Aircraft observations of the structure of the lower boundary layer over the sea

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SUMMARY

Results from an investigation of the turbulence structure of the lower half of the convective boundary layer over the sea around the UK in a wide variety of meteorological conditions are presented. The data were obtained on eight flights made by the Hercules aircraft of the Meteorological Research Flight.

Differences in structure between boundary layers over sea and over land are emphasized, the most notable being the relatively increased importance of mechanically driven mixing over the sea. This means that a more general scaling scheme is required which retains both $u_*$ and $z_i$ as scaling parameters. Such a scheme orders the results quite effectively. Dimensionless profiles of momentum, sensible heat and latent heat fluxes are presented, together with dimensionless variance profiles. Excellent agreement is found between these and theoretical predictions from numerical models. Surface fluxes are estimated and compared with bulk aerodynamic formulae.

The turbulence statistics are shown to be in good agreement with those published by other workers.

LIST OF SYMBOLS

Basic quantities

$C_D$, $C_H$, $C_E$ transfer coefficients for momentum, sensible heat and water vapour

$c_p$ specific heat at constant pressure

$g$ acceleration due to gravity

$H$ sensible heat flux $= \bar{\rho} c_p w T$

$\mathcal{H}$ virtual heat flux $= \bar{\rho} c_p w T$

$k$ von Kármán's constant (0.4)

$L_E$ latent heat of vaporization of water

$Q$ latent heat flux $= \bar{\rho} L_E w q$

$q$ humidity mixing ratio – both total and fluctuating values

$q_s$ saturation value of $q$ at temperature $T_s$

$T$ air temperature – both total and fluctuating values

$T_s$ sea surface temperature

$(U, V, W)$ total wind components

$(u, v, w)$ fluctuating wind components

$z$ height above the sea surface

$z_i$ height of the base of the lowest inversion above the sea surface

$z_0$ roughness length

$\Delta q$ $q_{10} - q_s$

$\Delta T$ $T_{10} - T_s$

$\nu$ kinematic viscosity of air

$\rho$ density of air

$\sigma_x$ standard deviation of $x$

$\tau$ shear stress

$\theta$ potential temperature – both total and fluctuating values
Derived quantities

\[ L = -u_*^2 \left( k(g/T_v)w_{T_v0} \right)^{-1} \]

\[ q_* = \frac{wq_0}{u_*} \]

\[ T_* = \frac{wT_0}{u_*} \]

\[ u_* = (-\overline{uv})^{\frac{1}{2}} \]

\[ w_* = \left( \frac{g(T_v)}{T_v} \right) w_{T_v0}^{\frac{1}{2}} \]

\[ \theta_* = \frac{wT_0}{w_*} \]

\[ \theta_{v*} = \frac{wT_{v0}}{w_*} \]

Subscripts

0, 10 estimated surface value (at reference height 10 m) (except \( z_0 \), the roughness length)

\( \nu \) virtual quantity

An overbar denotes a mean value (flight leg average).

1. **INTRODUCTION**

Fundamental to our understanding of the convective boundary layer over the sea is a knowledge of its structure under a wide variety of meteorological situations, ranging from cloud-free suppressed conditions with a well-defined capping inversion to ensembles of developed precipitating cumulus. One could start by concentrating on the more tractable suppressed conditions before moving on to consider the more convective situations. An alternative approach, which is adopted in the present paper, is to begin by examining the lower parts of the boundary layer. Here variations arising from changes in cloud dynamics or entrainment mechanisms should be a minimum, allowing a wide variety of convective situations to be considered *en masse*. Later work will extend to the upper levels, where surface effects are no longer dominant.

In 1975, a Hercules aircraft of the Meteorological Research Flight (MRF) began a series of flights designed to investigate the structure of the lower part of the unstable atmospheric boundary layer over open sea. Results from eight of these flights are considered here. These covered a wide range of meteorological conditions, ranging from a very suppressed, cloud-free situation, to a severe gale with scattered cumulonimbus. However, at the levels flown the mean profiles were very similar, reflecting a well-mixed layer beneath a capping inversion.

After a brief outline of instrumental techniques and data analysis, the results are compared with those reported by other workers. The structure of the boundary layer over the sea is contrasted with that over land. Surface values (i.e., at 10 m) are then estimated and drag coefficients evaluated. A suitable scaling scheme is proposed and used to produce mean profiles of fluxes, variances and mean quantities. These are compared with measurements by other workers and predictions of numerical models.

2. **THE DATA**

The Hercules instrumental system is fully described in Nicholls (1978), but briefly the three wind components were measured by a combination of wind vanes, a high quality pitot-static system and an inertial platform, drifts being removed by reference to a Doppler radar and a radar altimeter. Air temperature was measured by a platinum resistance element, and humidity by a microwave refractometer referenced to an automatic dew point hygrometer (the refractometer does not function in rain, hence the omission of \( Q_0 \) in flight
BOUNDARY LAYER OVER THE SEA

H083 (see Table 1)). The sea surface temperature was derived from the output of a Barnes PRT-4 radiometer.

The data were recorded on magnetic tape and reduced to a common sampling frequency of 20 Hz, corresponding to one measurement approximately each 5 m along track. Fluctuations were calculated as differences from the mean of each run following the removal of any linear trend. The velocity components were transformed into a right-handed coordinate system with \( U \) component positive along the mean wind direction, defined on any particular day by averaging data over all the runs.

All the data considered in this paper were obtained on eight flights over the sea around the United Kingdom. Experimental areas were selected so that ‘over sea’ fetches were at least 250 km. In many instances they were considerably more. Flights were made in a wide variety of meteorological situations (see Table 1) but on only one occasion (13 Feb. 1976) did conditions change sufficiently during the flight for sub-division to be necessary. The sub-division of the flight on 10 Sept. 1975 reflects a spatial change across the chosen experimental area rather than a temporal variation.

During these flights, the aircraft executed a series of L-shaped patterns (of side length 60 km) at various heights over the same ground positions. Overall developments were monitored by occasional ascents and descents. The ‘L’s’ were orientated with their sides perpendicular and parallel to the mean wind direction measured at a height of 150 m.

In all, 74 of these 60 km runs have been analysed, 66 of which lay in the lower half of the boundary layer (as defined by reference to \( z_i \), the height of the base of the lowest inversion). All runs were below cloud base and precipitation was avoided on most occasions (see Table 1). The method used to derive the surface fluxes listed in Table 1 is described in section 5(b). In presenting these results no distinction will be made in this paper between data measured along and across wind since the runs were long enough to obtain adequate samples in both directions (see, for example, Nicholls 1978). It will be shown in a later paper that there is a marked change in the spectral distribution of energy between along wind and across wind measurements, although the total variance (or covariance) remains unaffected.

In what follows, the terms ‘surface’ layer, ‘mixed’ layer and ‘free-convection’ layer refer to those parts of the boundary layer where surface layer scaling, mixed layer scaling and free-convection scaling (see section 4), respectively, are appropriate.

3. Surface transfers

According to the bulk aerodynamic theory (e.g. Businger 1973), surface fluxes (i.e. at 10 m) may be related to mean quantities using:

\[
\begin{align*}
\tau_{\theta}^2 & = -\tau_{0}/\rho_0 = C_D \bar{U}_{10}^2 \quad \text{(momentum flux)} \\
H_{\theta}/\rho_0 & = C_H \Delta T \bar{U}_{10} \quad \text{(sensible heat flux)} \\
Q_{\theta}/\rho_0 L_E & = C_E \Delta q \bar{U}_{10} \quad \text{(latent heat flux)}.
\end{align*}
\]

Following Pennell and LeMone (1974), values of \( \bar{U}_{10} \) were obtained by extrapolation from the lowest flight level using the flux-profile relations of Dyer and Hicks (1970). This assumes that the lowest flight level is in the surface layer, which is probably a reasonable assumption given the present values of \( L \) (see Table 1). \( \Delta T \) and \( \Delta q \) were calculated from \( T_{10} \) and \( q_{10} \) (estimated by linear extrapolation) and the observed values of sea surface temperature. The measured fluxes were extrapolated to 10 m by fitting mean profiles (see section 5(b)) to the flux data for each day using a least-squares technique. Thus by plotting surface fluxes against \( \bar{U}_{10} \) etc., drag coefficients may be derived (see Fig. 1). The vertical lines in Fig. 1(a)
<table>
<thead>
<tr>
<th>Flight</th>
<th>Date</th>
<th>Flight area</th>
<th>Time of start of experiment (GMT)</th>
<th>$U_{10}$ (m s$^{-1}$)</th>
<th>Approx. mean wind direction (°)</th>
<th>$\Delta q$ (g kg$^{-1}$)</th>
<th>$z_1$ (m)</th>
<th>$H_0$ (W m$^{-2}$)</th>
<th>$Q_{0}$ (W m$^{-2}$)</th>
<th>$u_*$ (m s$^{-1}$)</th>
<th>$-L$ (m)</th>
<th>$-z_1/L$ (z = 10 m)</th>
<th>$-z/L$ (z = 10 m)</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>H076</td>
<td>26 Feb.</td>
<td>Southern North Sea</td>
<td>1100</td>
<td>11.1</td>
<td>140</td>
<td>1.0</td>
<td>1.8</td>
<td>350</td>
<td>18.9</td>
<td>40.0</td>
<td>0.41</td>
<td>280</td>
<td>1.3</td>
<td>0.036 No cloud</td>
</tr>
<tr>
<td>H083</td>
<td>2 Apr.</td>
<td>St George's Channel</td>
<td>1000</td>
<td>11.7</td>
<td>300</td>
<td>0.35</td>
<td>700</td>
<td>5.3</td>
<td>0.45</td>
<td>1300*</td>
<td>0.5</td>
<td>0.008</td>
<td>8/8 Ns. All runs in rain, *$Q_0$ assumed zero</td>
<td></td>
</tr>
<tr>
<td>H109</td>
<td>10 Sept.</td>
<td>Atlantic Ocean W of France</td>
<td>1200</td>
<td>5.6</td>
<td>285</td>
<td>1.85</td>
<td>6.4</td>
<td>600</td>
<td>18.8</td>
<td>156.0</td>
<td>0.30</td>
<td>82</td>
<td>7.3</td>
<td>0.12 4/8 Cu. Flight split due to spatial variation</td>
</tr>
<tr>
<td>J109</td>
<td></td>
<td></td>
<td>1200</td>
<td>6.8</td>
<td>285</td>
<td>1.65</td>
<td>6.0</td>
<td>600</td>
<td>18.8</td>
<td>156.0</td>
<td>0.34</td>
<td>120</td>
<td>5.0</td>
<td>0.08</td>
</tr>
<tr>
<td>H111</td>
<td>15 Sept.</td>
<td>Southern North Sea</td>
<td>1030</td>
<td>7.7</td>
<td>340</td>
<td>1.7</td>
<td>4.2</td>
<td>450</td>
<td>22.4</td>
<td>96.0</td>
<td>0.40</td>
<td>195</td>
<td>2.3</td>
<td>0.05 2/8 Cu below 8/8 Sc</td>
</tr>
<tr>
<td>H113</td>
<td>23 Sept.</td>
<td>Southern North Sea</td>
<td>0930</td>
<td>13.9</td>
<td>240</td>
<td>0.75</td>
<td>2.3</td>
<td>435</td>
<td>34.1</td>
<td>53.0</td>
<td>0.66</td>
<td>695</td>
<td>0.6</td>
<td>0.014 3/8 St Fra below 8/8 Sc</td>
</tr>
<tr>
<td>H120</td>
<td>18 Jan.</td>
<td>St George's Channel</td>
<td>1245</td>
<td>15.2</td>
<td>250</td>
<td>-0.05</td>
<td>1.65</td>
<td>1000</td>
<td>17.4</td>
<td>32.0</td>
<td>0.62</td>
<td>1090</td>
<td>0.9</td>
<td>0.009 8/8 Sc</td>
</tr>
<tr>
<td>H122</td>
<td>21 Jan.</td>
<td>Atlantic Ocean NW of Ireland</td>
<td>1030</td>
<td>21.0</td>
<td>280</td>
<td>1.65</td>
<td>3.2</td>
<td>2000</td>
<td>70.2</td>
<td>209.0</td>
<td>1.06</td>
<td>1250</td>
<td>1.6</td>
<td>0.008 Severe gale, scattered Cb</td>
</tr>
<tr>
<td>H129</td>
<td>13 Feb.</td>
<td>St George's Channel</td>
<td>1130</td>
<td>10.3</td>
<td>015</td>
<td>1.3</td>
<td>2.8</td>
<td>850</td>
<td>26.1</td>
<td>70.0</td>
<td>0.45</td>
<td>260</td>
<td>3.3</td>
<td>0.028 3/8 Cu. Flight split due to temporal variation</td>
</tr>
<tr>
<td>J129</td>
<td></td>
<td></td>
<td>1415</td>
<td>9.5</td>
<td>015</td>
<td>1.2</td>
<td>2.9</td>
<td>850</td>
<td>23.7</td>
<td>59.0</td>
<td>0.40</td>
<td>205</td>
<td>4.1</td>
<td>0.049</td>
</tr>
</tbody>
</table>
Figure 1. Bulk aerodynamic relations. The solid line in (a) represents $C_D = 1.8 \times 10^{-3}$, the dashed curve is Eq. (4) with $c = 11.0$ (Charnock 1958). The line in (b) is for $C_H = 1.6 \times 10^{-3}$ and in (c) $C_E = 1.0 \times 10^{-3}$.

represent the standard errors in $u_*$, assuming random errors in the individual flux measurements, while the horizontal lines show the standard experimental error in determining $U_{10}$. To avoid confusion, only typical errors are included in Figs. 1(b) and 1(c) (see the lower right-hand corner of each graph). Best-fit straight lines constrained to pass through the origin, which therefore correspond to a constant drag coefficient, are plotted on all the figures (solid lines).

The slope of the solid line in Fig. 1(a) corresponds to $C_D = 1.8 \times 10^{-3}$. This is higher than the generally accepted value of $1.3 \times 10^{-3}$ (see Busch 1977), though reliable estimates
vary between $1.1 \times 10^{-3}$ and $1.75 \times 10^{-3}$. $U_{10}$ and $u_*$ are unlikely to err by as much as 20% as neither the uncertainties associated with extrapolation, nor instrumental errors, are large enough. Furthermore, the cospectra (to be discussed in a later paper) reveal no spurious correlations, their shapes agreeing well with those published by other workers. However, the value of $C_D$ is sensitive to the profile shape used to estimate the 10 m fluxes (see Fig. 4(a)). If linear relations are fitted to the $\overline{wv}$ data on each day, the average value of $C_D$ would be $1.5 \times 10^{-3}$, but such a procedure is difficult to justify given the small amount of scatter about the mean profile (Fig. 4(a)) and the theoretically predicted profile shapes (section 6), which agree well with that actually used.

McBean and Paterson (1975) reported a similar discrepancy when they compared aircraft measurements of momentum flux with those determined from directly measured surface winds, the flights being over the Great Lakes. At winds speeds of $10 \text{ m s}^{-1}$, the aerodynamic flux estimates were about 40% smaller than those measured by the aircraft, if a value of $1.3 \times 10^{-3}$ was used for $C_D$. For equality, $C_D$ would have to be increased to $1.82 \times 10^{-3}$, close to the present value.

This would also accord with the work of Sethuraman and Raynor (1975). They grouped values of $C_D$ (obtained from profile measurements) according to the state of the sea surface, defined by values of the ‘roughness Reynolds number’, $= u_* z_0 / v$, and found that in a fully rough state $C_D = 1.75 \times 10^{-3}$. They also discovered that all values of $u_*$ > $0.2 \text{ m s}^{-1}$ lay in this regime. As the smallest value of $u_*$ listed in Table 1 is $0.3 \text{ m s}^{-1}$, it seems reasonable to conclude that the present data also fall into this regime. Thus a value of $1.75 \times 10^{-3}$ would be expected, which is close to the value $C_D = 1.80 \times 10^{-3}$ shown in Fig. 1(a). Their results imply that $C_D$ increases with increasing windspeed. Many previous authors have discussed this possibility but no clear conclusions have emerged as the measurements are very difficult to make, especially in strong winds. Most workers appear to favour an increase in $C_D$ with increasing windspeed (e.g. Sheppard et al. 1972; Sethuraman and Raynor 1975; Smith and Banke 1975) but no quantitative relations have been agreed, though several have been proposed. Charnock (1955) suggested $z_0 = \beta u_*^2 / g$ and hence

$$U / u_* = (1/\kappa) \log (gz / u_*^2) + c$$

(1)

for neutral conditions. This relation, which implies an increasing $C_D$ with windspeed, is shown in Fig. 1(a). The data presented here, from occasions biased towards higher-wind situations, do suggest a mean value of $C_D$ somewhat in excess of the ‘normal’ value (based on a lower average windspeed) but individual $C_D$ measurements are too uncertain to attempt to fit a specific relationship with windspeed. More detailed conclusions must await further measurements.

The value of $C_H = 1.6 \times 10^{-3}$, represented by the solid line in Fig. 1(b), is in good agreement with previously reported values as summarized by Busch (1977) and by Freihe and Schmidt (1976).

The latent heat data (Fig. 1(c)) fit a $C_E$ of $1.0 \times 10^{-3}$, which is slightly lower than the value of $1.32 \times 10^{-3}$ proposed by Freihe and Schmidt, and the value of $1.2 \times 10^{-3}$ proposed by Smith (1974). This may reflect the shape of the mean $wq$ profile with the low-level maximum (see section 5).

The values of both $C_H$ and $C_E$ could depend on the way in which $\Delta T$ and $\Delta q$ are measured since both depend on $T_s$ ($\Delta q$ to a lesser extent than $\Delta T$). Here $T_s$ was measured using a radiometer whereas most experimenters have to rely on a direct method, which samples the temperature some distance below the surface. In general, there will be a difference between the two (see Grassl 1976), but the radiation temperature should be the more relevant provided appropriate corrections are made (see Nicholls 1978).
4. Scaling

Before proceeding to consider vertical profiles, the data will be non-dimensionalized so that universal curves can be discussed. Two scaling schemes current in the literature will be mentioned: 'surface layer' similarity and 'mixed layer' similarity (see Kaimal et al. 1976). The former assumes that the parameters governing the turbulence structure are $\tau_0$, $z$, $g/T_v$, $\mathcal{H}_0$, $\rho_0$ and $c_p$. Temperature and velocity scales may be formed from these, giving

$$u_* = (-\tau_0/\rho_0)^{1/3} \text{ and } T_{c*} = \mathcal{H}_0/u_*\rho_0c_p.$$  

Parameters made dimensionless by $u_*$ and $T_{c*}$ should then be functions of $z/L$ only, where $L = -u_*^3\{k(g/T_v)(\mathcal{H}_0/\rho_0c_p)\}^{-1}$ is the Monin-Obukhov length, as this is the only dimensionless group that can be formed from these parameters.

Mixed layer similarity is applied to situations where the turbulence structure is no longer influenced by $\tau_0$ but does vary with $z_0$, the corresponding scaling parameters being

$$w_* = \{(g/T_v)(\mathcal{H}_0/\rho_0c_p)z_0\}^{1/3} \text{ and } \theta_{v*} = (\mathcal{H}_0/\rho_0c_p)(1/w_*).$$

Parameters scaled by $w_*$ and $\theta_{v*}$ should then be functions of $z/z_0$ only.

Although virtual temperature is used in the formulae above, explicit temperature and humidity scales (e.g. $T_v$ or $q_*$) can also be derived (see Kaimal et al. 1976).

Surface layer scaling should be valid only up to heights at which $\tau_0$ is still an important parameter ($z < -L$) while mixed layer scaling can apply only when the converse is true. Recent measurements in strongly convective conditions (Kaimal et al. 1976) imply that Monin-Obukhov scaling applies only if $z < 0.1z_0$, with mixed layer scaling valid above. This idealized structure is summarized in Fig. 2. In the strongly unstable conditions typical of clear days over land ($L$ is small, $z_0$ large and $-z_0/L$ may be of the order of 100). There are then two distinct layers (see Fig. 2) in which surface and mixed layer scaling schemes may be applied independently. In between there may be a 'matching' layer where both types of scaling apply, the layer of 'free convection' (e.g. Wyngaard 1973; Panofsky 1976), where the turbulence is no longer sensitive to $z_0$ or $L$.

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**Figure 2.** Schematic diagram showing idealized limits of validity of certain scaling techniques (see text for descriptions). For example, in order of increasing height from the surface in very unstable conditions (left to right along the line $-z_0/L = 100$) one would encounter the surface layer, a layer of free convection, followed by the mixed layer.
All the present data were collected in conditions where \(-z_i/L < 10\), a common occurrence over the sea. Thus there is a considerable depth of boundary layer where neither mixed layer nor surface layer scaling is adequate (region 'A' in Fig. 2), where a more general approach is required which retains both \(\tau_0\) and \(z_i\). Physically, this means that the effects of surface-induced mechanical mixing cannot be neglected even though \(z_i/z_1\) is quite large. This is not generally the case over land, where surface heat fluxes are much larger.

Thus it is logical to use \(T_*\) and \(u_*\) (defined for the surface layer) to scale the present data, though this means there are now two independent non-dimensional groups, \(z/z_1\) and \(z_i/L\). Here the data are grouped according to \(z_i/L\) and plotted against \(z/z_1\), as \(z/z_1\) is easily interpreted as a physical height and \(z_i/L\) as a bulk stability parameter governing the overall convective structure of the boundary layer. Deardorff (1972) adopted the same approach in presenting his slightly unstable data.

5. THE VERTICAL PROFILES

In order to investigate possible changes in vertical profiles with changing stability, the data have been divided into two classes, A and B, according to the value of \(z_i/L\). The value of \(z_i/L = -1.5\), chosen to divide the classes, is not in any sense a critical value, but serves to group the data to enable the overall changes with stability to be observed despite the scatter of individual data points. This also allows direct comparison with numerical models (see section 6). However, the actual grouping was, to some extent, influenced by the results of such models, which suggested that significant changes in certain parameters would be observed over such a range of \(z_i/L\). With this in mind, class A is defined as the more unstable category, where \(z_i/L < -1.5\) (average value \(-3.9\)) and class B the less unstable, for which \(z_i/L > -1.5\) (average value \(-0.9\)). The values of \(z_i/L\) for each flight are listed in Table 1. Class A contains forty-six runs and class B twenty-eight.

(a) The mean quantities

Fig. 3 shows the mean profiles with the velocities scaled using \(u_*\) and the runs classified according to stability using the classes defined above. Error bars represent standard errors in the means.

Wind profiles are much as expected, with the more unstable data (for which mixing should be more vigorous) exhibiting a smaller momentum deficit and a deeper region of near-constant \(U\). \(V\) does not appear to separate according to stability, showing a consistent behaviour when averaged over all flights. If extrapolated to the surface the intercept implies a veer of approximately \(4^\circ\) between \(0.5z_i\) and the surface (\(V\) is not expected to be zero with the present coordinate system, see section 2). An anticlockwise rotation of \(2^\circ\) would make both \(V\) and \(\bar{w}\) zero at the surface.

The lower parts of the boundary layer were obviously well mixed for, despite quite large variations in \(\Delta T\) (see Table 1), the temperature lapse rate is very close to the dry adiabatic value for both classes (i.e. \(\bar{\theta} - \bar{\theta}_{10} \approx 0\)). This may be contrasted with the \(\bar{q} - \bar{q}_{10}\) profile, which exhibits a marked linear trend with height, independent of stability or \(\Delta T\), probably reflecting the presence of a sink of water vapour aloft. Without such a sink, \(\bar{q} - \bar{q}_{10}\) would tend to zero provided the boundary layer was well mixed. Both profiles can be explained in terms of a well-mixed layer eroding a relatively dry overhead inversion which would act as a sink of water vapour while serving as a source of sensible heat. Thus provided the atmosphere was well mixed one would expect \(\bar{\theta} - \bar{\theta}_{10} \approx 0\) and a linear decrease in \(\bar{q} - \bar{q}_{10}\) as water vapour is transferred from the surface to the inversion. The rate of decrease will depend on the magnitude of \(\bar{w}\), the efficiency of the entrainment process, and the
dryness of the air above the mixed layer. These profiles suggest that the conditions are fairly close to equilibrium, otherwise the data would be more scattered and the trends less regular.

Similar profiles including the linear decrease in humidity have been observed by other workers (e.g. Pennell and LeMone 1974; Donelan and Miyake 1973). The former also reported a low-level wind maximum within the mixed layer which they argued was caused by baroclinicity. The profiles shown in Fig. 3 suggest that the present data are unlikely to have been unduly influenced by this effect.

(b) **The fluxes**

Flux measurements were also split into the two classes, A and B, mean values for each class being plotted in Fig. 4. However, as there does not appear to be any significant
separation according to stability, mean profile curves (the full lines in Fig. 4) were fitted to the data as a whole, ignoring stability classifications. The shapes of the curves were determined empirically and fitted to the data points shown in the figure by a least-squares technique. (These curves were also fitted to the data obtained on individual flights to define the surface fluxes and scaling parameters listed in Table 1.)

Both $\overline{\nu\nu}$ and $\overline{\nu T}$ show steady decreases with height but $\overline{\nu\nu}$ and $\overline{\nu q}$ have maxima near $0.2z_1$. $\overline{\nu\nu}_0$ can be brought to zero by rotating the axes by $2^\circ$ in an anticlockwise direction (see above), but this does not change the shape of the profiles significantly. The curvatures of the $\overline{\nu\nu}$ and $\overline{\nu T}$ profiles are greatest near the ground, $\overline{\nu\nu}$ decreasing by about 25% and $\overline{\nu T}$ by about 30% in the first 0.1$z_1$.

Although lack of data leaves the form of the profiles at higher levels open to question, the results are clearly not incompatible with negative values of $\overline{\nu T}$ and positive values of $\overline{\nu q}$ at heights near $z_1$, where entrainment would be expected.

Figure 4. Dimensionless flux profiles. Classes as Fig. 3.
The $\bar{w}$ profile is of particular interest since there appears to be a maximum near $0.2z_i$. This would indicate a slow drying of the lowest layers associated with a moistening of the upper levels. This is compatible with the observed decrease in $\bar{q}$ throughout the mixed layer (see Fig. 3) as it suggests the mean profile is being slowly altered by turbulent mixing to one which is more constant with height. It seems unlikely that the fluxes could have been systematically under-estimated by 10–20% near the surface as, although the frequency response of the refractometer falls off above about 1 Hz, the cospectra (to be published later) do not indicate losses of this magnitude.

It has also been suggested by previous authors (e.g. Donelan and Miyake 1973) that such behaviour could be caused by the evaporation of spray droplets, which would produce a local source of water vapour. Extensive white capping was observed on all flights but no data are available to check this hypothesis.

(c) The variances

The behaviour of the variances can now be discussed using the scaling scheme outlined in section 4. The scaling parameters are calculated using the surface estimates listed in Table 1. Fig. 5 shows the variations of $\sigma_w/\mu_*$, $\sigma_v/\mu_*$ and $\sigma_u/\mu_*$ with height, retaining the usual stability classes. Again the data collapse quite well onto distinct lines. Also plotted are mean surface layer measurements compiled from previously published over-sea data (sources: Davidson 1974; Donelan and Miyake 1973; Elder 1973; Miyake et al. 1970; Pond et al. 1971; Smith 1974; Wieringa 1972; Leavitt and Paulson 1975). The scatter in these estimates is considerable, but there is little indication of any significant variation with stability in the range $0 \leq -z/L \leq 1.0$.

Within the mixed layer, the profiles of the vertical component show the greatest overall change with stability, values of $\sigma_w/\mu_*$ increasing with instability above 0.15$z_i$. Presumably this reflects the dominance of mechanical mixing below this level with convective activity becoming increasingly important above. Class A data exhibit a clear maximum just below 0.5$z_i$. A similar feature is apparent in the more neutral data, but just above 0.1$z_i$ and less marked. Maximal values of $\sigma_w$ have also been observed between 0.4$z_i$ and 0.5$z_i$ by Willis and Deardorff (laboratory simulation, 1974) and Nicholls and LeMone (private communication), based on GATE data obtained in moderately convective conditions for which $10 \leq |z_i/L| \leq 100$. Over land a clearly discernible maximum is also often observed between 0.5$z_i$ and 0.6$z_i$ in unstable conditions (Caughey, private communication) where $|z_i/L| \approx 100$. Thus a picture emerges of a maximum of $\sigma_w/\mu_*$ becoming more prominent and whose height above the surface increases as instability increases, as was originally suggested by the numerical experiments of Deardorff (1972, see section 6 below). However, more data are required to quantify this behaviour accurately.

There is still considerable discussion about the value of $\sigma_w/\mu_*$ in the surface layer, with reported 'best' estimates ranging from 1.05 (Lumley and Panofsky 1964) to 1.47 (Miyake et al. 1970). The data presented here suggest a value, over the sea, of $1.1 \pm 0.1$, which is slightly lower than the generally accepted value of 1.3 for over-land measurements (e.g. Panofsky et al. 1977). This is consistent with an increase in the correlation coefficient $-r_{uw}$ over the sea. The average near-surface value of $-r_{uw}$ was found to be $+35$ in the present data, compared with the maximum value of $+30$ observed over land by Haugen et al. (1971). This partly reflects the overall stability range of present measurements, being much closer to neutral than those reported by Haugen ($-r_{uw}$ is observed to decrease with increasing instability, e.g. Deardorff (1973)), but also may be related to the increased horizontal homogeneity of the state of the sea surface compared with a land site.

The lateral component $\sigma_v/\mu_*$ is strongly dependent on the value of $z_i/L$. In the more
unstable case $\sigma_u/u_*$ is considerably less than $\sigma_v/u_*$ especially at small $z/z_1$. In a later paper it will be shown that this reflects an increase in the spectral energy of the $v$ component at long wavelengths with increasing instability, possibly associated with return flows induced by convection. This concurs with the observations of Kaimal (1978) that at longer wavelengths, horizontal velocity spectra scale with $z_1$ even within the surface layer.

Panofsky et al. (1977) have suggested that within the surface layer

$$\sigma_u/u_* = \sigma_v/u_* = f(z_1/L) = (12 - 0.5z_1/L)^{4}. \quad (2)$$

Although this is clearly not the case in class B, where $\sigma_v < \sigma_u$, Eq. (2) is a reasonable approxi-
mation to the class A results (for which the mean value of \( z_i/L = -3.9 \)) where \( \sigma_u/\ell_* \approx \sigma_z/\ell_* \approx 2.5 \) at low levels. Equation (2) predicts a value of 2.43. Generally we must conclude that this model is valid only when \(-z_i/L\) is greater than some critical value. As values of \(-z_i/L\) are generally quite small over the sea, a more sophisticated model is needed to describe the marine boundary layer.

Similar results have been reported by Pennell and LeMone (1974), who studied the structure of a fair-weather boundary layer over the sea for which \(-z_i/L \approx 1.5\). This lies between the average values of the two classes considered here, being slightly nearer the less unstable one (class B). Their profile follows the same trend, lying between the pairs of curves shown in Fig. 5, with a slight bias towards the more neutral case.

Temperature fluctuations are also strongly dependent on stability, the two \( \sigma_T/T_* \) lines being clearly displaced, with magnitudes decreasing with increasing convective activity (see Fig. 6). This decrease does, however, owe more to variation of \( T_* \) than to any change in \( \sigma_T \), which remains fairly constant between the two classes. Similar comments apply to \( \sigma_q/q_* \) (also shown in Fig. 6), although in this case the separation of the two curves reflects changes in \( \sigma_q \) rather than in \( q_* \). Previous surface layer measurements have also shown that both quantities tend to increase rapidly as neutral conditions are approached. The points included in Figs. 6(a) and (b) are thus comparable with class A data only (sources: Leavitt and Paulson 1975; Phelps and Pond 1971: values taken at \(-z/L \approx 0.1\)).

6. Comparisons with Numerical Models

Unfortunately, with the exception of near-surface data, very few experimental results are available to compare with the measurements presented here. However, there have been several relevant numerical simulations, notably those of Deardorff (1972) and Wyngaard et al. (1974), and it is instructive to compare these theoretical results with measurements made with the Hercules.

There are many reasons why experimental results should not be expected to produce exactly the same characteristics as those derived from small-scale models, since the latter
consider only a limited number of physical processes (see, for example, the discussion by Sommeria and LeMone 1978). Experimental data may include mesoscale variability or entrainment processes not considered by models. However, since the main vertical turbulent transports generally occur on scales common to both model and experimental data, a good degree of correspondence should be observed. The overall behaviour should remain similar even though actual numerical values may vary. Model results generally refer to horizontally homogeneous, steady-state conditions. However, as the data selected for this paper appear to have been drawn from situations where these two conditions were satisfied, comparisons should be valid.

Fig. 7 shows the mean wind profiles (from Fig. 3) compared with those predicted by Wyngaard et al. The coordinate system has been rotated by 2° to make the two sets of data compatible, \( V_0 \) going to zero (see section 5 above). The results of Deardorff are not included here as the actual values of \( \overline{U}/u_\ast \) depend upon the lower boundary conditions chosen in the model. Those used by Wyngaard et al. are fairly compatible with the measurements. The results from both models show the same profile behaviour, illustrating that the turning of the wind with height decreases with increasing instability. This is also reflected in the momentum flux profiles shown in Fig. 8. Any change in these profiles due to the variation in stability from class A to class B is undetectable in the data although the overall shape of the mean profile curve is in excellent agreement with the models. The models show clearly the change from the strongly curved, near-neutral profiles (\( -z_i/L \) small) to the more linear profiles as \( -z_i/L \) increases. There are significant differences between the predictions from the two models, but these are of the same order as the scatter in the data. Neither considers entrainment across the upper boundary.

Fig. 9 shows further comparisons between the virtual heat flux profile and the heat flux profile of Deardorff (this model did not explicitly consider the effects of water vapour or entrainment). Again there is excellent agreement.

Vertical velocity variance comparisons are shown in Fig. 10. These show reasonable
Figure 8. Comparisons with models – dimensionless momentum flux profiles. Curves shown are from Wyngaard et al. (1974) and Deardorff (1972). The curves are labelled with values of $z_i/L$. Also shown in (a) is the mean profile obtained from the data presented here (from Fig. 4, mean $z_i/L = -2.4$).
agreement. As expected, $\sigma_w/u_*$ increases above 0.1$z_i$ as conditions become more unstable. Below 0.1$z_i$, Deardorff's curves depend strongly on the sub-grid-scale assumptions made. Wyngaard et al. chose to keep the surface value constant at 1.3, therefore close agreement at low levels cannot be expected.
7. CONCLUDING REMARKS

In demonstrating the close agreement between the present results and those published by other workers, this paper has highlighted differences between marine and over-land boundary layers, notably the increased importance of mechanically, relative to buoyancy, induced mixing in the former. This is due to the reduced surface buoyancy flux generally found in the marine boundary layer compared with that typically observed over land, and is reflected in the values of $z_L/L$ observed in these data, which are typically at least one order of magnitude less than those encountered in convective boundary layers over land. A scaling scheme which retains both $u_*$ and $z_l$ is required to describe the data in contrast to simpler schemes which are applicable either to surface or convectively mixed layers only. Dimensionless profiles have been derived which demonstrate the effectiveness of this more general method of scaling, which should be applicable to a wide range of convective boundary layers over the sea.

The aircraft results presented here have covered a larger height range than most previous measurements over the sea. This enables profiles of the turbulent fluxes of momentum, sensible and latent heat to be obtained as functions of non-dimensional height and stability. These are shown to agree very well with the predictions of turbulence simulation models. By extrapolating these flux profiles, surface fluxes may be estimated. A comparison of these values with corresponding estimates derived from the bulk aerodynamic method shows reasonable agreement, although a relatively high value for the (10 m) drag coefficient is suggested. This is shown to be in agreement with previously reported work in similar conditions, providing some evidence for an increasing drag coefficient with increasing windspeed.

A complementary investigation into the spectral and cospectral properties of the data presented here is nearing completion and will be reported in a later paper. Flights are also being made through the depth of the boundary layer, designed to increase our knowledge of mixing processes at heights beyond those reported here.

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