxes. (The estimates for temperature and specific humidity yielded corrected values for mean wet-bulb depression. With eq (17-19) it was then possible to calculate the change (q') in specific humidity for a small change in T_w , using both measured and corrected mean values for T_w , and thus to arrive at a new value for E . The procedure was repeated with the revised values of E inserted in eq (32). Further iterations produced negligible changes in E). The procedure was applied only to data obtained at heights below 100m, but excluding run 5 where the correction was very uncertain because of the significantly stable stratification. The average adjustment was 3%.

Table 13 lists the various flux estimates and includes values for C_E obtained using the measured fluxes (assumed best estimated by those given

Table 13 Vertical moisture fluxes during JASIN

Run No.	Instrument heights(m)	$\langle U_{10} \rangle$ (ms^{-1})	FLUXES ($\text{mg m}^{-2} \text{s}^{-1}$)			(d) Estimated surface flux	C_E $\times 10^3$	C_E (N) $\times 10^3$
			(a) Unaveraged data	(b) 8-sec averaged data	(c) Cospec smoothed to zero at $f=1$			
2	185	~ 2	4.0	3.7	3.8	9		
3	185	~ 2	5.5	5.2	4.8	9		
4	185	~ 2	6.7	6.3	6.2	9		
5	90	2.6	9.0	8.1	8.5	5		
	185		1.2	0.5	0.2			
6	90	5.3	23.2	20.7	22.1	14	2.05	2.00
	185		14.7	13.4	13.4			
7	90	2.3	2.6	2.1	2.2	4	0.81	0.77
	185		3.8	2.6	2.2			
8	90	~ 1	0.1	0	0.1	2		
	185		0.4	-0.1	0			
9	95	8.1	44.3	44.0	46.9	34	1.97	1.83
10	95	8.1	34.5	31.8	30.6	28	1.51	1.43
11	45	7.7	25.4	18.9	20.1	19	1.45	1.41
	140		5.7	5.2	5.7			
12	45	6.9	32.2	29.8	32.8	21	2.12	2.03
	140		38.5	38.2	38.0			
13	90	5.7	19.8	18.6	19.4	22	1.26	1.14
	185		19.1	19.1	18.8			
14	95	7.7	16.1	12.6	15.0	21	0.91	0.94
15	95	8.5	55.0	49.9	52.7	21	3.29	3.35
17	95	7.3	82.5	81.8	82.8	41	2.82	2.70
18	95	7.8	45.3	40.8	44.9	44	1.43	1.32
19	95	8.1	45.5	41.7	43.7	37	1.64	1.51

in Column (c)) in eq (9a): the final column gives $C_E(N)$, the transfer coefficient corrected to neutral stability using Figure 1. There was generally close agreement between the three different estimates for the measured fluxes, as would be anticipated from the comparatively small apparent influence of sensor motion on the qw-cospectra.

The fluxes showed comparatively small variations with height, not only when there was a significant sea-air temperature difference (run 13) but also on occasions with much of the buoyancy flux due to the net upward transfer of water-vapour (runs 7, 12 and to a lesser degree 6 also). For this reason it is thought that the surface flux can be closely approximated by that measured at a height of even around 100m in convective conditions. The mean value for C_E was $(1.78 \pm 0.68) \times 10^{-3}$: this may be compared with 1.3×10^{-3} found by Robinson and Chamberlain: the difference is significant at the 4% level. However the results are strongly weighted by the very high values from runs 15 and 17. In the case of the former the heat flux (Table 11) was directed upwards at instrument level in spite of an estimated negative surface flux and this strongly suggests that the observations of sea temperature from the ship were not representative on this occasion, the true values being higher than those used to calculate Δq so that C_E was overestimated. There seems to be no such explanation for run 17. (It may be noted that if these two runs are discarded the resulting average for C_E is $(1.51 \pm 0.42) \times 10^{-3}$ which differs from Robinson's and Chamberlain's estimate at the 15% level only).

The observed relation between vapour flux and the product of wind speed and sea-air humidity difference is shown graphically in Figure 31. The regression lines were fitted by least-squares and gave values for C_E of 1.94×10^{-3} using all the data points or 1.52×10^{-3} after discarding runs 15 and 17 (observations from these are represented by the circled points in the diagram).

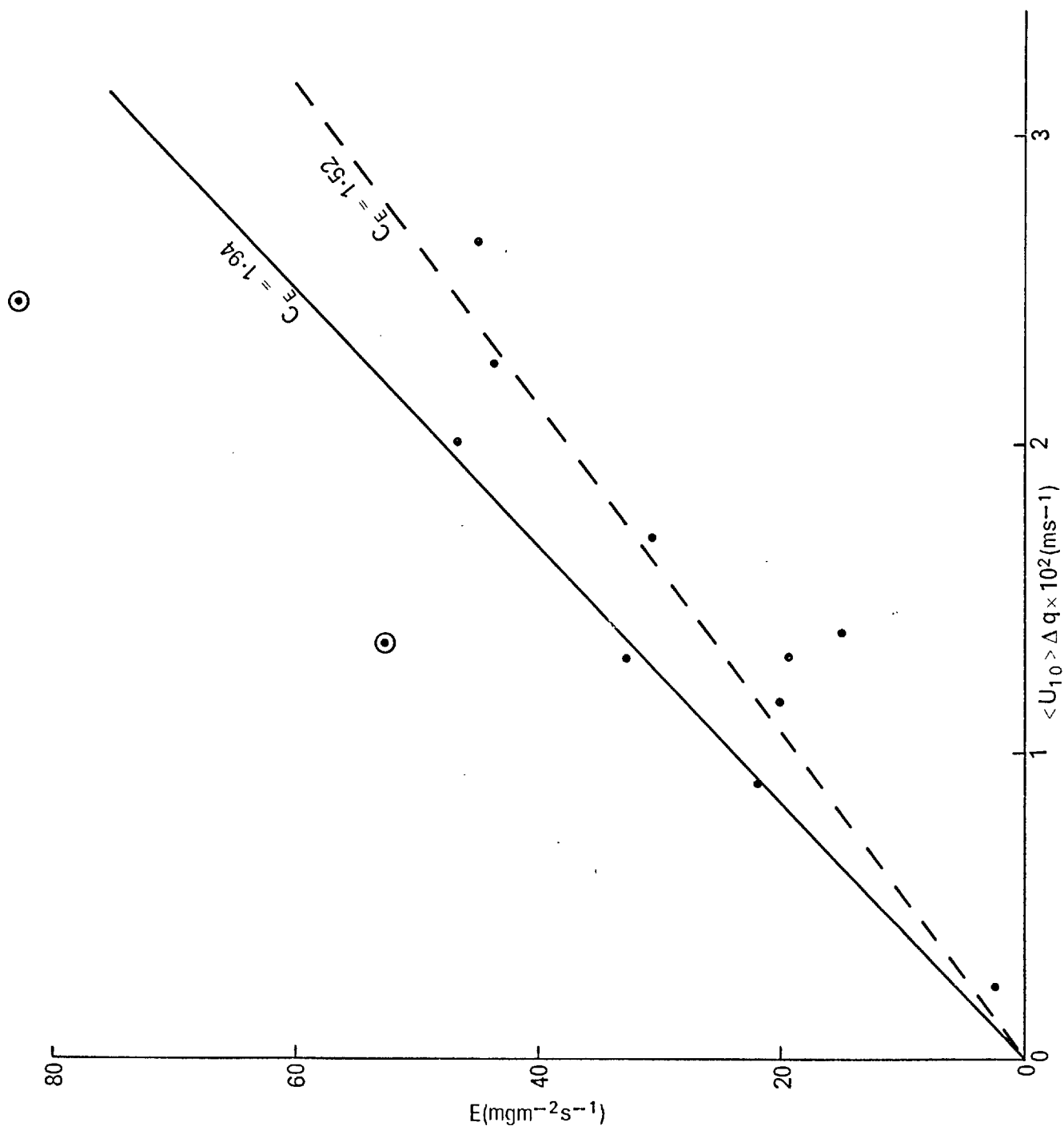


Figure 31 The relation between measured moisture fluxes and $\langle U_{10} \rangle \Delta q$

Corresponding cospectra (normalised by $\langle qw \rangle$) are shown in Figure 32 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 (1970b).

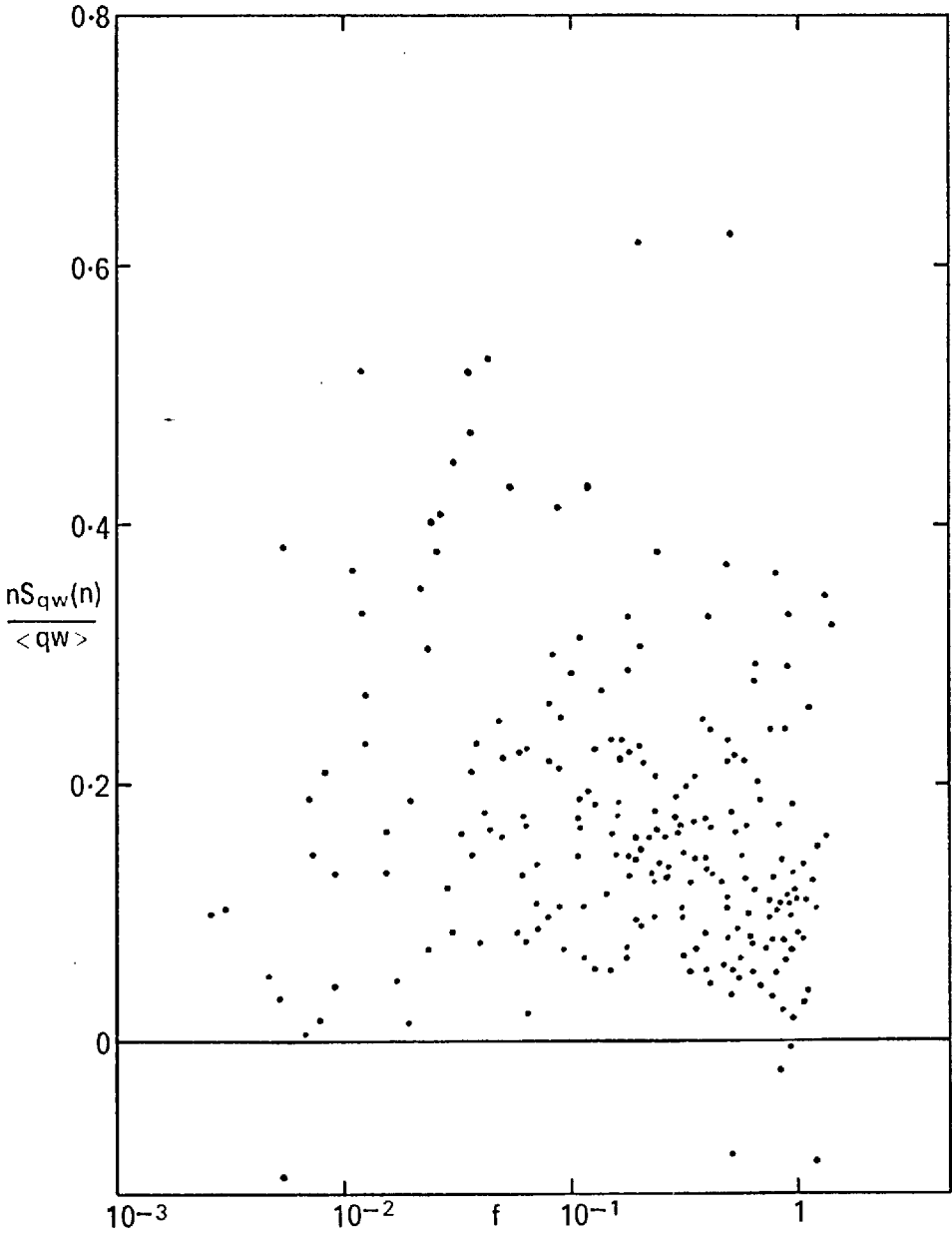


Figure 32 Normalised qw – cospectra

7.7 STANDARD DEVIATIONS OF TURBULENT FLUCTUATIONS

Studies such as those of McBean (1971) have shown that close to the surface at least the dimensionless standard deviations of wind, temperature and humidity (i.e. normalised by u_* , T_* and q_* respectively) have reasonably well-defined values which appear to vary comparatively slowly with changes in the stability parameter z/L . McBean gives the following value:

	$z/L \sim 0$	$z/L \sim -0.5$
σ_u / u_*	2.2	2.6
σ_w / u_*	1.4	1.7
σ_T / T_*	-1.6	-1.1
σ_q / q_*	-1.5	-1.2

It is difficult to compare the wind statistics obtained from the JASIN data with these values because of the substantial contributions to the variances due to sensor motion. Thus estimates for σ_u (and σ_w) derived from 4- and 8- second averaged data obtained in moderate winds usually differed considerably (Table 14) chiefly because of larger noise contributions to the former. From figure 3, and noting that the average cut-off of the 8-second running-average filter was around $f = 0.6$ for the runs listed in the Table, the values for $(\sigma_u)_{8\text{-sec av.}}$ were underestimates by about 5% (it is assumed here that this averaging time was sufficiently large to attenuate almost completely the noise contributions). The mean value for $(\sigma_u)_{8\text{-sec}}/u_*$ was 1.9, or with a correction for filter attenuation ~ 2.0 : here u_* was obtained from the estimated vertical momentum flux at the surface. The average values for z/L was -0.9 if L was calculated using the estimated surface fluxes, with a similar value if measured fluxes were used in the derivation. McBean's results suggest in this case that σ_u/u_* is around 3.0 near the surface so there appeared to be a breakdown of similarity in the case of the JASIN results (the ratio is increased to 2.3 if σ_u was

Table 14 Standard deviations of wind fluctuations, JASIN

Run No.	Height (m)	u_* (ms^{-1})	$\sigma_{u(\text{ms}^{-1})}$		$\sigma_{w(\text{ms}^{-1})}$		σ_u/u_* (8-sec av.data)	σ_w/u_*
			(a) From 4-sec av. data	(b) 8-sec av. data	(a) 4-sec av. data	(b) 8-sec av. data		
6	90	0.19	0.37	0.27	0.39	0.29	1.5	1.5
	185		0.30	0.29	0.22	0.17	1.6	0.9
9	95	0.29	0.64	0.59	0.69	0.52	2.0	1.8
10	95	0.28	0.53	0.47	0.66	0.45	1.7	1.6
11	45	0.27	0.51	0.43	0.71	0.39	1.6	1.4
	140		0.49	0.43	0.60	0.38	1.6	1.4
12	45	0.24	0.48	0.41	0.67	0.37	1.7	1.5
	140		0.51	0.46	0.61	0.42	1.9	1.7
13	90	0.21	0.49	0.40	0.59	0.33	1.9	1.6
	185		0.39	0.35	0.56	0.39	1.7	1.9
14	95	0.27	0.53	0.50	0.39	0.28	1.9	1.1
15	95	0.29	0.39	0.35	0.45	0.36	1.2	1.2
17	95	0.26	0.83	0.80	0.76	0.56	3.1	2.2
18	95	0.28	0.80	0.76	0.71	0.55	2.7	2.0
19	95	0.29	0.89	0.86	0.75	0.62	3.0	2.1
Means							1.9	1.6

normalised by the measured value for u_* , still significantly different from McBean's value). However Busch and Panofsky (1968) have commented on the considerable low-frequency variability which is observed when spectra obtained at different sites and heights (chiefly over land) are plotted against reduced frequency, and so the difference between McBean's and the present results should be interpreted perhaps as more evidence for this variability rather than demonstrating any special characteristics of the JASIN data.

The corresponding average value for σ_w/u_* was 1.6, close to that expected but the agreement here is partly fortuitous since figure 3 demonstrates that the 3-second filter reduces σ_w by about 15%: in the present case the filter attenuation is presumably compensated by the noise peak of the w-spectrum which is much less attenuated by the filter than the narrower U-spectrum noise peak.

A large scatter in the normalised value for σ_T was expected because of the comparatively small thickness of the constant-flux layer for the heat flux in many of the runs but even so there were **some unexpected results**. Thus in run 17 T_* was small, whether based on measured fluxes or those estimated at the surface, but σ_T was one of the largest found in all the runs (Table 15) implying that a substantial contribution to the temperature variance can be "inactive", that is, not correlated with significant vertical motions, and therefore not contributing to the heat flux. The four runs with significant sea-air temperature differences and apparently small variations of heat flux with height yielded a mean value around -1.4 for σ_T/T_* , with z/L about -1, somewhat smaller than anticipated from McBean's results and supporting therefore a significant contribution to σ_T from "inactive" fluctuations.

Table 15 Standard deviations of temperature
and humidity fluctuations, JASIN

Run No.	Height (m)	$\sigma_T \times 10^2 (^{\circ}\text{C})$		σ_{T/T_*} (4-sec av)	$\sigma_q \times 10^4$		σ_{q/q_*} (4-sec av)
		(a) 4-sec av. data	(b) 8-sec av. data		(a) 4-sec av. data	(b) 8-sec av. data	
6	90	3.86	3.70		0.99	0.94	-1.02
	185	3.35	3.27		1.01	0.98	
9	95	15.5	15.3		1.74	1.72	-1.30
10	95	4.44	4.28	-1.9	1.57	1.55	-1.79
11	45	8.21	8.10		1.54	1.51	-2.57
	140	5.49	5.35		1.36	1.33	
12	45	6.12	5.96		2.74	2.70	-2.51
	140	14.1	14.1		2.21	2.19	
13	90	3.43	3.23	-1.2	1.08	1.05	-1.40
	185	2.91	2.78		1.03	1.01	
14	95	6.86	6.76		1.47	1.44	-3.20
15	95	4.71	4.54		2.32	2.27	-1.59
17	95	13.2	13.1		3.05	3.04	-1.16
18	95	12.9	12.7	-1.6	1.57	1.55	-1.20
19	95	6.73	6.49	-0.9	1.82	1.80	-1.48

The normalised standard deviations of humidity fluctuations (Table 15) showed much less scatter than those for σ_T . The normalisation was done here using the vapour flux at the lowest height and the estimated surface value of u_* to derive q_* : the average value of σ_q/q_* was -1.7 after discarding data from runs 11 and 12 where the winding of the temperature element was occasionally bridged and therefore shunted by rain drops, thereby producing spurious temperature and humidity fluctuations. Except for the very high (and inexplicable) value from run 14 the scatter is fairly small but the mean value is significantly less than that found by McBean and Miyake et al (1970b) however and this implies at least some proportion of σ_q is inactive. The values found for runs 15 and 17 (-1.6 and -1.2) were fairly close to the mean and this gives some support to the validity of the measured moisture fluxes for these occasions (where it will be recalled the estimated surface fluxes were less than half the measured values).

7.8

TURBULENT ENERGY BALANCE

In the inertial subrange the following relations may be used to calculate rates of dissipation of turbulent fluctuations.

$$S_{uu}(\kappa) = C_1 \epsilon^{2/3} \kappa^{-5/3} \quad (33)$$

$$S_{vv,ww}(\kappa) = C_2 \epsilon^{2/3} \kappa^{-5/3} \quad (34)$$

$$S_{\tau\tau}(\kappa) = C_3 \chi_\tau \epsilon^{-1/3} \kappa^{-5/3} \quad (35)$$

$$S_{qq}(\kappa) = C_4 \chi_q \epsilon^{-1/3} \kappa^{-5/3} \quad (36)$$

Here $C_1 = 0.75C_2 \simeq 0.47$ (Panofsky and Pasquill 1963) and $C_3 \simeq 0.7$ (Panofsky 1969): κ is in radians per metre. C_4 probably has a value similar to that for C_3 . Using the conventional Taylor transformation ($\kappa = 2\pi n / \langle U \rangle$) eq (33) may be rewritten

$$\kappa S_{uu}(\kappa) = n S_{uu}(n) = 0.47 \epsilon^{2/3} (2\pi n / \langle U \rangle)^{-2/3}$$

leading to

$$\epsilon(u) = 19.5 (S_{uu}(n))^{3/2} n^{5/2} / \langle U \rangle \quad (37)$$

and alternatively (from eq 34)

$$\epsilon(w) = 12.6 (S_{ww}(n))^{3/2} n^{5/2} / \langle U \rangle \quad (38)$$

In practice estimates of dissipation from horizontal and vertical velocity spectra are unlikely to agree even if data are of very high quality because of the lack of isotropy often observed in the $-5/3$ region for the atmosphere over the sea (Weiler and Burling (1967), Miyake et al (1970 a,b)), and also because the horizontal velocity rather than the component of this velocity along the mean direction is usually used to calculate the spectral density appearing in eq (37). Dissipation rates for the temperature and humidity variances are given by

$$\begin{aligned} \chi_{\tau,q} &= ((2\pi)^{2/3} / 0.7) S_{\tau\tau,qq}(n) \epsilon^{1/3} n^{5/3} \langle U \rangle^{-2/3} \\ &= 4.9 S_{\tau\tau,qq}(n) \epsilon^{1/3} n^{5/3} \langle U \rangle^{-2/3} \end{aligned} \quad (39)$$

The turbulent energy equations are, for horizontally homogeneous and stationary turbulence,

$$-\langle uw \rangle \frac{\partial \langle u \rangle}{\partial z} - \langle vw \rangle \frac{\partial \langle v \rangle}{\partial z} - g \langle ew \rangle / \langle e \rangle - \epsilon - \frac{\partial}{\partial z} \langle E'w \rangle - \frac{\partial}{\partial z} \langle pw \rangle / \langle e \rangle = 0 \quad (40)$$

$$-\langle Tw \rangle \frac{\partial \langle \theta \rangle}{\partial z} - \chi_T - \frac{\partial}{\partial z} \langle T'^2 w \rangle / 2 = 0 \quad (41)$$

$$-\langle qw \rangle \frac{\partial \langle q \rangle}{\partial z} - \chi_q - \frac{\partial}{\partial z} \langle q'^2 w \rangle / 2 = 0 \quad (42)$$

Here $E' = (\overline{u'^2} + \overline{v'^2} + \overline{w'^2})/2$: it has been assumed that $\langle w \rangle = 0$. In eq (40) it is conventional to neglect the pressure transport term though this may not be always justifiable (Wyngaard and Coté 1971), and as a first approximation the first term may be replaced by $-\langle Uw \rangle \partial \langle U \rangle / \partial z$: also near the surface over the sea the second term can probably be neglected by comparison with the first. Taking into account the contribution of turbulent fluxes of both heat and moisture to the buoyant production of turbulent kinetic energy eq (40) can then be written

$$-\langle Uw \rangle \frac{\partial \langle U \rangle}{\partial z} + g(\langle Tw \rangle + 0.61 \langle T \rangle \langle qw \rangle) / \langle T \rangle - \epsilon - \frac{\partial}{\partial z} \langle E'w \rangle = 0 \quad (43)$$

Additionally $\langle U'^2 \rangle \approx \langle u'^2 \rangle \approx \langle v'^2 \rangle$ (McBean 1971) and so

$E' \approx U'^2 + w'^2/2$. In the constant-flux layer the following substitutions may also be made:

$$-\langle Uw \rangle \frac{\partial \langle U \rangle}{\partial z} \approx u_*^3 \phi_m(z/L) / \kappa z \quad (44)$$

$$-\langle Tw \rangle \frac{\partial \langle \theta \rangle}{\partial z} = \langle Tw \rangle^2 \phi_h(z/L) / \kappa u_* z \quad (45)$$

$$-\langle qw \rangle \frac{\partial \langle q \rangle}{\partial z} = \langle qw \rangle^2 \phi_e(z/L) / \kappa u_* z \quad (46)$$

When these are inserted in eq (41) - (43), it is seen that provided simultaneous measurements have been made at more than one height it is then possible to investigate turbulent energy balances between measurement levels using the JASIN data.

Run 11, 12 and 13 were occasions on which data were obtained at two heights and were particularly suitable for the present purpose because they were convective situations with, in general, significant vertical fluxes. The spectra which had been calculated already for these runs were extended to higher frequencies by adding spectral estimates deduced from the data sampled at five times per second ($\S 5$). It had been hoped originally that the spectral regions at frequencies significantly higher than the main contribution due to sensor motion would show insignificant noise from the same source but this was unduly optimistic. Figure 33 for example demonstrates that the vertical velocity spectrum had a periodic structure at frequencies well above those of the main noise peak. It is perhaps interesting to note that the peaks of the various periodicities on this occasion occurred at the approximate frequencies 0.02, 0.04, 0.08, 0.16, 0.33, ~ 0.6 and 1.25Hz, and also at 0.06, 0.12 and ~ 0.25 Hz. Thus there appeared to be two independent series of harmonically-related distortions of the spectrum. In these circumstances it was clearly not feasible to attempt to obtain estimates for ϵ from spectral estimates on the high-frequency side of the main peak resulting from sensor motion (the U-spectrum also had similar characteristics). On the other hand there was a possibility of using spectral estimates at frequencies lower than the major noise peak provided the $-5/3$ region extended far enough. The measurements of Miyake et al (1970b) suggested this limit for w was around $f = 1$, though perhaps a little lower for U . This corresponds to about 0.17Hz for runs 11 and 12 at 45m and 0.07Hz at 90m in run 13. Since the former lay close to the centre of the main noise peak it was still necessary to extrapolate spectral estimates from lower frequencies in order to obtain the estimate at the lower end of the $-5/3$ region but at greater heights the estimate could be extracted with less ambiguity: in either case before this was done the identifiable harmonically-related peaks between about 0.02 and 0.08Hz were smoothed out. Table 16 lists values for

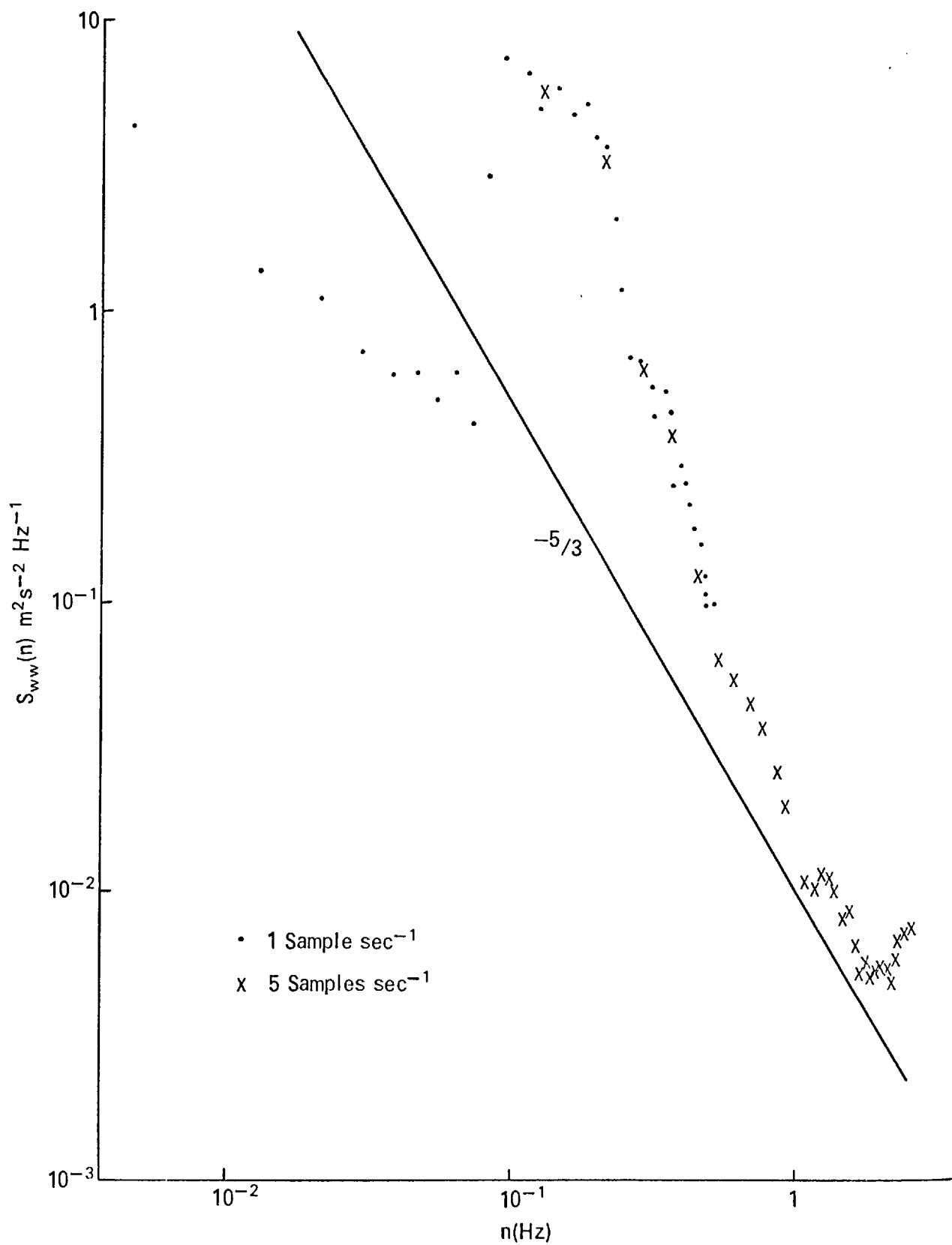


Figure 33 Composite W – spectrum, 140m, 1610 –1738 Z, 17.6.70 (JASIN Run 11)

calculated from both U and w spectra in this way, their mean (assumed to be the best estimate), and also calculated values for the various terms of eq (43). Here u_* was the estimated surface value obtained using eq (3):

ϕ_m was deduced from the results of Dyer and Hicks (1970) using the values for L given in Table 3. No attempt was made to estimate mechanical production of turbulent kinetic energy at the upper measurement levels in this way since these heights were above the constant-flux layer and in a region where the variation of wind speed with height was likely to be affected to some extent by Coriolis forces: in these conditions eq (44) - (46) are invalid. For these reasons even the value given at 90m (the lower level)

Table 16 Turbulent kinetic energy balance

Run No	Ht (m)	$\epsilon(U)$	$\epsilon(w)$	$\langle \epsilon \rangle$ (in $m^2 s^{-3}$: all estimates are $\times 10^{-4}$)	$u_*^3 \phi_m / \kappa z$	$-g \langle \epsilon w \rangle / \langle \epsilon \rangle$	$\langle w(u'^2 + w'^2/2) \rangle$ $m^3 s^{-3}$ ($\times 10^{-1}$)
11	45	4.6	3.9	4.3	7.4	+1.3	+1.0(1.5)
	140	5.5	3.7	4.6	-	-1.8	+2.2(2.7)
12	45	3.4	5.1	4.3	4.7	+3.7	+2.5(2.9)
	140	5.3	3.0	4.1	-	-0.6	+3.6(5.0)
13	90	4.5	3.6	4.1	(1.0)	+2.8	-0.5(-0.1)
	185	3.7	2.6	3.1	-	+1.4	+1.7(2.2)

in run 13 is very doubtful. The buoyant production of turbulent kinetic energy was obtained from the measured **vertical fluxes of heat and moisture**. The vertical flux of turbulent kinetic energy was derived using 8-second averaged data but was found to be rather sensitive to averaging time, presumably because of the very large influence of ship motion on this term. On the other hand increasing the averaging time to 32 seconds yielded values (shown in parentheses) which, while greater than those from 8-second averaged data, showed differences between the two levels in the mean only about 30% larger

than given by the shorter averaging time: thus the divergence of the vertical flux of kinetic energy was less sensitive to averaging time (the values from 8-second averaged data are those which will be used).

The main features demonstrated by the table are (i) only small variations of dissipation with height, (ii) at the lower height (45m) a substantial excess of production over dissipation, and (iii) assuming that mechanical production is roughly proportional to height, dissipation greater than production at the levels 90m and 185m. A similar structure was found over land by Lenschow (1970) though in this case production exceeded dissipation up to heights around 500m because of a much less rapid variation of heat flux with height. The results generally do not support a close relation between dissipation and total production as suggested by Busch and Panofsky (1968), or a simple relation between dissipation and mechanical production (Lumley and Panofsky 1964). The prospect of being able to estimate vertical momentum flux merely from measurements of high-frequency turbulent fluctuations of wind and the vertical gradient of wind speed appears therefore to be very remote.

Table 17 gives average values of dissipation and mechanical and buoyant production between the two levels for each run. Here mechanical production was calculated directly from the speed difference between the two measurement heights and the average measured value for u_*^2 . Dissipation

Table 17 Mean Kinetic energy balance

Run No.	$\epsilon \times 10^4$ ($m^2 s^{-3}$)	$-\langle U w \rangle \frac{\partial \langle U \rangle}{\partial z} \times 10^4$ ($m^2 s^{-3}$)	$-g \langle e w \rangle / \langle e \rangle$ $\times 10^4$ ($m^2 s^{-3}$)	$\frac{\partial}{\partial z} \langle E' w \rangle \times 10^4$ ($m^2 s^{-3}$)
11	4.5	2.1	-0.3	12.6 (11.5)
12	4.2	1.8	1.5	11.6 (5.3)
13	3.6	0.1	2.1	23.1 (14.7)

is seen to exceed total production in all three cases but the calculated divergence of vertical flux of kinetic energy is in each case of the wrong sign and much too large to affect a balance. The divergence term may have been incorrectly represented by assuming that $\langle wv'^2 \rangle \sim \langle wu'^2 \rangle$ since $\langle u'w \rangle \gg \langle v'w \rangle$ and if the contribution of the $\bar{v}^2$ term to the kinetic energy balance is ignored therefore then the divergence values given in parentheses are obtained: magnitudes are now significantly less but the incorrect sign remains. It appears therefore that the triple correlation terms are too severely influenced by sensor motion to allow them to be satisfactorily calculated from even low-pass filtered data (Wyngaard and Cote (1971) suggested that over land in unstable conditions the pressure transport term in eq (40) was not negligible but neglect of this seems a less-likely explanation than the effect of ship motion for the discrepancy in the present case).

Values for terms in the balance equation for the temperature variance are given in Table 18. Dissipation could not be estimated with great accuracy since there are usually no well-defined $-5/3$ region. The data of Miyake et al (1970b) suggest a lower-frequency limit for this region somewhat higher than $f = 1$, but at the lower levels this was in the spectral region significantly affected by noise due to sensor motion (figure 34): there was less ambiguity at higher levels.

The production term was calculated on the assumption that the lower level was in the constant-flux layer, but using the measured heat flux at this level. At the upper level for runs 11 and 12 the potential temperature gradient was almost certainly positive (the measured heat flux was directed

Table 18 Temperature variance balance

Run No.	Height (m)	$\chi_T \times 10^5$ ($^{\circ}\text{C}$) ² s ⁻¹	$\langle wT' \rangle^2 \phi_h / k u_{xz}$ $\times 10^5$ ($^{\circ}\text{C}$) ² s ⁻¹	$\langle wT'^2 \rangle / 2$ $\times 10^4$ ($^{\circ}\text{C}$) ² s ⁻¹ m
11	45	1.6	0.01	-0.8
	140	0.9	-	+4.7

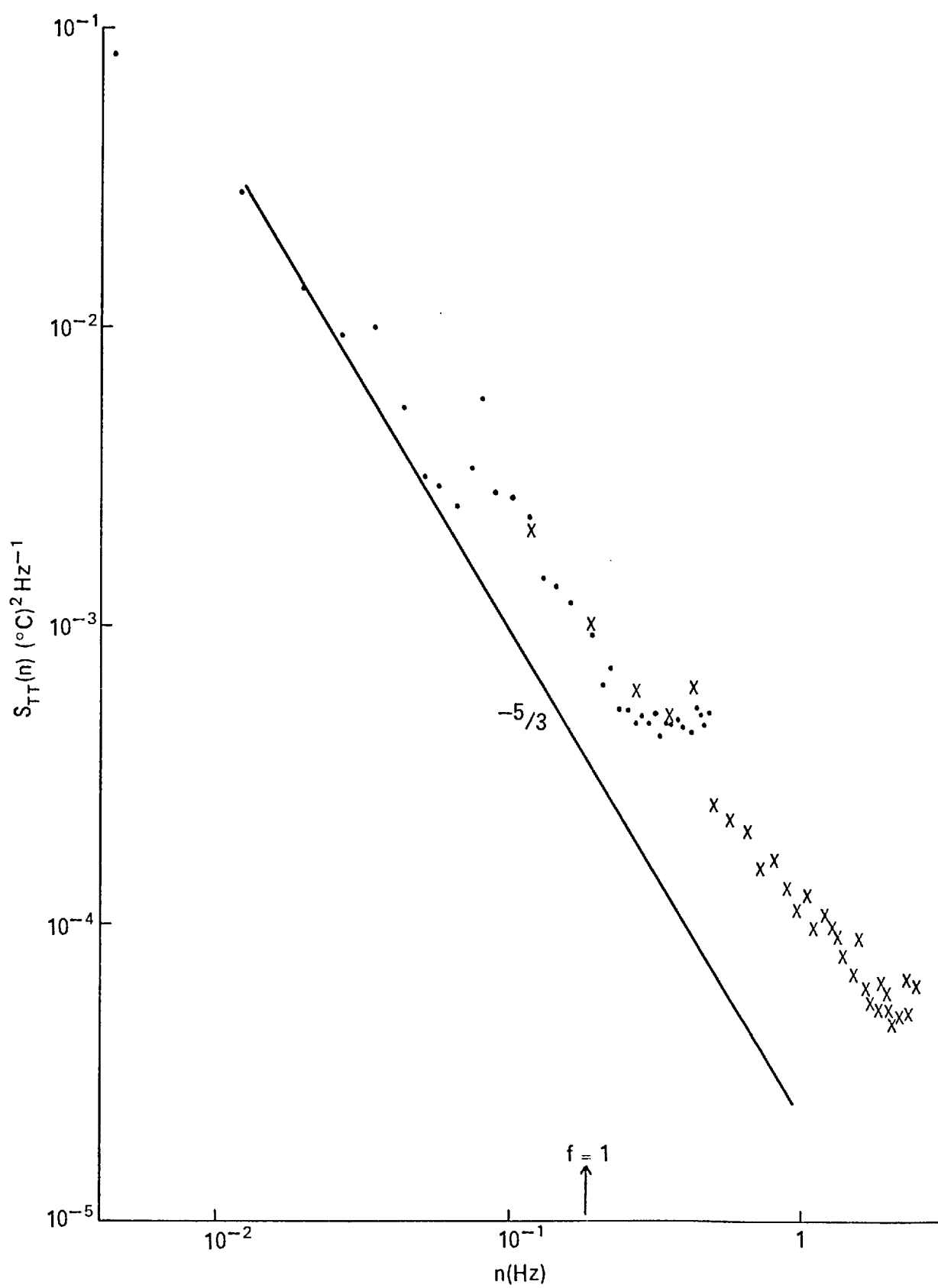


Figure 34 Composite T-spectrum, 140m, 1610 -1738 Z, 17.6.70 (JASIN Run 11)

Table 18 contd.

Run No.	Height (m)	$\chi_T \times 10^5$ ($^{\circ}\text{C}$) ² s ⁻¹	$\langle wT' \rangle^2 \phi_H / k u_* z$ $\times 10^5$ ($^{\circ}\text{C}$) ² s ⁻¹	$\langle wT' \rangle^2 / 2$ $\times 10^4$ ($^{\circ}\text{C}$) ² m s ⁻¹
12	45	1.3	0.33	+1.5
	140	0.6	-	-23.0
13	90	1.4	(0.08)	+3.7
	185	0.5	-	+0.4

downwards), probably a few tenths of a degree centigrade per kilometre, giving positive production around 2×10^{-6} ($^{\circ}\text{C}$)²s⁻¹ or again less than dissipation: in run 13 because of the small heat flux the production also would have been less than dissipation. Thus for all runs and levels dissipation was greater than production. As in the case of turbulent kinetic energy the calculated divergences between the two levels (respectively $+0.6 \times 10^{-5}$, -2.6×10^{-5} , -0.3×10^{-5} ($^{\circ}\text{C}$)²s⁻¹) did not agree well with the estimated average difference of production and dissipation (around -1×10^{-5} ($^{\circ}\text{C}$)²s⁻¹) though the sign at least was correct for runs 12 and 13.

Dissipation of humidity variance (Table 19) was calculated after correcting spectra for the estimated instrumental inertia. Humidity spectra showed comparatively little influence from ship motion (§ 7(b)) and -5/3 regions were found in all cases, extending below $f = 1$ as also found by Miyake et al (1970b), though there was significant noise at high frequencies (above about 0.2Hz): this was probably due to the telemetry and recording system and the phase difference between wet and dry-bulb sensor outputs. Production at the lower level was obtained by assuming a constant-flux layer to this height and using the measured water-vapour flux (eq 46). It was found that on average dissipation decreased with height, though less rapidly than for the temperature variance. At 45m it is seen that production and dissipation were closely balanced. If it is assumed that production was roughly inversely proportional to height above the surface then at greater

heights dissipation was on average larger than production. Table 20 gives average values for the various terms between the two measurement levels. Mean production was estimated on the basis of an inverse proportionately with height and was probably overestimated therefore in run 11 where the vapour flux was much smaller at the higher level. The calculated divergence

Table 19 Humidity variance balance

Run No.	Height (m)	$\chi_q \times 10''$ (s ⁻¹)	$\langle wq \rangle^2 \phi_\epsilon / \kappa u_{xz}$ $\times 10''$ (s ⁻¹)	$\langle wq'^2 \rangle / 2$ $\times 10''$ (m s ⁻¹)
11	45	2.0	2.5	+2.0
	140	2.3	-	-0.8
12	45	4.9	6.2	+5.9
	140	2.6		+4.3
13	90	1.6	(0.6)	+2.8
	185	0.9	-	+2.1

of the vertical flux of humidity variance is seen to have the correct sign in each case to balance the variance budget though the value was on average a factor of two too large.

Table 20 Mean humidity variance balance

Run No.	Dissipation $\times 10''$ (s ⁻¹)	Production $\times 10''$ (s ⁻¹)	(Production- dissipation) $\times 10''$ (s ⁻¹)	Divergence $\times 10''$ (s ⁻¹)
11	2.1	1.3	-0.8	-2.9
12	3.7	3.3	-0.4	-1.7
13	1.3	0.4	-0.9	-0.7

7.9

BOUNDARY-LAYER STRUCTURE

A typical boundary layer near OWS I or J is convectively unstable but topped by a stable layer, with the cloud usually non-precipitating cumulus (Charnock and Ellison 1967): such a structure occurs widely in the Trade wind regions also and was indeed the predominant type during ATEX and also frequently present in JASIN. The results from these expeditions have shown in general in these circumstances that the upward flux of heat from the surface decreases rapidly with increasing height but the moisture flux in contrast varies much more slowly. Thus this type of marine boundary layer with its turbulence structure controlled to a large extent by the upward flux of water-vapour from the surface, differs essentially from that over land in unstable conditions where it is the heat flux which exerts the predominant control over the structure.

A simple boundary-layer model which is compatible with the present observations may be developed along the following lines. It is assumed that mixing is maintained primarily by the convective action of moist air in a layer (usually less than 3,000m deep) which is capped by a dry inversion. Typically there will be condensation and cumulus formation at heights greater than a few hundred metres and the released latent heat will allow further vertical development. Now particular features of the present observations are the downward heat fluxes and positive potential temperature gradients below cloud base, implying a source of sensible heat in the upper part of the boundary layer: however the cumulus clouds decay continuously and the latent heat released through condensation is therefore constantly re-absorbed, the net heating rate being zero in these circumstances. On the other hand the moist rising elements will penetrate occasionally the warm dry lid at the top of the mixed layer and entrain air from there. This guarantees a source of sensible heat near the top of the layer and retards

the rate of increase of specific humidity in this region. Subsequently the entrained heat is transported downward by mixing. Another way in which a net release of latent heat can occur is as a result of precipitation but although this process exists frequently in deeper boundary layers it was not relevant to ATEX where showers were very rare.

Theories of the boundary layer structure such as those described by Smith (1968) in general do not appear to be appropriate to describe the type of structure now being discussed since they ignore the crucial contribution of entrained warm air at the top of the layer. That of Charnock and Ellison while specifically including entrainment has important limitations in other directions: for example the assumption that the profiles of potential temperature, humidity and virtual potential temperature are similar is completely at variance with observations from the present experiments, and the model's neglect of turbulence resulting from the release of latent heat is incorrect. The model of Ball (1960) also considered entrainment but was based on the assumption that the production of turbulent kinetic energy by buoyancy forces was balanced chiefly by increases of gravitational potential energy due to change of height of the capping inversion caused by turbulent entrainment of air from it into the boundary layer: viscous dissipation was ignored by comparison. However the present results (§ 7(h)) and those of Lenschow (1970) suggest that below cloud base at least the majority of the turbulent kinetic energy is dissipated by viscous forces and not exported to higher levels to erode the capping inversion. Clearly then the necessary turbulence is generated in the upper part of the boundary layer as a result of the release of latent heat associated with the moist convection process.

It appears that a practical model of the convective marine boundary layer in order to be realistic will have to be significantly more complicated

than most of those which have already been proposed. The appropriate external parameters will include presumably z , h (the boundary-layer depth), f , V_g , $\Delta\theta$ and Δq (differences across the full boundary-layer depth) and g/θ , but if baroclinicity is too large to be neglected a further parameter $\langle \partial V_g / \partial z \rangle_{z=0 \text{ to } h}$ will then be required. The importance of this last was well demonstrated by the results of Charnock et al (1956) in the highly baroclinic Trades. They found the typical ratio of the near-surface wind to the surface geostrophic wind to be around 0.5, whereas Findlater et al's result showed the ratio of surface to 900mb wind to be around 0.9 near OWS I and J in slightly unstable conditions (here the 900mb wind can be identified approximately with the surface geostrophic wind because the effect of baroclinicity was likely to have been small on average). The rate of sensible heat transfer at the surface is usually small compared with that over land and in these circumstances the boundary-layer structure is likely to be significantly influenced by net radiational cooling which thus is another variable to be expressed in terms of the external parameters along with quantities such as τ , H , E , u , v , w and α . Release of latent heat and the development of cloud will usually occur over the sea and this must also be included in any realistic model as well as entrainment through the upper boundary. Additionally once the cloud has developed the radiation structure of the layer will be substantially modified: in particular there may be significant radiational warming at cloud base. The relatively straightforward similarity treatments of the boundary layer such as that of Gill (1968) are quite inadequate to deal with the complex structure just described and indeed it is possible in these circumstances that no adequate similarity treatment will be simple enough to form a useful basis for parameterizing for example sub-grid scale processes.

Chapter 8

SUMMARY AND CONCLUSIONS

A tethered-balloon system operated from a ship has been developed for measuring turbulent fluctuations of horizontal and vertical wind speed, temperature and humidity at several heights up to a few hundred metres over the open sea, and has been used near OWS 'Juliett' and in prototype form in the Atlantic North-Easterly Trades. The ship's wave - induced motion produced an alternate tightening and slackening of the balloon-cable catenary which led to substantial, nearly horizontal, oscillations of the sensors and very large contamination of both horizontal and vertical wind-speed signals (the system for sensing vertical wind was pendulum-orientated to the apparent vertical). The temperature and particularly humidity signals were much less affected. The majority of the noise was concentrated in a comparatively-narrow frequency band which lay at the high frequency end of the cospectra, and could be filtered out therefore without much attenuation of turbulent contributions. A similar treatment when applied to the measured variance spectra resulted in significant losses of genuine turbulent fluctuations.

Vertical fluxes of momentum, moisture and heat were calculated by the eddy-correlation method, using the corresponding cospectra to identify and quantify the contributions due to sensor motion. The results usually supported a comparatively rapid decrease of upward heat flux with height but in contrast the measured water-vapour fluxes typically showed much slower variations. The measurements near Juliett produced a value for C_E of $(1.8 \pm 0.7) \times 10^{-3}$. The data from the Trades suggested a value for the drag coefficient C_D (referred to 10m wind) probably somewhat in excess of 1×10^{-3} but because of strong baroclinicity and the comparatively large average height above surface ($\sim 170\text{m}$) the estimate was very tentative.

However there was evidence that the measured momentum fluxes were under-estimates since the Uw -cospectra usually showed positive contributions at around 3×10^{-2} Hz, due presumably to sensor motion. This feature was confirmed by the data from near Juliett where a value for C_D of 1.1×10^{-3} was found.

From data obtained at two heights on three slightly unstable occasions there appears to be a very rough balance between production and dissipation of vapour fluctuations, and mechanical production and dissipation of turbulent kinetic energy at heights of a few tens of metres but dissipation was found to exceed production at all heights for the temperature fluctuations, and at heights above about 100m for moisture and kinetic energy. The technique was unable to produce plausible estimates for the divergence terms which being triple correlations were strongly influenced by sensor motion.

The heat and vapour flux results from both sets of experiments suggested that in slightly unstable conditions the buoyancy flux and hence the turbulence structure in the bulk of the boundary layer is determined primarily by the vertical flux of moisture. It appears that existing similarity treatments are quite inadequate to represent the boundary layer over the sea: suitable models may have to pay attention to not only turbulent entrainment at the layer top but also the radiation balance within the layer.

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APPENDIX:COPIES OF RELEVANT PAPERS BY THE AUTHOR

- Paper 1 Relation between turbulent fluxes over the sea and near-surface
mean parameters
- Paper 2 The effect of sensor motion on data collected by captive
balloon-borne turbulence instrumentation.

Meteorological Office
Boundary Layer Branch (Met O 14)

Turbulence and Diffusion Note No 14 (May 1971)

RELATIONS BETWEEN TURBULENT FLUXES OVER THE
SEA AND NEAR-SURFACE MEAN PARAMETERS

by N Thompson

Note As this paper has not been published, permission to quote from it should be obtained from the Head of the above Branch of the Meteorological Office.

May 1971

Relations between turbulent fluxes over the sea and
near-surface mean parameters

by N Thompson

Deardorff (1968) discussed the variations with stability of the transfer coefficients which are often used to relate vertical fluxes of momentum, heat and moisture to near-surface mean parameters over the sea. In the course of the analysis of the data obtained during the Joint Air-Sea Interaction Experiment (JASIN), it has been necessary to estimate the stability dependence of these coefficients: Deardorff's formulations, which are based on analytical forms for the empirical functions in the Monin-Oboukhov relations for the constant flux layer and which give the results in terms of bulk Richardson numbers, have not been found convenient to use so alternatives have been derived which are described in this note. A brief discussion is also included of the likely quantitative relations between the three coefficients.

The present treatment is also based on similarity theory for the constant flux layer and makes use of the recent estimates of Dyer and Hicks (1970) for the necessary empirical functions. The governing equations are with the usual notation,

$$\frac{\partial U}{\partial z} = \frac{u_*}{kz} \phi_u(z/L) \quad (1)$$

$$\frac{\partial \theta}{\partial z} = \frac{T_*}{kz} \phi_\theta(z/L) \quad (2)$$

$$\frac{\partial q}{\partial z} = \frac{q_*}{kz} \phi_q(z/L) \quad (3)$$

where $u_* = (\tau/\rho)^{1/2}$, $T_* = -H/\rho c_p u_*$, $q_* = -E/\rho u_*$

and $L = -u_*^3 T \rho c_p / k_3 (H + 175E)$: in this last T has been taken as 15°C , H is in Watts cm^{-2} and E is $\text{gm cm}^{-2} \text{sec}^{-1}$ (c_p is $1.00 \text{ J gm}^{-1} \text{ }^\circ\text{C}^{-1}$). Integrating (1) - (3),

$$\frac{k}{u_*} (U - U_0) = \log z/z_0 - \psi_u(z/L) \quad (4)$$

$$\frac{k}{T_*} (\theta - \theta_0) = \log z/z_0 - \psi_\theta(z/L) \quad (5)$$

$$\frac{k}{q_*} (q - q_0) = \log z/z_0 - \psi_q(z/L) \quad (6)$$

where
$$\psi = \int_{z_0}^z \frac{1 - \phi}{\xi/L} d(\xi/L)$$

Only the meteorologically important case of unstable density stratification will be considered here, and in these circumstances the tabulations of ψ given by Dyer and Hicks can be used.

Eq. (4) can be written

$$u_*^2 = C_D (U - U_0)^2$$

where $C_D = \left\{ k / (\ln z/z_0 - \psi_n(z/L)) \right\}^2$: in neutral stability $C_D(N) = (k / \ln z/z_0)^2$
Conventionally U is measured at a height of 10m. Wind tunnel data of Keulegan (1951) suggest that U_0 (the sea-surface drift velocity) is about 4% of U . Brocks and Krugermeyer (1970) have analysed numerous wind profiles obtained over the sea in near-neutral conditions in winds up to 12 m sec^{-1} and concluded (assuming $U_0 = 0$) that $C_D(N) = 1.3 \times 10^{-3}$, with $z_0 = 1.2 \times 10^{-2} \text{ cm}$: there was some evidence from their data for a few percent increase in C_D in going from low speeds to the highest observed. Smith (1970) found $C_D = 1.35 \times 10^{-3}$ after averaging values found in both stable and unstable conditions in winds between 7 and 16 m sec^{-1} (he also assumed $U_0 = 0$): there was again an indication of a very small increase with increasing speed. Other recent determinations of C_D (in wind speeds below 12 m sec^{-1}) gave values of 1.21×10^{-3} (Hasse 1968) and 1.09×10^{-3} (Miyake et al 1970) - these authors also assumed U_0 to be zero. All these data suggest that C_D varies with wind speed by no more than a few percent in winds up to about 15 m sec^{-1} , and it will be assumed therefore that $C_D(N)$ is a constant, value around 1.3×10^{-3} (here assuming $U_0 = 0.04U$): the uncertainty is probably less than about 20%.

The assumption is also made that z_0 , z_T and z_p are constants, independent of wind speed. This is likely to be the case because z_0 at least is virtually independent of wind speed in neutral conditions, and a stability change for a given 10m wind speed results in a change of wind speed at lower heights - this is equivalent, as far as conditions near the surface are concerned, to a change of wind speed at 10m in neutral stability.

Equations (4) and (5) can be combined to give

$$\left\{ (U - U_0) / (\theta - \theta_0) \right\} T_* / u_* = \left\{ \ln z/z_0 - \psi_n(z/L) \right\} / \left\{ \ln z/z_T - \psi_n(z/L) \right\}$$

$$\text{or } H = \rho C_p (U - U_0) (\theta - \theta_0) C_D(N) \left\{ \ln z/z_0 / (\ln z/z_0 - \psi_n(z/L)) \right\}^2 \left\{ \frac{\ln z/z_0 - \psi_n(z/L)}{\ln z/z_T - \psi_n(z/L)} \right\}$$

with a similar expression for E . If θ_0 (and q_0) are equated to the surface values of

temperature and saturation specific humidity, as would be done conventionally, the variations of all the exchange coefficients with stability can be obtained. Thus

$$C_{H,E} / C_{H,E}(N) = \log z/z_0 \cdot \log z/z_{T,E} / \{ (\log z/z_0 - \psi_H(z/L)) (\log z/z_{T,E} - \psi_{H,E}(z/L)) \} \quad (8)$$

$$C_D / C_D(N) = \{ \log z/z_0 / (\log z/z_0 - \psi_H(z/L)) \}^2 \quad (9)$$

Table 1 tabulates these variations with stability, assuming that z_0 , $z_{T,E}$ and z_q are equal, and $C_D(N) = 1.3 \times 10^{-3}$. Selecting a different value for $C_D(N)$

Table 1 Variation of transfer coefficients with stability (using values for ψ given by Dyer and Hicks (1970))

$-z/L$	$-L (z=10m)$ (m)	$C_D / C_D(N)$	$C_{H,E} / C_{H,E}(N)$
0.01	1000	1.00	1.00
0.02	500	1.00	1.01
0.05	200	1.01	1.03
0.10	100	1.03	1.06
0.20	50	1.07	1.11
0.50	20	1.14	1.21
1.00	10	1.21	1.31

makes little difference: for example, with $C_D(N) = 1.1 \times 10^{-3}$ and $z/L = -1.0$, the ratio $C_D/C_D(N)$ becomes 1.19, a change of only 2%.

Now,

$$L = -u_*^3 T_e c_p / k_g (H + 175E)$$

$$= -(U-U_0)^3 C_D^{3/2} T_e c_p / k_g \rho c_p C_{H,E} (U-U_0) (\theta_0 - \theta_{10} + 175(q_0 - q_{10}))$$

Multiplying top and bottom by $(C_D(N))^{1/2} (\approx 3.6 \times 10^{-2})$ and putting $T = 288^\circ K$, this reduces to

$$L(m) = \left\{ -2.6 (U-U_0)^2 / (\theta_0 - \theta_{10} + 175(q_0 - q_{10})) \right\} \cdot \frac{C_D^{3/2}}{C_{H,E}(C_D(N))^{1/2}}$$

The maximum error in L , for $-z/L \leq 1$, which results from assuming $C_D = C_{H,E} = C_D(N)$ is less than 3%, which produces less than 1% error in the estimates for C_D and $C_{H,E}$.

Hence we can write adequately

$$L = -2.4 U^2 / (\theta_s - \theta_o + 175(q_s - q_o)) \quad (10)$$

If values for $10/L$ are calculated for given U and $(\theta_s - \theta_o + 175(q_s - q_o))$ and the corresponding $C/C(N)$ are extracted from table 1, then simple nomograms showing the variations of $C/C(N)$ with stability may be constructed (figure 1). Thus, for $\theta_s - \theta_o = 2^\circ\text{C}$, $q_s - q_o = 5 \times 10^{-3}$ and $U = 5 \text{ m sec}^{-1}$, then $C_u/C_u(N) = 1.10$. We see immediately that for winds above about 10 m sec^{-1} the effects of stability on the transport coefficients are small even when very large values are entered for air-sea temperature and humidity differences. In lighter winds the effects are significant but would not be easily discernible in measured fluxes unless winds are below about 5 m sec^{-1} .

It is of interest now to consider the likely relations between C_D , C_u and C_e if it is no longer assumed that z_o , z_T and z_a are equal, and if account is taken of the fact that for practical reasons q_o , θ_o and U_o are not necessarily measured (or estimated) at the same height. The published data are in some conflict. For example eddy correlation measurements by Hasse (1968) suggest that in near-neutral conditions $C_u \approx 1.0 \times 10^{-3}$ but budget studies by Robinson (1966) give $C_u \approx 1.3 \times 10^{-3}$ a value similar to that found by Chamberlain (1968) for C_e from wind-tunnel investigations of evaporation.

Owen and Thomson (1963) derived the empirical relation

$$1/B = \frac{1}{k} \log z_o/z_{T,a} = 0.52 \left(\frac{30 u_* z_o}{\nu} \right)^{0.45} \left(\frac{\nu}{kD} \right)^{0.8} \quad (11)$$

valid for $\frac{u_* z_o}{\nu} > 3.3$: here ν is the kinematic viscosity and k and D are the thermometric conductivity, and the molecular diffusion coefficient for water vapour. Now $z_o \approx 1.2 \times 10^{-2}$ and it follows that the relation is not expected to hold for $u_* < 40 \text{ cm sec}^{-1}$ or in winds less than about 12 m sec^{-1} . Chamberlain (1968) carried out wind tunnel experiments on the evaporation of water vapour from rough surfaces which gave results in approximate agreement with eq. (11) but also had no data corresponding to winds less than about 15 m sec^{-1} over the sea. However, using his

extrapolating curve to the value for $1/B$ corresponding to flow over smooth surfaces, then for $U = 10 \text{ m sec}^{-1}$ or $\frac{U_* z}{\nu} \approx 3$, we obtain a value for $1/B$ around 2. In this case, $z \approx 2.2 z_0$, which in near-neutral conditions gives $C_h/C_s \approx 0.94$. A similar calculation for $U = 20 \text{ m sec}^{-1}$ gives $C_h/C_s \approx 0.91$.

Because of the factor $(\nu/k, D)^{-0.8}$ in eq. (11) it follows that $1/B$ for heat transport is 1.08 times that for water vapour, which suggests that C_h and C_s can be equated with negligible error. It appears therefore that the ratio C_h/C_s , in near-neutral conditions, decreases slowly with increasing wind speed but that the ratio departs by only about 10% from unity even in very strong winds (this is of course neglecting evaporation of spray droplets).

Stewart (1961) pointed out that significant momentum is likely to be transported to the sea in the process of wave generation and maintenance, and because the effective level at which this momentum is given up must be above that where wind speed is equal to the phase velocity of the waves, then the vertical flux of momentum must decrease as the surface is approached. Hence the turbulent mixing near the surface is less than would otherwise be expected and so in this region the transfer coefficients for heat and water vapour will also be less than the measurements at greater heights might suggest. As a result there will be an increase in gradients of temperature and moisture which will help to maintain fluxes close to those anticipated in a wave-free situation, but it is not clear that complete compensation will occur: the ratio C_h/C_s may thus be decreased.

Finally, in practical applications of the bulk formulations for heat and moisture transfers it is usual to equate Θ_0 (and hence q_0) to temperature measured at a depth of a few tens of centimetres. This introduces two sources of error. First, it is well known that in unstable conditions a cool "skin" forms at the surface where the temperature is typically around 0.2 deg C lower than the bulk water temperature. Thus the use of the bulk value should produce overestimates for both heat and moisture fluxes. Secondly, it is not correct to equate Θ_0 and q_0 to the surface (skin) values Θ_s and q_s , but if this is done a simple order of

magnitude argument suggests that overestimates of a few percent in the heat and moisture fluxes may result. Thus the thickness of the viscous boundary layer (δ_v) is of order ν/u_* (Kraus 1966), and this is approximately equal to δ_T and δ_q , the thermal and water vapour boundary layer thicknesses (Schlichting 1968). Taking z_0 as 1.2×10^{-2} it then follows that for winds less than about 10 m sec^{-1} , $\delta_{T,q} > z_0$. Hence except in strong winds we can assume that temperature varies linearly with height from surface up to $z = z_0$. Now, $z_{T,q} \sim z_0 \sim 1 \times 10^{-2} \text{ cm}$; hence,

$$H = \rho c_p C_H (\theta_0 - \theta_{1,0}) (u_{1,0} - u_0) = k (\theta_s - \theta_0) / 1 \times 10^{-2}$$

where k is the thermal conductivity. Putting $C_H = 1.3 \times 10^{-2}$ and $u_0 \sim 0$,

$$\begin{aligned} \theta_s - \theta_0 / (\theta_0 - \theta_{1,0}) &\sim (\rho c_p u_{1,0} / k) \cdot 1.3 \times 10^{-5} \\ &\sim 6 U_{1,0} \times 10^{-5} \end{aligned}$$

The corresponding expression for q is

$$q_s - q_0 / (q_0 - q_{1,0}) \sim 5 U_{1,0} \times 10^{-5}$$

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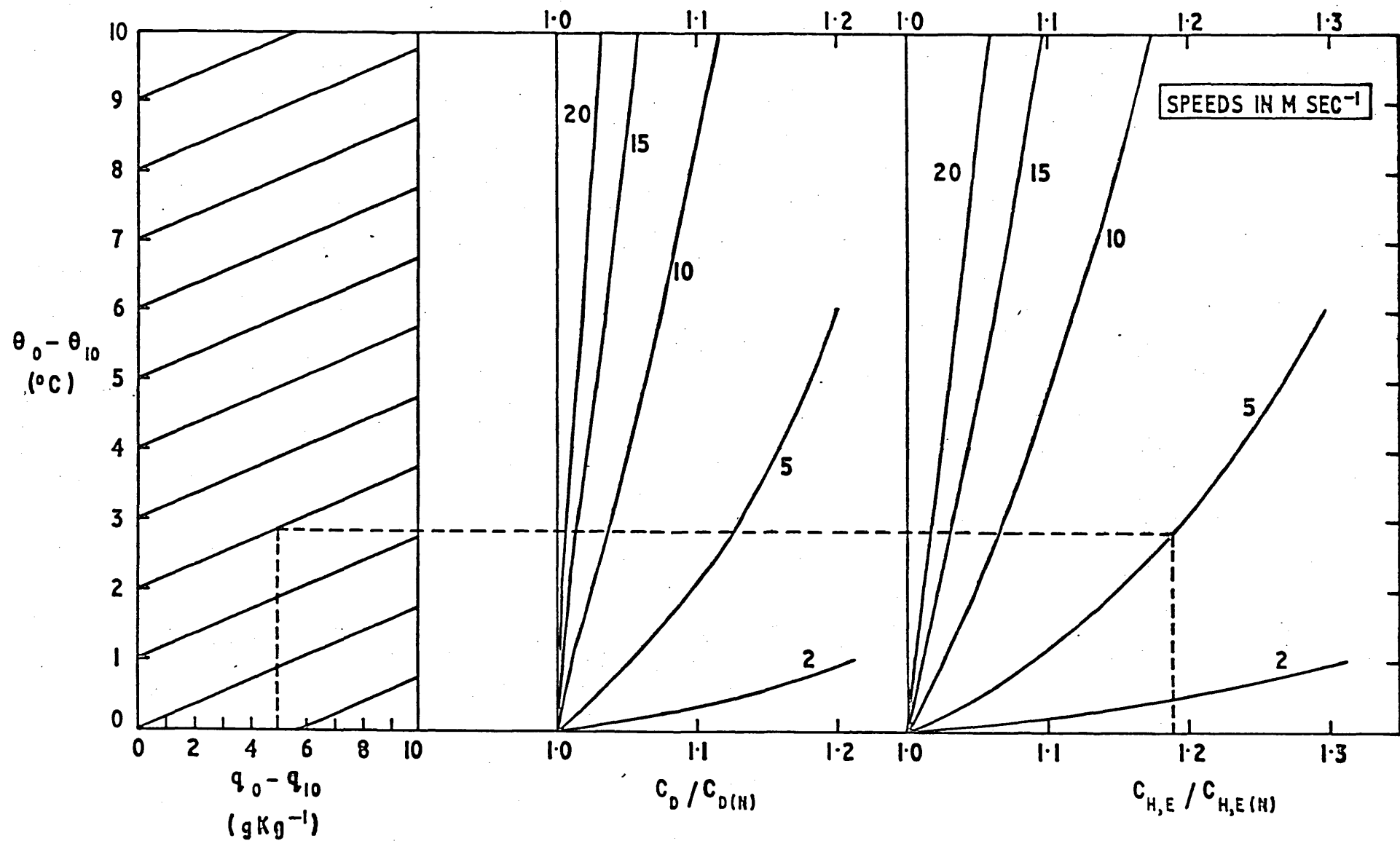


FIGURE 1. VARIATIONS OF TRANSFER COEFFICIENTS
WITH WIND SPEED AND AIR-SEA
TEMPERATURE AND HUMIDITY DIFFERENCES

THE EFFECT OF SENSOR MOTION ON DATA COLLECTED BY
CAPTIVE-BALLOON-BORNE TURBULENCE INSTRUMENTATION

by N. Thompson

1. Introduction

Captive balloons have been used for a number of years in the Meteorological Office research programme for supporting turbulence sensors above ground level, but only a limited effort has been made to establish the effects of balloon movements on sensor outputs (e.g. Jones (J.I.P.) and Butler 1958, Jones (R.A.) 1958). These effects have been considered small usually, but there is obviously a need for a more critical investigation before drawing more definite conclusions. The present note describes the results of a number of experiments carried out specifically to measure balloon-induced sensor motions, and their effect on vertical momentum fluxes.

2. Background to the experiments

If u_s, v_s, w_s are the three components of sensor motion at some instant of time, and u, v, w are the corresponding wind velocity components (with the usual orientation of axes), then the horizontal and vertical wind components inferred from the sensor outputs are $\left((u-u_s)^2 + (v-v_s)^2 \right)^{1/2}$ and $w-w_s$. The Cardington instrumentation cannot measure the individual horizontal components, so that even if all three components of sensor motion can be measured, it is not yet possible to correct the measured horizontal wind for sensor motion (However, later in this note an approximate method for correcting the horizontal wind is suggested, which may have some uses). Nevertheless, corrections for vertical sensor motion alone may still be very valuable in the context of the vertical flux of momentum. Thus, the measured momentum flux is

$$\tau = - \rho (w-w_s) \left\{ \left((u-u_s)^2 + (v-v_s)^2 \right)^{1/2} - \left((u-u_s)^2 + (v-v_s)^2 \right)^{1/2} \right\}$$

(It is assumed that $\bar{w} = \bar{w}_s = 0$).

A closer approximation to the true flux is likely to be obtained by correcting for vertical sensor motion, equivalent to putting $W_s = 0$ in the above expression. Further, in the event of u_s and v_s being uncorrelated with W , and with u and v respectively, one would have then the true flux, but in practice there is no good reason to suppose that this is the case. A complete correction procedure must await the development of a technique for measuring wind direction by balloon-mounted instrumentation.

When considering the vertical flux of quantities other than momentum, although the sensor motion may still introduce spurious correlations between vertical velocity and fluctuations of the quantity being transported in the vertical, the effect is likely to be smaller than in the case of momentum, and may well be eliminated almost entirely by a correction for vertical sensor motion.

Most of the remainder of this note describes a series of experiments in which measurements of the movements of the "BALGUST" instrument⁽¹⁾ when attached to the cable of a captive balloon were made using theodolite techniques, the intention being to obtain data on the typical magnitude of balloon-induced sensor movements, and particularly their effect on measured momentum fluxes.

3. The experiments

A short pilot series was carried out in September 1967, in which a single theodolite was mounted very near the tethering point of a captive balloon, and used to follow balloon movement over periods of about twenty minutes. The turbulence-measuring instrumentation was not flown on these occasions. The theodolite was one of a pair normally used at Porton for pilot-balloon tracking, in which helical potentiometers were connected to the azimuth and elevation controls by suitable gearing. The outputs were voltages proportional to azimuth

(1) As used at Cardington since 1968: described in MRCP 234

and elevation and these were recorded on the Cardington data-logger (DDR8) at 5-second intervals (on subsequent analysis this 5-second interval was found to be too large, and later experiments used 1-second sampling). Use of a single theodolite for these measurements was based upon the assumption that the balloon cable was an effectively rigid structure which pivoted around its tethering point. On this basis, observations of azimuth and elevation were converted to components of balloon motion in all three dimensions. Of course, such a technique provided no information on vertical velocities of the balloon resulting from changes in shape of the cable catenary, and hence possibly underestimated the vertical velocities. However it should have given good estimates for horizontal velocities. "Good" in this context is a relative term, depending on both the skill of the observer in following the balloon motion and also the resolution of the data-logger. Because the Porton theodolites had been modified specifically for tracking pilot balloons, the resolution was rather low for the present usage (0.06 degrees for elevation, 0.2 degrees for azimuth). Maximum vertical velocities appeared to be in excess of 20 cm/sec^{-1} , and horizontal velocities up to about 1 m sec^{-1} were also measured.

In the main series of experiments the two theodolites were set up at either end of a base-line about 700m long. Optimum resolution of the vertical motion of the sensor system attached to the balloon cable is obtained when the angle of elevation of the sensors with respect to each theodolite is 45 degrees. It was not possible to make the base-line long enough to achieve this, so that the elevations were usually in the range 50 to 60 degrees. The base-line orientation was dependent on the position of various obstructions on the Cardington site, and it was not the ideal one for resolving horizontal sensor motions (the ideal occurs when the bearing of the sensors from the base-line is 45 degrees at each theodolite: The angles usually subtended were between 20 and 30 degrees). Errors in tracking the horizontal motion of the sensors of course introduce errors in calculated values of the vertical sensor motion so that in the present series of experiments, the

/conditions

conditions for minimising random errors in the measured vertical velocities were not achieved. It was therefore anticipated that smoothing of the apparent sensor velocity components would be necessary to obtain coherent data on the sensor motions.

In this second series of experiments the "Balgust" instrument was mounted on the balloon cable about 70m below the balloon and approximately 600m above ground, and movements of this instrument, rather than those of the balloon, were followed using the theodolites. The "Balgust" inclination and wind speed signals, together with the theodolite outputs, were led to the data logger which sampled the six inputs every second. Details of the experiments appear in the Appendix. All experiments were carried out in unstable conditions, with the exception of No. 9, when the "Balgust" instrument remained above the convective layer throughout the experiment.

4. Data analysis

The data tapes were processed on the "Mercury" computer at Porton. The theodolite data were reduced to give "Balgust" positions at 1-second intervals, using a right-handed coordinate system with origin at the Southern theodolite, X - axis along the base-line and Z - axis vertical (Thyer 1962). The components of sensor velocity along the axes can then be evaluated but usually these axes are not conveniently aligned with respect to the mean wind direction at the sensors. However a method of alignment can be obtained from consideration of the general characteristics of the balloon motion. Thus, a feature of the behaviour of these large captive balloons is that they tend to drift from side to side, across the mean wind direction, in preference to alongwind. Therefore one expects the successive sensor positions, referred to a horizontal plane, to lie near an arc symmetrically orientated acrosswind. This suggests a method of estimating the mean alongwind direction at the balloon level (more approximately at the sensor level also) from the theodolite data. Referring to Figure 1, the dotted line has been drawn to enclose the projection of the successive sensor positions onto the

/horizontal

horizontal ($x-y$) plane. If a line AA' is drawn through the points as shown, by a least-squares method, then the perpendicular BB' defines approximately the mean wind direction at the sensor level (note in this example that the wind blows from B to B'). If BB' makes an angle Θ with the x -axis, then by rotating the horizontal axes about the vertical through this angle, the new positions of these axes (x', y') define the mean alongwind and acrosswind directions. In practice this technique fails to distinguish between BB' and $B'B$, and supplementary wind data are needed to do this.

The method outlined above was used with the present data to define the successive coordinates of the "Balgust" instrument (x', y', z) and by subtraction the sensor velocities (u_s, v_s, w_s) were obtained. Due to the lack of resolution involved in the theodolite technique, and also the impossibility of following the instrument motion exactly, the sensor velocities fluctuated wildly in magnitude, and it was clear that some quite drastic smoothing was necessary to obtain a coherent picture. Initially this was done by using observations in pairs up to 6 seconds apart (i.e. $u_s = (x'_{n+6} - x'_n)/6$), but this smoothing was found to be insufficient, and finally an additional smoothing by groups was incorporated:

$$u_s = (x'_{n+7} + x'_{n+6} + x'_{n+5} - x'_{n+1} - x'_n - x'_{n-1})/18 \quad \text{etc}$$

Even this smoothing was insufficient to remove all "noise", and because w_s was usually the smallest of the components, its behaviour was still very erratic. By comparison, u_s and v_s showed smoother variations with time. It was felt that further smoothing was undesirable, leading as it would to further reduction in the magnitude of the true fluctuations of sensor velocity.

In all, there were seven experiments in which usable data were obtained. In three of these, the wind occasionally became too light to operate the anemometer; the wind speed was then taken to be the stopping speed of the anemometer. The data were used to compute fluxes of momentum in two ways. The "Balgust" instrument gave outputs proportional to total wind speed and wind inclination, and from these the instantaneous apparent horizontal and vertical velocities u_a and w_a were
/obtained.

obtained. In the first flux computation, it was assumed that the sensor motion was negligible; then

$$\tau_1 = -\rho \overline{(w_a - \bar{w}_a)(u_a - \bar{u}_a)}$$

$\bar{w}_a$ and $\bar{u}_a$ were evaluated from the whole length of each record, but the record was also divided into 1-minute lengths and the contributions to τ_1 were also evaluated for these short periods: the results appear in Table 1.

The second computation involved correcting w_a for both non-zero mean and for vertical sensor motion, giving

$$\tau_2 = -\rho \overline{(w_a + w_s - \bar{w}_a - \bar{w}_s)(u_a - \bar{u}_a)}$$

This was again done for the whole record, and for one-minute periods, for direct comparison with τ_1 , and the results appear also in Table 1.

(No use has been made of the wealth of data on horizontal sensor motion. Any attempt to take this into account in the present context can only be speculative because of lack of measured horizontal wind data in component form, and for this reason it was felt that any approximate treatment to try to correct momentum fluxes for horizontal as well as vertical sensor motion would be too crude to be of real value. On the other hand, if the data could be used in a plausible way to provide better estimates for mean wind speed this would be very valuable. The nature of the data precludes precise treatment: but making the assumption that alongwind movements of the sensors can be ignored because their effect will tend to be averaged out of the mean speed, then, provided the lateral fluctuations of the actual wind are not too large one can write

$$u_a^2 \approx V^2 + u_s^2 \quad (\text{where } V \text{ is the true horizontal wind speed})$$

$$\text{or } V \approx (u_a^2 - u_s^2)^{1/2}$$

Now U_z is approximately a linear function of height (because the balloon cable tends to swing as a whole), whereas V increases with height as Z raised to a power less than 1. Hence $\frac{\partial U_a}{\partial Z}$ will be greater than $\frac{\partial V}{\partial Z}$, a point of some importance in kinetic energy balance studies).

5. Discussion of results

The computed vertical (downward) momentum fluxes are given in detail in Table 1. They appear here as the 1-minute averages, calculated using values for $\overline{W}$ and $\overline{U_a}$ derived for the complete runs. Thus, the averages of these 1-minute fluxes is the momentum flux for the whole period (Table 2). An obvious feature of the results is the enormous variation that can occur in the flux on both short and long term bases. The experiments listed were all carried out in unstable conditions, and it seems that very long sampling times, or alternatively ensemble averaging, are required in order to obtain meaningful momentum fluxes in convective conditions with cumulus development.

Table 2 summarizes the main results. The point of interest is that there appears to be a strong correlation between wind speed and the difference between corrected and uncorrected values for momentum flux. Apparently even in winds as light as 5 m sec^{-1} it is not permissible to ignore the effects of sensor motion on computed momentum fluxes. However, as pointed out earlier it is uncertain how horizontal sensor motion affects the measured flux. If the correlation between vertical wind speed and horizontal sensor motion is small (as would probably be the case if the balloon was several hundred metres above the sensors), then a precise correction for vertical sensor motion should effectively eliminate the contribution of sensor motion to the measured momentum flux. Confirmation of this point must await instrumented developments now in progress to enable measurements of wind direction to be made by balloon-borne devices.

It appears from Table 2 that the momentum flux was directed upwards rather than downwards in a number of the experiments, implying a decrease of wind with height at the sensor level on these occasions. To see whether this was the case,

the wind speeds were extracted from the relevant Balthum ascents (Table 3). However, none of the Balthum ascents coincided with the experimental runs, and so no satisfactory confirmation of this point could be obtained.

Finally, Table 4 gives data to assess the effectiveness of the technique used to define the mean wind direction from the double theodolite observations. Of the six occasions when the wind direction was reasonably specified by the vane mounted on the 40m tower, in four cases the directions from this agreed to within about 20 degrees with those derived from the theodolite data. Errors in the other two cases were 50 and 60 degrees, and while in one of these the wind was very light, in the other the wind speed was around 5 m sec^{-1} . This casts doubt on the effectiveness of this method for assessing the wind direction at the balloon: unless the point is clarified by further experiments, it suggests the tower wind direction is probably a more reliable estimate of the wind at the sensors in cases where the boundary layer is well-mixed.

6. Conclusions

The results of these experiments appear to indicate that except in very light winds, vertical eddy-fluxes of momentum derived from Balgust data may be in error due to balloon-induced motions of the Balgust instrument. This conclusion must remain tentative until both wind speed and direction are measured by the balloon-supported equipment, so that corrections for sensor motion can be applied to both the measured vertical and horizontal wind speeds. Until this is done it is recommended that in future measurements by the Balgust instrument, the balloon should be flown several hundred metres above the sensor level and the vertical sensor motion should be observed. Then, correcting the Balgust data for vertical sensor motion should lead to better estimates for the vertical eddy flux of momentum.

The technique described in this note for determining the sensor motion is cumbersome and it would appear worthwhile to investigate the use of accelerometers for this purpose. Strictly the accelerometer measurements should be supplemented by observations on the variation with time of the tilt of the accelerometer array from the true vertical, as well as its orientation in the horizontal plane, but

the former is likely to be small in practice: if circumstances are found when this is not the case, then it would be necessary to define the vertical by, for example, a gyroscopic technique.

Acknowledgements

H. E. Butler undertook the Balgust measurements in the experiments. The computer programme was written by Miss S. A. Mathews.

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Table 1

Downward momentum fluxes over 1-minute periods (dynes cm⁻²) (a) uncorrected (b) corrected for vertical sensor motion

22nd May		5th June (am)		5th June (pm)		21st Aug. (am)		21st Aug. (pm)		22nd August		27th August	
(a)	(b)	(a)	(b)	(a)	(b)	(a)	(b)	(a)	(b)	(a)	(b)	(a)	(b)
0.47	0.59	2.22	2.21	-5.61	-6.69	10.21	9.72	39.10	30.38	-0.87	-1.51	-4.33	-5.85
0.18	0.10	0.41	0.36	-1.18	-2.10	-1.67	-1.85	3.08	3.20	-0.15	-0.14	8.32	2.78
0.16	0.09	-0.69	-0.80	3.20	3.21	-1.38	-1.39	6.36	5.47	-2.35	-2.84	2.73	1.41
0.12	0.04	0.93	0.92	3.19	3.26	-4.50	-4.94	-11.73	-15.31	0.71	0.51	3.21	3.30
-0.04	-0.05	-1.52	-1.61	7.48	6.77	-3.06	-3.39	0.35	0.53	0.29	0.18	12.34	12.52
0.48	0.50	3.06	3.12	2.99	2.97	-2.19	-2.98	-2.61	-2.20	-1.38	-1.21	12.66	11.57
-0.10	-0.08	-0.13	-0.16	-4.19	-4.32	1.69	1.73	-1.02	-1.18	2.36	2.33	7.27	6.99
-0.46	-0.57	4.18	4.27	-10.67	-11.08	-2.35	-2.25	-0.40	-0.40	-0.95	-0.88	-1.81	-2.04
-0.10	-0.12	3.36	3.28	-1.75	-2.18	-3.22	-3.05	-3.47	-3.73	1.86	1.27	-3.25	-3.25
0.57	0.65	3.70	3.74	-0.92	-1.30	-16.72	-16.73	-1.34	-2.26	2.31	2.15	-4.91	-5.16
-0.01	-0.05	3.71	3.65	-1.78	-1.82	-8.55	-9.07	-0.09	-0.34	-2.31	-2.45	-4.17	-3.88
0.51	0.60	2.01	1.99	1.38	1.02	-8.21	-8.43	-4.42	-4.48	3.43	3.45	-7.58	-6.23
0.39	0.37	1.89	1.97	-2.05	-2.13	4.03	3.82	-2.68	-3.95	3.63	3.20	-4.18	-3.68
-0.11	-0.05	0.19	0.23	-4.34	-4.15	1.77	1.31	0.77	0.03	-1.43	-1.25	-1.09	-1.25
0.24	0.30	0.43	0.43	1.59	1.63	1.76	-0.31	-2.38	-2.41	3.40	3.15	6.35	5.37
0.65	0.69	0.46	0.46	-0.99	-1.08	-1.30	-1.91	5.96	5.11	0.71	0.31	4.90	3.27
0.47	0.48	0.07	0.09	-5.11	-5.51	-11.36	-13.28	-1.45	-1.54	2.07	2.08	-3.01	-2.90
0.33	0.41	3.08	3.08	-3.05	-3.11	-14.04	-18.04	-0.30	-0.06	3.17	2.68	6.76	-0.54
0.81	0.86	-0.20	-0.19	4.45	2.89	6.03	6.07	2.00	1.80	0.39	0.30	1.47	0.71
0.50	0.41	-0.66	-0.64	-2.33	-2.67	-8.98	-9.91	0.53	0.49	-0.35	-0.85	5.40	5.84
0.29	0.28	-2.23	-2.32	-1.12	-1.13	0.95	-0.78	-1.85	-3.74	-21.00	-21.68	8.30	8.05
0.89	0.95	1.84	1.82	-8.07	-7.91	-5.45	-5.23	5.47	3.76	-33.80	-33.81	-1.09	-4.70
1.78	1.83	0.83	0.72	-4.24	-4.15	-2.87	-2.72	1.19	0.84	-1.34	-1.51	5.32	5.13
0.98	0.98	0.75	0.80	0.93	0.73	-3.61	-3.81	-11.64	-11.81	-2.98	-3.71	0.51	0.00
1.05	1.05	-0.13	-0.21	0.16	0.25	-0.38	-0.67	-7.12	-7.11			-0.14	-0.52
-0.07	-0.06	-1.48	-1.28	-0.52	-0.50	-18.31	-19.27	-0.62	-1.13			0.49	0.55
		-2.07	-2.20	-0.71	-0.60	-1.55	-1.65	1.15	0.96			-1.37	-2.08
		1.92	1.84	-2.25	-2.36	-4.18	-4.44	-0.43	-0.28			2.07	2.03
				-0.29	-0.33	-0.74	-0.83	0.34	0.35			1.03	0.52
				-0.89	-0.84	6.76	6.30	-0.53	-0.34			0.55	0.61
				-1.05	-1.11	4.39	4.55	-5.29	-4.91			-0.84	-1.78
				0.35	0.24	-2.73	-3.86	-6.72	-7.65			0.62	-0.18
				-0.12	-0.22	13.46	11.98	1.21	0.76			0.03	-0.08
				-0.29	-0.32	-1.87	-1.99					-3.52	-2.26
				0.11	0.08	1.48	1.27					3.71	-2.09
				0.01	-0.01	6.49	6.26					3.24	2.58
				-0.11	-0.09	2.43	2.30					0.57	0.63
				-0.11	-0.13								
				-1.41	-1.42								
				1.98	2.16								
				2.32	1.68								
				-4.37	-4.60								

Table 2

Mean downward momentum fluxes and mean wind speeds

Date	Downward momentum fluxes (dynes cm ⁻²)		Wind Speed m sec
	uncorrected	corrected	
22nd May	0.32	0.33	1.2
5th June(am)	0.92	0.91	2.0
5th June(pm)	-0.95	-1.12	5.1
21st Aug.(am)	-1.87	-2.37	4.6
21st Aug.(pm)	0.04	-0.64	5.0
22nd Aug.	-2.53	-2.79	2.7
27th Aug.	1.35	0.62	9.8

Table 3

Mean wind speeds (m sec⁻¹) from Balthum ascents

Height (m)	22nd May		5th June			21st Aug.			22nd Aug		27th Aug	
	1200Z	1800	0600	1200	1800	0600	1200	1800	1200	1800	1200	2400
300	02	04	08	02	06	06	05	05	04	04	08	08
450	01	04	07	01	06	06	05	05	04	04	09	10
600	02	04	06	02	07	05	06	05	04	04	09	08
750	02	05	05	01	-	05	07	05	03	05	10	10
900	02	03	06	04	-	06	07	04	02	07	10	08

Table 4

Date	Mean wind directions (degrees)		
	From tower vane		From least-squares fit of double-theodolite data
22nd May	010 → 050		020 or 200
5th June(am)	300		170 or 350
5th June(pm)	260		020 or 200
21st Aug.(am)	220		020 or 200
21st Aug.(pm)	230		040 or 220
22nd Aug.	Variable		180 or 360
27th Aug.	030		030 or 210

Appendix: Experimental details

Balloon: Approximately 25m long, 90 cubic metres
BALGUST instrumentation: Inclinator, air damped, double "V" hot-wire
 Anemometer, 3 - cup contact (expts. 1-6): 3 - cup photo-electric (expts. 7-11)
Baseline: Orientation, 140°-320°
 Length, 661m (expts. 1 & 2): 735m (expts. 3-11)

Experiment Number	Date	Period (GMT)	Length of cable to sensors (m)	Mean Wind m sec ⁻¹ (From sensor data)	Mean wind at 40m (degrees, m sec ⁻¹)	Cloud	Notes
1	21-5-68	1514-1544	610	Not evaluated	010/4	3-6/8 Cu. 700m 8/8 Sc. 1200m	(i) Anemometer output intermittent due to faulty contact, output virtually unusable (ii) Zero of inclinometer displaced about 15°, but sensitivity apparently unaffected (iii) Data logger produced many punching faults.
2	22-5-68	1451-1517	610	1.2	030/1-2	3/8 Cu. 700m 4/8 Sc. 1200m 7/8 Ac. 2500m	(i) Wind fell very light at times, anemometer apparently stopped rotating on occasions, mainly in first half of run
3	5-6-68	0945-1013	610	2.0	300/2	4/8 Cu. 800m 5/8 Sc. 1100m 7/8 Ac. 2500m	(i) Wind fell almost calm for short periods a few minutes before end of run
4	5-6-68	1515-1557	610	5.1	260/5	3/8 Cu. 900m 8/8 Sc. 1400m	Nil
5	6-6-68	0910-0955	610	Not evaluated	300/6	4/8 Cu. 700m	} Overnight rain which had entered a plug and socket rendered the data from the southern theodolite useless. No analysis of data attempted
6	6-6-68	1530-1615	690	Not evaluated	270/6	4/8 Cu. 900m 5/8 Sc. 1400m	
7	21-8-68	0932-1009	533	4.6	220/3-4	4/8 Cu. 800m 5/8 Ac. 3000m	Nil
8	21-8-68	1417-1450	610	5.0	230/4	3/8 Sc. 1300m 4/8 Ac. 3500m	Nil
9	22-8-68	0915-1000	610	Not evaluated	170/3	1/8 St. 400m	(i) Convection had not built up to level of sensors: insignificant sensor motion throughout period: no data analysis attempted
10	22-8-68	1415-1440	610	2.7	Variable 1 - 5	Nil	(i) Wind fell very light on a number of occasions during the period
11	27-8-68	1338-1415	457	9.8	030/8	4-6/8 Cu. 500m	(i) Sensors set at lower level than before because of low cloud base

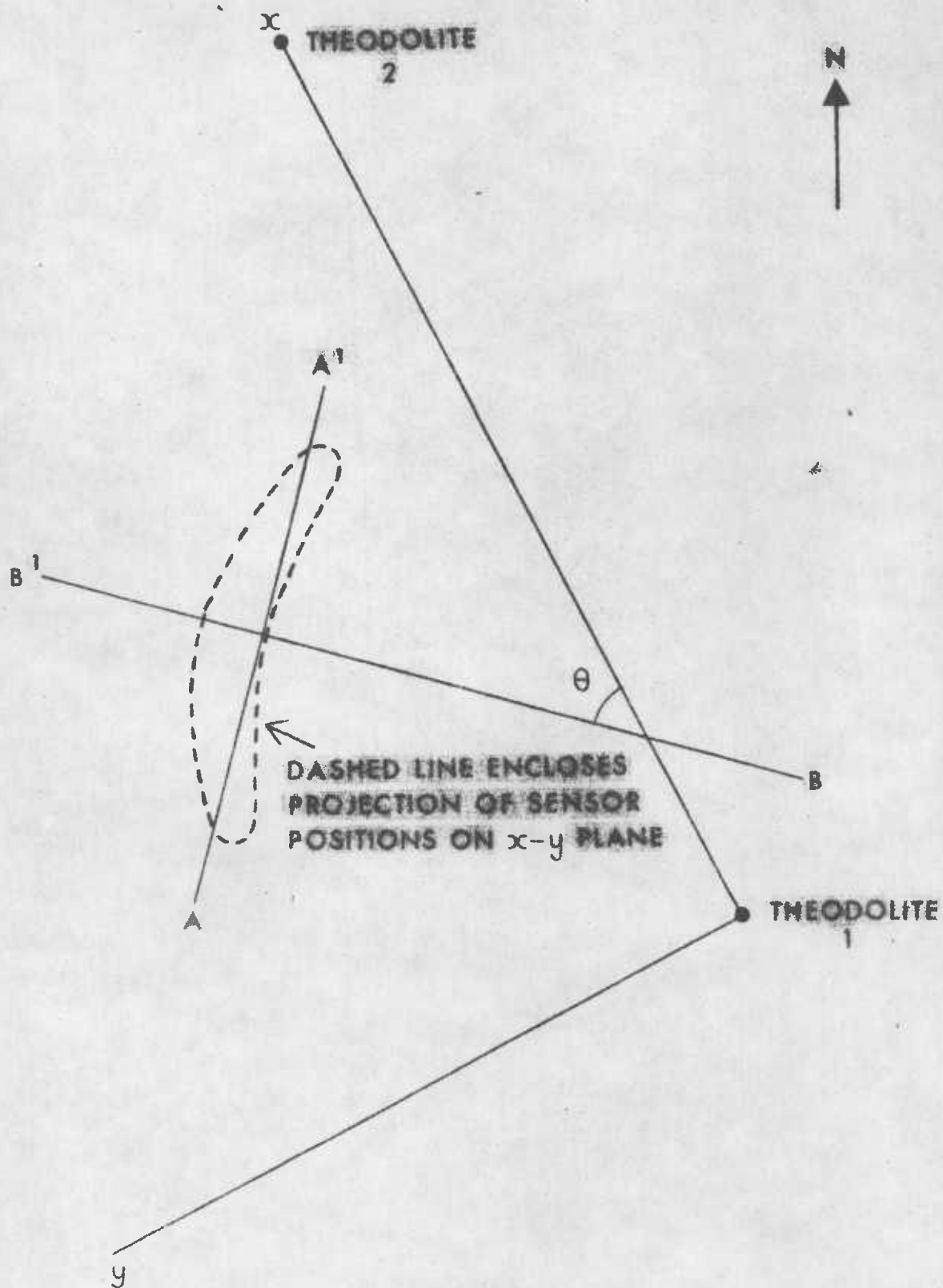


FIGURE 1