Aircraft observations of the Ekman layer during the Joint Air–Sea Interaction Experiment

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SUMMARY

Results are presented from an observational study of the mid-latitude atmospheric boundary layer over the sea. The data were obtained mainly by instrumented aircraft as part of the Joint Air–Sea Interaction Experiment (JASIN).

The eight occasions chosen for study are characterized by small surface buoyancy fluxes and the conditions are well described as near neutral and barotropic. In the absence of low-level inversions, a well-mixed Ekman layer is observed on each occasion but is limited to a depth of approximately 0-2u/a. Accurate measurements of horizontal gradients (including pressure) together with improved wind observations and turbulent flux measurements enabled each of the terms of the momentum balance to be evaluated throughout the depth of this layer. These terms were found to balance quite closely, were well described by Ekman scaling and were consistent with the requirements of the measured turbulent kinetic energy balance. The latter suggests that the main shear production terms are dissipated locally and that little is exported to upper levels in the Ekman layer to enable significant deepening by entrainment although only a slight increase in instability appears to be needed to alter this balance. Conditions in which such boundary layer structure might be observed are suggested. Values of various coefficients used in schemes for relating surface fluxes to mean quantities (C_D, C_E, C_G, A, B) are derived and compared with previous measurements.

Spectral analysis reveals that most of the turbulent transport is confined to a distinct high wavenumber region whose characteristics vary as a function of Ekman layer depth and stability parameters. This is superimposed on larger-scale fluctuations which do not vary appreciably within the Ekman layer and which therefore dominate the variances in the upper regions as the intensity of the smaller-scale turbulence decreases strongly with height.

Finally, further implications of this interpretation are discussed with particular reference to the heat and water vapour balance. These imply that removal of water vapour from the Ekman layer is accomplished by transfer related to cloud activity, marking a significant change in the mechanism of turbulent transport at the top of this layer. The relationship between these and similar results obtained concurrently by different methods is also discussed.

1. INTRODUCTION

This paper is concerned with the structure of the atmospheric boundary layer (ABL) at a mid-latitude, open ocean site and is based upon results obtained during the Joint Air–Sea Interaction experiment (JASIN) which took place during the summer of 1978 in the NE Atlantic to the north-west of Scotland. As a result of combination of factors, this type of ABL has different characteristics from those which have received most previous observational attention. For example, the physical nature of the underlying water surface and the small departures from neutral stability which are a feature of such open ocean sites result in a very different structure from that observed at other mid-latitude locations. Also, the features of the large-scale flows associated with mid-latitude westerlies give rise to a boundary layer structure which is significantly different from that found over oceans at lower latitudes.

The data on which this paper is based were obtained primarily by aircraft: the C130 of the Meteorological Research Flight, Farnborough, and the NCAR Electra of the National Center for Atmospheric Research, Boulder, U.S.A., supplemented by surface observations from four ships. (A full description of the complete experiment is contained in Royal Society, 1979.) The data therefore extend throughout a considerable depth of the boundary layer in contrast to most previous observations which have tended to concentrate on the few tens of metres immediately adjacent to the surface. Those
previous results which have extended to higher levels also fall into one of two categories depending upon the type of measurement strategy employed: those based upon direct, *in situ* measurements of turbulent fluctuations and those relying on mean flow methods.

Mean flow methods are usually based on profile measurements made by frequent radiosonde launches from a fixed array of stations. These enable budgets of various quantities to be estimated for the area in question. The turbulent transport may then be estimated from the budget residuals if the fluxes are known at some level. In practice, however, the accuracy requirements for determining the fluxes to even a reasonable standard are stringent and additional constraints are usually necessary to produce meaningful results even with the highest quality data. Another disadvantage is that statistically reliable results require a very long averaging period even with frequent radiosonde launches. This can often be a number of days which has the effect of smoothing out real changes occurring on shorter timescales. Thus the interpretation of the results is made more difficult since the final quantities are integrated over quite large space- and timescales and a diagnostic model is often required to relate these to smaller-scale transporting elements e.g. Esbensen (1978). An associated limitation is the possibility that spatial inhomogeneities may be misinterpreted as small-scale transport since the variables are assumed to vary linearly between observing stations. The method is thus best suited to very uniform areas, e.g. the trade wind regions.

Stress profiles have been calculated by this method in the trade wind boundary layer by Brümer *et al.* (1974), Charnock *et al.* (1956) and Holland and Rasmussen (1973). In each case the alongwind stress component was assumed zero at the level of maximum wind although the authors found it necessary to adjust the geostrophic wind profiles in both the latter two papers to yield reasonable stress estimates. Holland and Rasmussen and Augstein *et al.* (1973) also obtained values for the heat and water vapour fluxes which can be derived by analogous methods but require additional information about cloud, radiative heat transfer and precipitation. Clarke (1970) and Clarke and Hess (1973, 1974) deduced boundary layer flux profiles by similar methods at land sites in Australia; however, difficulties were again experienced in determining the geostrophic and thermal winds to sufficient accuracy (Hess *et al.* 1981). Similar problems were also encountered by Taylor *et al.* (1983) who made measurements across the 200 km sided triangle of the JASIN area. These results are discussed later. Despite the known shortcomings of this method, such studies have supplied much of the available information on the structure of the boundary layer throughout its whole depth.

Direct measurements of turbulent fluctuations in the ABL over the sea are still scarce above the surface layer. Thompson (1972) employed a ship-borne tethered balloon system although ship and balloon motion placed considerable limitations on the data. Nicholls and Readings (1979) used aircraft data to make measurements of turbulence throughout a substantial part of the ABL in a narrow band of stability close to neutral over the sea around the U.K. These observations demonstrated the importance of mechanically generated turbulence in maintaining mixing in this type of boundary layer although the limited accuracy of the wind measurements then available together with the limited sampling capability of a single aircraft precluded any attempt to investigate the heat or momentum balance. A spectral analysis of part of this data showed excellent agreement with previous surface layer measurements in samples taken parallel to the wind, but showed that acrosswind (i.e. across shear) measurements were significantly different (Nicholls and Readings 1981). Lenschow *et al.* (1980) again used aircraft data to investigate more convective boundary layers in cold air outbreaks over the ocean, but an attempt to close the momentum budget was not particularly successful due to the limited geostrophic wind information available, the neglect of horizontal acceleration
terms and the limited accuracy of the wind measurements. Budgets of a number of second-order moments were, however, more successful.

A number of investigations have also concentrated on the interaction of the ABL with clouds, usually in the tropics. Here even small amounts of cumulus appear to strongly affect the heat and water vapour fluxes near cloud base (e.g. Pennell and LeMone 1974; LeMone and Pennell 1976; Nicholls and LeMone 1980) although the effect on the momentum fluxes is less clear.

Many of these results have been generalized by the use of similarity theory which can provide an elegant, universal description as well as a framework for the analysis of data and comparison of results. However, these remain valid only under certain restricted conditions and observations are needed to test their utility and to establish numerical values. Previous results have covered only a limited range of conditions and in many cases measurements of important parameters have not been available. Some tests have therefore often been rather inconclusive.

The JASIN aircraft programme was designed to avoid many of the shortcomings of earlier investigations and the choice of measurement strategy together with new methods of analysis has enabled both the mean flow and the turbulent fluctuations to be sampled simultaneously and with sufficient accuracy to enable the variation with height of the balance between the turbulent transport and the mean flow to be measured directly in a consistent manner. Also, since the aircraft made direct measurements of turbulent fluctuations, it is possible to make inferences about the physical nature of the energy transfer processes.

The paper is organized as follows. First, details of the measurement programme and analysis methods are presented followed by a brief section concerned with instrumentation. The majority of the results are presented in the subsequent three sections followed by discussion. A list of notation appears at appendix II. To avoid undue repetition, reference is often made to other JASIN results which have already appeared in Nicholls et al. (1983a), henceforward called RS.

2. THE MEASUREMENT PROGRAMME

Flight patterns were designed to enable both horizontal and vertical gradients to be determined accurately. To achieve this, both aircraft were operated in loose formation wherever possible, flying around a square box pattern of side length l at a range of altitudes from 30 m upwards as described in RS.

The equation expressing one component of the horizontal mean motion may be expressed using usual notation (see appendix) as

$$\frac{\partial U}{\partial t} + U \frac{\partial U}{\partial x} + V \frac{\partial U}{\partial y} + W \frac{\partial U}{\partial z} = -\frac{1}{\rho} \frac{\partial p}{\partial x} + fV - \frac{\partial u'w'}{\partial z}$$

If each of these terms is considered as an average over the area enclosed by the flight pattern (i.e. $L^2$) and over a corresponding length of time ($-\ell |\textbf{V}|$) during which the measurements are made, it is possible to estimate each term, using the continuity equation to recover $W$. The heat and water balance may be investigated by a directly analogous procedure. Therefore, fluxes and other statistical moments were computed from the fluctuation data along each of the sides of the boxes and by averaging measurements made around the circumference, quantities representative of the area enclosed were obtained. Horizontal gradients were determined by a method explained below and vertical gradients were measured by repeating box patterns at a series of altitudes.
The method is limited to fairly steady conditions where departures from linearity across the experimental area remain small and the size of $l$ is a compromise between various constraints:

(i) A longer sample reduces the statistical uncertainty implicit in correlation measurements in stationary conditions.

(ii) In order to maintain reasonably linear variations across the area, $l$ must be considerably less than the scales of synoptic features, which are typically a few hundred kilometres or more.

(iii) The maximum vertical resolution is achieved with the minimum box size since measurements at a greater number of levels can be made in the time available. The endurance of the aircraft set an upper limit of about 6 h in the experimental area.

A value of $l = 70$ km was eventually chosen. This would be expected to give an uncertainty of about $\pm 25\%$ in a single $\overline{u'w'}$ measurement (Nicholls et al. 1983b) and resulted in a complete box taking approximately one hour to complete at each level. This satisfies the above constraints and allows a useful number of levels to be flown. The terms of Eq. (1) are therefore considered to be averaged over an area of approximately $(70\text{km})^2$ and a timescale of a few hours.

However, in spite of the use of two aircraft, in certain situations the vertical resolution of the measurements was still not felt to be adequate, so L-patterns with 50 km side were interspersed between the boxes. Also, since it was unlikely that sufficient levels could be repeated to give satisfactory local rate of change terms, the box patterns were usually chosen (when conditions permitted) to be centred on a ship making meteorological measurements with the expectation that these terms could be estimated from the ship data. A number of slow rate of climb ascents and descents were also regularly interspersed between the straight and level parts of the box and L-patterns. The detailed vertical structure revealed in these soundings would then be used to assess the gross characteristics of the boundary layer and to locate significant levels. This information was also used in choosing levels for the horizontal flight legs since the soundings could be partially plotted in real time aboard the aircraft.

Data from eight occasions were selected for analysis on the basis of the following criteria:

(i) At least one box pattern was flown.

(ii) Conditions were reasonably uniform across the area selected, not affected by frontal activity.

(iii) The air–sea temperature difference was not such as to suppress completely turbulent mixing below the lowest flight levels.

(iv) Sufficient levels were sampled close to the surface.

Further details are contained in Table 1.

### 3. Instrumentation and Data Processing

The instrumentation on the C130 and the Electra has been discussed at length in many previous publications and a full summary referring to the configurations current at the time of this particular experiment is contained in RS. One significant new feature was the introduction of methods to improve the accuracy of horizontal wind measurements. This was achieved by continuously recording accurate position fix information (from Decca Navigator and LORAN-C) on the C130 which subsequently allowed the low frequency velocity errors inherent in inertial platforms, commonly referred to as 'Schuler' errors, to be accurately determined and hence corrected. As the Electra generally flew in formation with the C130, a similar procedure was adopted using the
### TABLE 1. DETAILS OF CASES CHOSEN FOR STUDY

<table>
<thead>
<tr>
<th>Date (1978)</th>
<th>Synoptic type</th>
<th>Mean wind direction at 50 m</th>
<th>Patterns flown*</th>
<th>Aircraft data**</th>
<th>Start GMT</th>
<th>End GMT</th>
</tr>
</thead>
<tbody>
<tr>
<td>26 Jul.</td>
<td>Cyclonic SW'ly</td>
<td>176°</td>
<td>2</td>
<td>C</td>
<td>1200</td>
<td>1600</td>
</tr>
<tr>
<td>28 Jul.</td>
<td>Cyclonic SW'ly</td>
<td>208°</td>
<td>4</td>
<td>C, E</td>
<td>1000</td>
<td>1600</td>
</tr>
<tr>
<td>29 Jul.</td>
<td>Cyclonic SW'ly</td>
<td>203°</td>
<td>4</td>
<td>C, E</td>
<td>1030</td>
<td>1600</td>
</tr>
<tr>
<td>7 Aug.</td>
<td>N'ly</td>
<td>005°</td>
<td>5</td>
<td>C, E</td>
<td>1110</td>
<td>1720</td>
</tr>
<tr>
<td>21 Aug.</td>
<td>Cyclonic SW'ly</td>
<td>234°</td>
<td>4</td>
<td>C, E</td>
<td>1150</td>
<td>1550</td>
</tr>
<tr>
<td>23 Aug.</td>
<td>Anticyclonic W'ly</td>
<td>251°</td>
<td>2</td>
<td>C</td>
<td>1030</td>
<td>1720</td>
</tr>
<tr>
<td>31 Aug.</td>
<td>Anticyclonic W'ly</td>
<td>311°</td>
<td>2</td>
<td>C</td>
<td>0910</td>
<td>1620</td>
</tr>
</tbody>
</table>

* Only those within the main experimental area are included
** C = MRF C130, E = NCAR Electra

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Cloud observations

<table>
<thead>
<tr>
<th>Date</th>
<th>Observations</th>
</tr>
</thead>
<tbody>
<tr>
<td>26 Jul.</td>
<td>Broken St. Sc; base ~200 m</td>
</tr>
<tr>
<td>28 Jul.</td>
<td>3/8 Cu; base ~250 m</td>
</tr>
<tr>
<td>29 Jul.</td>
<td>3/8 Cu; base ~300 m</td>
</tr>
<tr>
<td>7 Aug.</td>
<td>4/8 Cu beneath 8/8 Sc; bases 350 m/900 m</td>
</tr>
<tr>
<td>21 Aug.</td>
<td>7/8 St and Sc; bases variable</td>
</tr>
<tr>
<td>23 Aug.</td>
<td>Broken St and Sc; bases ~300 m/850 m</td>
</tr>
<tr>
<td>25 Aug.</td>
<td>3/8 Cu and Sc; base ~250 m</td>
</tr>
<tr>
<td>31 Aug.</td>
<td>8/8 Sc; base ~900 m</td>
</tr>
</tbody>
</table>

C130 as a reference. These enabled errors of the order of ±2 m s\(^{-1}\) to be removed from both sets of data resulting in an overall accuracy in the horizontal wind components of about ±0.5 m s\(^{-1}\). This has been confirmed by intercomparisons (Nicholls et al. 1983b). Full details concerning this procedure are contained in Nicholls (1983). The wind measurements are therefore sufficiently accurate for the ageostrophic components of the momentum balance to be measured directly given that the pressure gradient can be determined with similar accuracy.

Considerable efforts were made to intercompare data from the two, independent aircraft systems (Nicholls et al. 1983b). To briefly summarize the conclusions, time-averaged (over 5 s) data could all be reduced to a single common standard while the high frequency (20 s\(^{-1}\)) measurements showed detectable but negligible small differences. Data from both aircraft may therefore be freely combined and no distinction between the two is made in what follows.

A fixed horizontal coordinate system was used for each day where the u component was aligned with the measured mean wind at a height of 50 m (as listed in Table 1). Statistical moments were calculated as described in Nicholls (1978).

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### 4. MEAN HORIZONTAL STRUCTURE AND SURFACE FLUXES

In all the cases selected for study, a well-mixed layer was observed to extend upwards from the surface. The depth of this layer and the associated vertical structure are addressed in the next section, here the emphasis is placed on horizontal variations within it.
### TABLE 2(a). HORIZONTAL GRADIENT MEASUREMENTS

<table>
<thead>
<tr>
<th>Date</th>
<th>DDD (deg)</th>
<th>FF (ms⁻¹)</th>
<th>( \delta p/\delta E ) (mb/1000 km)</th>
<th>( \delta p/\delta A ) (mb)</th>
<th>( V_i ) (m s⁻¹)</th>
<th>DDD (deg)</th>
<th>FF (ms⁻¹)</th>
<th>( \delta U(N)/\delta E ) (m s⁻¹)/100 km</th>
<th>( \delta U(N)/\delta A ) (m s⁻¹)</th>
<th>( \delta V(E)/\delta E ) (m s⁻¹)/100 km</th>
<th>( \delta V(E)/\delta A ) (m s⁻¹)</th>
<th>div (10⁻² s⁻¹)</th>
<th>( \alpha_0 ) (deg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>26 Jul.</td>
<td>176</td>
<td>10.4</td>
<td>-3.8(0.8)</td>
<td>21.9(0.7)</td>
<td>190</td>
<td>14.2</td>
<td>0.0-1(0.5)</td>
<td>0.0-1(0.5)</td>
<td>1.4(0.9)</td>
<td>1.4(0.9)</td>
<td>-1.4(0.9)</td>
<td>14</td>
<td></td>
</tr>
<tr>
<td>28 Jul.</td>
<td>208</td>
<td>6.9</td>
<td>-7.9(0.4)</td>
<td>10.8(0.5)</td>
<td>216</td>
<td>8.6</td>
<td>0.0-1(0.5)</td>
<td>0.0-1(0.5)</td>
<td>0.9(0.6)</td>
<td>0.9(0.6)</td>
<td>-1.7(1.0)</td>
<td>8</td>
<td></td>
</tr>
<tr>
<td>29 Jul.</td>
<td>203</td>
<td>8.5</td>
<td>-9.8(0.9)</td>
<td>11.9(0.9)</td>
<td>220</td>
<td>9.9</td>
<td>0.0-1(0.5)</td>
<td>0.0-1(0.5)</td>
<td>0.9(0.6)</td>
<td>0.9(0.6)</td>
<td>-1.7(1.0)</td>
<td>8</td>
<td></td>
</tr>
<tr>
<td>7 Aug.</td>
<td>200</td>
<td>6.8</td>
<td>-19.0(0.5)</td>
<td>-12.1(0.5)</td>
<td>200</td>
<td>7.8</td>
<td>0.0-1(0.5)</td>
<td>0.0-1(0.5)</td>
<td>0.9(0.6)</td>
<td>0.9(0.6)</td>
<td>-1.7(1.0)</td>
<td>8</td>
<td></td>
</tr>
<tr>
<td>21 Aug.</td>
<td>234</td>
<td>11.4</td>
<td>-21.4(0.4)</td>
<td>7.6(0.5)</td>
<td>250</td>
<td>14.5</td>
<td>0.0-1(0.5)</td>
<td>0.0-1(0.5)</td>
<td>0.9(0.6)</td>
<td>0.9(0.6)</td>
<td>-1.7(1.0)</td>
<td>8</td>
<td></td>
</tr>
<tr>
<td>23 Aug.</td>
<td>251</td>
<td>11.5</td>
<td>-20.4(1.2)</td>
<td>-2.9(1.5)</td>
<td>262</td>
<td>13.2</td>
<td>0.0-1(0.5)</td>
<td>0.0-1(0.5)</td>
<td>0.9(0.6)</td>
<td>0.9(0.6)</td>
<td>-1.7(1.0)</td>
<td>8</td>
<td></td>
</tr>
<tr>
<td>25 Aug.</td>
<td>273</td>
<td>6.4</td>
<td>-12.4(1.1)</td>
<td>-2.7(0.8)</td>
<td>282</td>
<td>8.1</td>
<td>0.0-1(0.5)</td>
<td>0.0-1(0.5)</td>
<td>0.9(0.6)</td>
<td>0.9(0.6)</td>
<td>-1.7(1.0)</td>
<td>8</td>
<td></td>
</tr>
<tr>
<td>31 Aug.</td>
<td>311</td>
<td>10.2</td>
<td>-10.9(0.8)</td>
<td>-12.8(0.7)</td>
<td>320</td>
<td>10.8</td>
<td>0.0-1(0.5)</td>
<td>0.0-1(0.5)</td>
<td>0.9(0.6)</td>
<td>0.9(0.6)</td>
<td>-1.7(1.0)</td>
<td>8</td>
<td></td>
</tr>
</tbody>
</table>

Standard errors in parenthesis

### TABLE 2(b). HORIZONTAL GRADIENT MEASUREMENTS

<table>
<thead>
<tr>
<th>Date</th>
<th>DDD (deg)</th>
<th>FF (ms⁻¹)</th>
<th>( \delta T/\delta E ) (°C/100 km)</th>
<th>( \delta T/\delta A ) (°C)</th>
<th>( \delta Q/\delta N ) (g kg⁻¹/100 km)</th>
<th>( \delta Q/\delta A ) (g kg⁻¹)</th>
<th>( V_{H,T} ) (m s⁻¹)</th>
<th>( V_{H,T} ) (°C/100 km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>26 Jul.</td>
<td>176</td>
<td>10.4</td>
<td>0.08(0.13)</td>
<td>-0.07(0.13)</td>
<td>-0.30(0.10)</td>
<td>-0.22(0.10)</td>
<td>105</td>
<td>0.11</td>
</tr>
<tr>
<td>28 Jul.</td>
<td>208</td>
<td>6.9</td>
<td>0.02(0.06)</td>
<td>-0.01(0.06)</td>
<td>0.44(0.08)</td>
<td>-0.15(0.08)</td>
<td>169</td>
<td>0.10</td>
</tr>
<tr>
<td>29 Jul.</td>
<td>203</td>
<td>8.5</td>
<td>0.05(0.10)</td>
<td>-0.06(0.10)</td>
<td>0.37(0.20)</td>
<td>-0.52(0.20)</td>
<td>277</td>
<td>0.08</td>
</tr>
<tr>
<td>7 Aug.</td>
<td>200</td>
<td>6.8</td>
<td>-0.70(0.14)</td>
<td>-0.10(0.06)</td>
<td>0.16(0.08)</td>
<td>-0.10(0.07)</td>
<td>353</td>
<td>0.67</td>
</tr>
<tr>
<td>21 Aug.</td>
<td>234</td>
<td>11.4</td>
<td>-0.02(0.14)</td>
<td>0.28(0.14)</td>
<td>0.33(0.05)</td>
<td>-0.42(0.05)</td>
<td>335</td>
<td>0.51</td>
</tr>
<tr>
<td>23 Aug.</td>
<td>251</td>
<td>11.5</td>
<td>-0.26(0.17)</td>
<td>0.13(0.18)</td>
<td>0.42(0.10)</td>
<td>-0.26(0.11)</td>
<td>337</td>
<td>0.21</td>
</tr>
<tr>
<td>25 Aug.</td>
<td>273</td>
<td>6.4</td>
<td>0.07(0.06)</td>
<td>0.14(0.04)</td>
<td>0.37(0.12)</td>
<td>0.05(0.09)</td>
<td>229</td>
<td>0.20</td>
</tr>
<tr>
<td>31 Aug.</td>
<td>311</td>
<td>10.2</td>
<td>-0.11(0.11)</td>
<td>0.10(0.11)</td>
<td>0.26(0.16)</td>
<td>0.03(0.16)</td>
<td>299</td>
<td>0.13</td>
</tr>
</tbody>
</table>

Standard errors in parenthesis
(a) The measurement of horizontal gradients

The method used to estimate the horizontal gradients of wind, temperature, humidity and pressure in keeping with the averaging scheme outlined above is described in appendix I. The scheme seeks to define the value of a particular horizontal gradient representative of the area enclosed by the box pattern for the duration of the experimental part of a flight and makes use of the excellent differential accuracy available on individual measurement runs. The required spatial gradients are objectively determined by a least squares method, so full account is taken of all the measurements taken along each 70 km flight track. As shown in appendix I, this technique offers one of the most sensitive methods of determining the geostrophic wind over the sea since comparable pressure difference measurements from surface-based instrumentation can barely attain this standard of accuracy even in optimum conditions.

The horizontal gradient measurements obtained on each of the days analysed by this method are listed in Tables 2(a) and 2(b). Due to the slightly different correction methods applied to the Electra wind data, only C130 data were utilized when defining the velocity gradients. The technique also enables an assessment of the statistical reliability of the results to be made (the figures in brackets show standard errors).

From the results listed in these tables, it is immediately clear that the measured 50m and geostrophic winds are consistent, with $\alpha_g$ (the difference between their directions) consistently having the expected sign, suggesting that useful momentum budget studies are possible.

The velocity gradients are relatively more uncertain, though small, but do still account reasonably well for the observed variation across the boxes. The uncertainty in the horizontal divergence calculated from those values shown in Table 2(a) is approximately $\pm 10^{-5} \text{s}^{-1}$, the same as that derived by Nicholls (1983) for equivalent line integral methods.

The values of the measured horizontal virtual temperature gradients (which may be calculated from the data in the tables) show that there were only two occasions (7 and 21 Aug.) with any appreciable low-level baroclinicity, the larger of the two (7 Aug.) having a virtual temperature gradient of 0.7 K/100 km corresponding to a geostrophic wind shear of $2 \times 10^{-3} \text{s}^{-1}$. Only on 7 Aug. was there any appreciable temperature advection (cold advection, about $5 \times 10^{-5} \text{Ks}^{-1}$), values on all other days being less than $1.5 \times 10^{-5} \text{Ks}^{-1}$ (disregarding sign).

A number of comparisons were made between the gradients listed in the tables and corresponding estimates for the whole JASIN ship triangle determined from hourly surface observations. On some occasions (e.g. 7 Aug.) the agreement was good, but generally significant differences were found suggesting that the gradients changed in the remainder of the area not entered by the aircraft or that gradients within the triangle were not well determined by the observations at the vertices or both.

(b) Surface fluxes

One of the drawbacks of airborne measurements is the minimum altitude at which observations can be made (30 m a.s.l.) while the surface forcing is best described by surface (nominally 10 m) observations. This necessitates an extrapolation procedure which in turn requires some knowledge of the flux gradients between 10 m and the lowest flight level. However, in conditions where a mixed layer extends in depth to a few hundred metres, the corrections required are small fractions of the values of the lowest measurements and can be defined in a number of ways. Two possibilities are:

(i) By curve fitting and extrapolation.
(ii) By estimating the flux gradients between the lowest measurement level and 10 m using the budget equations (e.g. Eq. (1) for the momentum flux).

While there is no physical reasoning employed in method (i), other than by noting that excessively large flux gradients cannot be maintained, thus allowing a reasonably smooth curve to be fitted, method (ii) incurs large errors in individual cases. This is indeed one reason for evaluating the surface fluxes since the results may then be scaled and combined together to form a composite.

In practice, a combination of both methods was used. Flux gradients were initially estimated by fitting first- and second-order polynomials and also by fitting curves by eye to all the relevant data obtained on a particular day. The corrections derived from these different curves generally varied by very little. The surface momentum flux estimates were later updated in a second iteration by assuming a simplified momentum budget. It is shown below that a sufficiently accurate approximation to the flux gradients between the lowest flight level and 10 m is specified by

\[
\frac{1}{u_*} \frac{\partial (u'w')}{\partial z} = \frac{1}{u_*} (V - V_{\text{g}}) = 10
\]

(2)

giving a slightly different correction. All the resultant, second-iteration, \(u_*\) values were found to differ by less than 5% from the initial estimates. The box-average surface fluxes were then derived from these and are shown in Table 3 together with quantities calculated from them.

The surface virtual heat or buoyancy fluxes are small but positive in each case except 26 Jul. Data from the latter case are used only in section 6. In more stable situations characterized by larger negative buoyancy fluxes, mixing is suppressed at low levels and large gradients may exist in layers close to the surface, which are not accessible to aircraft. (Under these conditions stress measured at levels that are accessible to aircraft may be strongly modulated if the surface buoyancy flux changes sign owing, for example, to varying air or sea surface temperatures.)

While the sensible heat fluxes are small, the water vapour fluxes are reasonably large, being responsible for 20–50% of the virtual heat flux and actually changing the sign on 28 Jul.

\[(c)\] Transfer coefficient analysis

These allow the fluxes and mean quantities to be tested for compatibility and are often used to specify the value of the fluxes from measurements of simpler quantities made near the air–sea interface. These relationships are usually written:

\[
-(u'w')_{10} = C_D U_{10} \]

(3)

\[
(w'T')_{10} = C_H U_{10} (T_s - T_{10}) \]

(4)

\[
(w'q')_{10} = C_E U_{10} (Q_s - Q_{10}) \]

(5)

where 10 refers to values at a reference 10 m height. In order to compare the present results with previous measurements it is therefore again necessary to invoke an extrapolation procedure.

For this analysis, only data from runs at or less than 50 m a.s.l. have been used and in order to account for changes across the box pattern due, for example, to the horizontal gradients listed in the preceding section, the results are taken singly, run by run, instead of forming an area average. Values of \(U_{50}, T_{50}\) and \(Q_{50}\) were derived from run-average values at this flight level. By assuming a logarithmic profile and the stability corrections of Dyer and Hicks (1970) the measurements were reduced to a reference height of 10 m.
## TABLE 3. Surface fluxes and associated parameters

<table>
<thead>
<tr>
<th>Date</th>
<th>$-(u'w')_0$ $(\text{m}^2\text{s}^{-3}) \times 10$</th>
<th>$(w'T)_b$ $(\text{m}^{-1}\text{s}^{-3}) \times 10^3$</th>
<th>$(w'q)_b$ $(\text{m}^{-1}\text{s}^{-3}) \times 10^3$</th>
<th>$(w'T)_b$ $(\text{m}^{-1}\text{s}^{-2} \text{K}) \times 10^3$</th>
<th>$-L$ (m)</th>
<th>$h$ (m)</th>
<th>$hf/u_*$ $\frac{1}{L}$</th>
<th>$-u_*$ $\frac{1}{L}$</th>
<th>Symbol</th>
</tr>
</thead>
<tbody>
<tr>
<td>26 Jul.</td>
<td>0.75</td>
<td>-3.0</td>
<td>2.5</td>
<td>-2.6</td>
<td>-550</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>A</td>
</tr>
<tr>
<td>28 Jul.</td>
<td>0.36</td>
<td>-0.5</td>
<td>7.0</td>
<td>0.7</td>
<td>710</td>
<td>240</td>
<td>0.16</td>
<td>2.1</td>
<td>0.3</td>
</tr>
<tr>
<td>29 Jul.</td>
<td>0.72</td>
<td>2.5</td>
<td>12.0</td>
<td>4.6</td>
<td>310</td>
<td>340</td>
<td>0.17</td>
<td>6.9</td>
<td>B</td>
</tr>
<tr>
<td>7 Aug.</td>
<td>0.49</td>
<td>12.0</td>
<td>14.0</td>
<td>14.4</td>
<td>54</td>
<td>410</td>
<td>0.26</td>
<td>7.6</td>
<td>C</td>
</tr>
<tr>
<td>21 Aug.</td>
<td>1.30</td>
<td>2.0</td>
<td>13.0</td>
<td>4.3</td>
<td>790</td>
<td>450</td>
<td>0.16</td>
<td>3.6</td>
<td>D</td>
</tr>
<tr>
<td>23 Aug.</td>
<td>1.28</td>
<td>5.0</td>
<td>15.0</td>
<td>7.6</td>
<td>450</td>
<td>500</td>
<td>0.18</td>
<td>6.3</td>
<td>E</td>
</tr>
<tr>
<td>25 Aug.</td>
<td>0.32</td>
<td>2.5</td>
<td>5.0</td>
<td>3.4</td>
<td>120</td>
<td>210</td>
<td>0.16</td>
<td>12</td>
<td>F</td>
</tr>
<tr>
<td>31 Aug.</td>
<td>1.15</td>
<td>6.0</td>
<td>27.0</td>
<td>10.7</td>
<td>270</td>
<td>660</td>
<td>0.24</td>
<td>10</td>
<td>G</td>
</tr>
</tbody>
</table>

## TABLE 4. Recent evaluations of $C_D$ over the oceans

<table>
<thead>
<tr>
<th>Source</th>
<th>Method</th>
<th>Range of wind speeds</th>
<th>Corresponding range of $10^6C_D$ *</th>
</tr>
</thead>
<tbody>
<tr>
<td>Smith and Banke (1975)</td>
<td>Eddy correlation</td>
<td>5 m/s to 11 m/s</td>
<td>0.96 to 1.36</td>
</tr>
<tr>
<td>Garratt (1977)</td>
<td>Review</td>
<td>5 m/s to 11 m/s</td>
<td>1.08 to 1.49</td>
</tr>
<tr>
<td>Amoroco and DeVries (1980)</td>
<td>Review</td>
<td>5 m/s to 11 m/s</td>
<td>1.05 to 1.45</td>
</tr>
<tr>
<td>Large and Pond (1981)</td>
<td>Dissipation (JASIN)</td>
<td>&lt; 11 m/s</td>
<td>1.2</td>
</tr>
<tr>
<td>Present measurements</td>
<td>Aircraft eddy correlation</td>
<td>5 m/s to 11 m/s</td>
<td>1.29 ± 0.08</td>
</tr>
</tbody>
</table>

* At neutral stability and at $z = 10$ m

## TABLE 5. Measurements of $C_s$ over the sea

### JASIN flights

<table>
<thead>
<tr>
<th>Date</th>
<th>$10^2 C_s$</th>
<th>Previous results</th>
<th>Site</th>
<th>Stratification</th>
<th>$10^2 C_s$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Date</td>
<td></td>
<td>Reference</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>28 Jul.</td>
<td>2.2</td>
<td>Deacon (1973)</td>
<td>N. Atlantic</td>
<td>Nearly neutral</td>
<td>2.7</td>
</tr>
<tr>
<td>29 Jul.</td>
<td>2.7</td>
<td>Hasse and Dunckel (1974)</td>
<td>W. Baltic</td>
<td>Neutral-unstable</td>
<td>2.2</td>
</tr>
<tr>
<td>7 Aug.</td>
<td>2.9</td>
<td>Lettau (1958)</td>
<td>Scilly Isles</td>
<td>Unstable</td>
<td>2.2</td>
</tr>
<tr>
<td>21 Aug.</td>
<td>2.5</td>
<td>Lettau and Hоеber (1964)</td>
<td>Heligoland</td>
<td>Unstable</td>
<td>2.0</td>
</tr>
<tr>
<td>23 Aug.</td>
<td>2.7</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>25 Aug.</td>
<td>2.3</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>31 Aug.</td>
<td>3.2</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mean ($\sigma_s$)</td>
<td>2.6(0.1)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
using the appropriate values of $u_*$, $T_*$ and $q_*$ deduced from Table 3. This method of correction is valid in the surface layer only, but the degree of uncertainty introduced is expected to be reasonably small given the near-neutral conditions and the depth of the mixed layers (see below). The actual corrections required are quite small: $-1.5 \text{ m s}^{-1}$ for wind speed, negligible for temperature and $\sim 0.2 \text{ g kg}^{-1}$ for specific humidity. These are roughly the same as the variations observed across the box patterns. The fluxes were extrapolated to the $10 \text{ m}$ reference level using the vertical gradients determined in section 4(b) above.

The momentum flux results (corresponding to Eq. (3)) are shown in Fig. 1. A linear regression fits the data quite well, yielding a value of $C_D = 1.29 (\pm 0.08) \times 10^{-3}$ and an intercept which is close to the origin.

![Figure 1. $-\langle u'w' \rangle_{10}$ as a function of $U_{10}^2$ from individual runs. Letter code as Table 3.](image)

Much previous effort has been expended in defining values of $C_D$ which may depend upon stability, wind speed, wave effects and possibly other factors. A consensus has not yet been reached on the effects of the latter two of these but present opinion is that $C_D$ increases with wind speed although no quantitative details have yet been agreed (e.g. Smith and Banke 1975; Garratt 1977; Amorcho and DeVries 1980; Large and Pond 1981). The value of $C_D$ obtained above is in very good agreement with these results if account is taken of the nearly neutral, open sea conditions in which the present measure-
ments were made and the range of wind speeds encountered (<11 m s\(^{-1}\) at 10 m). Some comparisons are given in Table 4 for a comparable range of wind speeds.

Since the geostrophic wind speed, \(G\), was also measured, the geostrophic drag coefficient, \(C_g = u_s/G\), was also evaluated. The Ekman layer ‘resistance laws’ (e.g. Tennekes 1973) suggest that \(C_g\) should be a function of the surface Rossby number, \(Ro = u_s/f_0\), and a stability parameter, but that in marine boundary layers of the type studied here, this variation of \(C_g\) should be very weak. Accordingly, Table 5 lists the determinations of \(C_g\) on each day together with previous results obtained in similar conditions.

No determination of \(C_H\) was possible because of the small air–sea temperature differences and the correspondingly small fluxes; however, it was possible to estimate \(C_E\). This is made more difficult because undetermined offsets were known to be present in the dew-point and sea surface temperature measurements (Nicholls et al. 1983b) although the intercomparisons suggest that these remained constant while the data were obtained.

Assuming that the true \((Q^T)\) and measured \((Q^M)\) values are related by

\[
Q^T = Q^M + Q_1 \quad \text{and} \quad Q_{10}^T = Q_{10}^M + Q_2
\]  

(6)

where \(Q_1\) and \(Q_2\) are constants, Eq. (5) becomes

\[
(w'q')_{10}/U_{10} = C_E(Q^M_{10} - Q_{10}^M) + C_E(Q_1 - Q_2).
\]

(7)

Most of the variation of \((Q^M_{10} - Q_{10}^M)\) is due to \(Q_{10}^M\) as the sea surface temperature (s.s.t.) changed by very little. It is known that the s.s.t. measured by an airborne radiation thermometer will be systematically different to the ‘bulk’ or immersion temperature as measured from a surface platform due to non-blackness or ‘cool-skin’ effects (see e.g. Nicholls 1978). The measured s.s.t. is corrected for neither of these but the total effect is believed to be small (<0.5 K) for the conditions encountered here, which is confirmed by intercomparisons with bulk measurements from ships. The effect of disregarding these corrections is therefore to introduce a constant, or at most slowly varying, systematic error into \(Q^M\). This is considered to be part of the offset \(Q_1\).

Figure 2 therefore shows \((w'q')_{10}/U_{10}\) plotted against \((Q^M_{10} - Q_{10}^M)\) and from the gradient of the regression line, \(C_E = (0.9 \pm 0.1)\times10^{-3}\). This is slightly smaller than most previous determinations: 1.3\times10^{-3} by Friese and Schmitt (1976), 1.2\times10^{-3} by Smith (1974) and (1.15 \pm 0.2)\times10^{-3} by Large and Pond (1980). The latter was also determined during JASIN by the dissipation method. The value is, however, in closer agreement with the value of 1.0\times10^{-3} found by Nicholls and Readings (1979) using similar data and techniques. The reasons for measuring these slightly lower values are not clear. The results of the intercomparisons provide no evidence supporting any systematic over-evaluation of \((Q^M_{10} - Q_{10}^M)\) other than a constant offset, while the humidity flux measurements have compared well in previous experiments (e.g. GATE: Barnes et al. 1980; LeMone and Pennell 1980). The wind speed also appears to have been well determined although the extrapolation procedures used, though well tested in other experiments, cannot be examined more closely here in the absence of very-low-level profile data. The stability-related profile corrections of Dyer and Hicks (1970) used here to relate \(Q_{10}\) to \(Q_{50}\) amount to only 5–10% of \((Q^M_{10} - Q_{10}^M)\) so substantial errors in this procedure would be necessary to significantly affect \(C_E\). However, convincing observational confirmation of these corrections for humidity profiles in conditions such as those encountered here is still outstanding.
One final point which may be deduced from Figs. 1 and 2 is that the fluxes measured around the circumference of the box patterns are strongly correlated with small changes in the local mean conditions. This demonstrates not only the sensitivity of the measurement techniques to quite small changes, but also shows that the low-level mixing processes respond quickly to changes in external forcing.

Figure 2. $10^3 \frac{w'q'}{U_{10}}$ as a function of $(Q_{10}^M - Q_{10}^B)$. Letter code as Table 3.

5. VERTICAL STRUCTURE

A typical group of vertical profile observations is shown in Fig. 3. Due mainly to the small variation in stability, most other profiles measured in the JASIN flights were similar. These have been discussed in RS, so only a brief resume is given here. In increasing height order from the surface, the common features of the profiles were found to be:

(i) A well-mixed layer extending to a height $h$.
(ii) A layer slightly stable to dry convection where $\theta_v$ increased slowly with height.
(iii) A stratiform cloud layer beneath the main inversion.

Under these conditions, the level $h$ did not generally correspond with marked changes of any of the quantities measured (e.g. see Fig. 3) making the determination of $h$ from profile data very uncertain. However, the flux data provided a much clearer definition since the $u'w'$ component of the momentum flux was found to decrease approximately linearly with height on each of the days analysed. The height at which a linear regression to these data become zero served to define $h$ which was taken as the definition of the depth of this mixed layer. Subsequently, small discontinuities in the mean profiles could often be detected at about this same level, as can also be seen in Fig. 3. The other intercept provided a first estimate to the area-averaged stress, as mentioned in section 4(b) above.
Figure 3. Profiles measured on descent by the C130 at 1115 GMT on 7 August. The dashed line on the $q_1$ trace represents an adiabatic increase assuming cloud base is at 750 m.

Values of $h$ and $L$, the Monin–Obukhov length scale, are listed in Table 3. Since $L$ can be roughly interpreted as the height at which buoyant production of turbulent kinetic energy (TKE) exceeds shear production, only one occasion, 7 August, was relatively unstable ($-h/L = 8$). Table 3 also shows that $h$ is approximately proportional to $u_*/f$, which is a prediction of neutral Ekman layer similarity theory (see for example Zilitinkevich 1975) with a constant of proportionality of about 0.2. This value is close to those generally quoted for neutral stability determined from previous observational studies (0.25–0.3, see reviews by McBean 1979; Tennekes 1973). To avoid any possible confusion with current terminology where the term 'mixed layer' has come to imply a predominantly convectively mixed, inversion-capped layer, the region extending to $z = h$ is subsequently referred to as the Ekman layer and the scaled height $z f / u_*$ is denoted $\tilde{z}$.

The aircraft spent most of the time within the Ekman layer since $h$ was perceived in flight as the level at which turbulent mixing ceased. Most of the results therefore refer to this region. However, mixing was also occurring within the cloud layer as illustrated by the near-adiabatic increase of the specific liquid water content, $q_1$, in Fig. 3. This is promoted by the destabilizing influence of radiative effects (e.g. Slingo et al. 1982). However, coupling between the cloud and Ekman layers, if present at all, was observed to be weak and intermittent: both aircraft and the tethered balloon observations discussed in RS showed the intervening region to be essentially non-turbulent unless associated with cumulus convection occasionally seen rising from within the Ekman layer.

(a) The momentum balance

Within the Ekman layer the momentum balance equations may be non-dimensionalized using $u_*/f$ as the height scale to become

$$\partial (u'w'/u_*)/\partial (zf/u_*) = (1/u_*) (V - V')$$

(8)

$$\partial (v'w'/u_*)/\partial (zf/u_*) = (1/u_*) (U' - U)$$

(9)
where the virtual geostrophic wind \((U'_g, V'_g)\) is defined by
\[
(U'_g, V'_g) = \left\{ -\frac{1}{f} \left\{ \frac{1}{\rho} \frac{\partial p}{\partial y} + \frac{\partial V}{\partial t} \right\}, \frac{1}{f} \left\{ \frac{1}{\rho} \frac{\partial p}{\partial x} - \frac{\partial U}{\partial t} \right\} \right\}.
\] (10)

Each of the terms in Eqs. (8)–(10) were evaluated run by run. The geostrophic wind was calculated for each using the run-averaged value of density and the appropriate pressure gradient listed in Table 2(a). This was assumed to apply at the mean height of the observations used to compute the gradient, i.e. usually midway through the mixed layer. Small corrections were made to the geostrophic winds at other levels on 7 and 21 August to account for the non-negligible thermal wind using data from Table 2(b). (Corrections were <0.1 m s\(^{-1}\) on other days.)

Some approximation is necessary when deriving the acceleration terms but since these are usually smaller than the main terms in the momentum balance, a degree of uncertainty can be tolerated. The components of Eq. (10) were evaluated as follows:

(i) The local rate of change was obtained by a linear regression on surface observations over a period extending from two hours before the start of the first aircraft run to two hours after the aircraft departed. On the occasions when the box pattern was not centred on a ship, a weighted average of the rate of change measured by the closest ships was used.

(ii) The mixed layer average horizontal velocity gradients and the average wind components \(U\) and \(V\) determined at a height of 50 m (see Table 2(a)) were used at all levels. The vertical variation of the wind components can be neglected in comparison to the uncertainties present in the horizontal gradient measurements.

(iii) The term involving \(W\) was neglected. It can be shown (Nicholls 1983) that estimates of \(W\) derived from the continuity equation and velocity gradient measurements together with the observed vertical wind-shear combine to yield a term which is at most ~1.5\(u_*f\). This figure is an upper bound based on the assumption that the divergence is constant with height at a value of 1.5 \(\times 10^{-5}\) s\(^{-1}\) (cf. Table 2(b)) and applies at the top of the mixed layer where the implied vertical velocity is greatest.

This method of evaluating the acceleration terms tends to be biased towards the surface, but was assumed to hold throughout the mixed layer. The acceleration corrections therefore remain constant on each day and are listed in Table 6.

<table>
<thead>
<tr>
<th>Date</th>
<th>(DU/\text{Dt} \times 10^5) (m s(^{-2}))</th>
<th>(DV/\text{Dt} \times 10^5) (m s(^{-2}))</th>
<th>((1/u_*f)) (DU/\text{Dt})</th>
<th>((1/u_*f)DV/\text{Dt})</th>
</tr>
</thead>
<tbody>
<tr>
<td>28 Jul.</td>
<td>1</td>
<td>4</td>
<td>0.3</td>
<td>1.8</td>
</tr>
<tr>
<td>29 Jul.</td>
<td>4</td>
<td>13</td>
<td>1.2</td>
<td>3.7</td>
</tr>
<tr>
<td>7 Aug.</td>
<td>-6</td>
<td>3</td>
<td>-2.1</td>
<td>-1.3</td>
</tr>
<tr>
<td>21 Aug.</td>
<td>-7</td>
<td>2</td>
<td>-1.6</td>
<td>0.4</td>
</tr>
<tr>
<td>23 Aug.</td>
<td>9</td>
<td>-1</td>
<td>2.0</td>
<td>-0.1</td>
</tr>
<tr>
<td>25 Aug.</td>
<td>-9</td>
<td>-6</td>
<td>-4.2</td>
<td>-2.7</td>
</tr>
<tr>
<td>31 Aug.</td>
<td>-9</td>
<td>-0</td>
<td>-2.2</td>
<td>-0.1</td>
</tr>
</tbody>
</table>

The errors involved in this procedure are difficult to specify but a reasonable estimate might be \(\pm 5 \times 10^{-5}\) m s\(^{-2}\). If this is combined with the accuracies of the actual and geostrophic wind terms discussed earlier, the uncertainty in a single non-dimensional
geostrophic departure term averaged around one complete box pattern is estimated to be about ±3. This appears to be consistent with the scatter of the measurements discussed below.

The momentum flux and velocity defect profiles are shown in Fig. 4. The symbols represent the average from each particular box pattern, but since these only make up 70% of the total available data, the results as a whole have also been grouped into non-dimensional height classes as shown in the figure. The momentum flux profiles are well defined with little scatter, suggesting they are not significantly affected by the small changes in stability between the different occasions.

The balance between the momentum fluxes and velocity defects has already been discussed to some extent in RS and is shown by the dashed curves. Those drawn on the momentum flux profiles have been obtained by integrating the curves drawn through the velocity defect data upwards from the surface. Those drawn on the velocity defect profiles were defined by linear regression on the \((V - V_g)/u_*\) data, but the scatter of the other component is larger and a better constraint was provided through the TKE balance discussed below. Within the limits set by experimental error, the balance shown in Fig. 4 is quite good at all levels although the decrease in \(\bar{u}'\bar{w}'/u_*\) observed in the upper half of the Ekman layer would require negative values of \((U_g' - U)\) if Eq. (9) was to balance exactly. However, the winds at this level were not observed to be consistently super-geostrophic, suggesting that the flow aloft must have been accelerating relative to that at lower levels (recall that the determination of the acceleration terms was biased towards the surface). This probably reflects the different times taken by the flux profiles to reach equilibrium with the velocity profiles \((-h/u_* = 0.2/f)\), while the velocity profiles themselves require on the order of \(2\pi/f (\sim 14\, \text{h} \text{ at } 60^\circ \text{N})\) to approach a steady state, a factor 10\(\pi\) longer.

The non-dimensional acceleration terms listed in Table 6 are typically about ±2 and are therefore significant especially near the top of the Ekman layer. However, these are not the dominant terms in the momentum balance, suggesting that the adjustment of the mean profiles was substantially complete. These results should therefore be representative of steady, neutral, barotropic conditions.

No other comparable observations have yet been obtained, although comparisons of neutral wind defect profiles with those obtained during the Wangara experiment (Clarke and Hess 1974) and results from numerical simulations (Wyngaard et al. 1974; Deardorff 1972) were presented in RS. However, other investigators have made simpler measurements in attempts to determine the functions \(A\) and \(B\) of the ‘resistance laws’ which relate the surface stress to external variables by matching Ekman and surface layer profiles (e.g. Tennekes 1973). This approach predicts that in steady, neutral and barotropic conditions both \(A\) and \(B\) should be universal constants defined by

\[
A = \ln(u_*/fz_0) - kU_g/u_* \tag{11}
\]

\[
B = (k/u_*) (V - V_g)_0. \tag{12}
\]

\(B\) can be estimated directly from the measurements shown in Fig. 4. A linear regression on the class-averaged data yields a value of \(B = 4.2 \pm 0.6\).

\(A\) cannot be similarly determined since no concurrent measurements of \(z_0\) were made during JASIN. Nevertheless, estimates can be made from the data by employing surface layer formulae (e.g. Dyer and Hicks 1970) in the form

\[
U_{10}/u_* = (1/k) \{\ln(10/z_0) - \Psi_m(10/L)\} = C_d^{-1/2} \tag{13}
\]
Figure 4. Non-dimensional momentum flux and wind defect profiles. A symbol (code as Table 3) represents an average from one box pattern while the solid curves connect dimensionless height class mean values. The limits of these classes and the number of observations in each are also shown. The dashed curves represent a balanced momentum equation as described in the text.

to eliminate $z_0$ from Eq. (11). The stability correction $\Psi_m$ is negligible for these data so an expression for $A$ may be written as

$$A = \ln(u_*/10f) + kC_D^{1/2} - kU_g/u_*.$$  

(14)

Values computed from Eq. (14) for each day using the data listed in Tables 2 and 3 scatter between $-2$ and 4, the mean being estimated as $1.4 \pm 0.8$. 
These values of $A$ and $B$ are in good agreement with previously published values obtained using different techniques as listed in Table 7.

<table>
<thead>
<tr>
<th>Source</th>
<th>Details</th>
<th>$A$</th>
<th>$B$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Deacon (1973)</td>
<td>Mean values from 6 experiments (1959–70)</td>
<td>1.9 ± 0.35</td>
<td>4.7 ± 0.15</td>
</tr>
<tr>
<td>Clarke and Hess (1974)</td>
<td>Wangara data</td>
<td>1.1 ± 0.5</td>
<td>4.3 ± 0.7</td>
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<tr>
<td>Melgarejo and Deardorff (1974)</td>
<td>Reanalysis of Wangara data</td>
<td>0</td>
<td>4.5</td>
</tr>
<tr>
<td>Arya (1975)</td>
<td>Reanalysis of Wangara data</td>
<td>1.0</td>
<td>5.1</td>
</tr>
<tr>
<td>This work (± standard error) 90% confidence interval</td>
<td></td>
<td>1.4 ± 0.8</td>
<td>4.2 ± 0.6</td>
</tr>
<tr>
<td></td>
<td></td>
<td>± 1.6</td>
<td>± 1.1</td>
</tr>
</tbody>
</table>

(b) The turbulent kinetic energy balance

The scaled TKE balance may be expressed in the form

$$
\frac{1}{u_*^2 f L} \frac{\partial \overline{\varepsilon}}{\partial t} = \frac{1}{k} \left( -u_* \right) \left( \frac{w' T_v}{w' T_v} \right) - \frac{\partial}{\partial \varepsilon} \left( \frac{w' p'}{\rho u_*^2} + \frac{w' e}{u_*^2} \right) - \frac{\partial (U/u_*)}{\partial \varepsilon} - \frac{\partial (V/u_*)}{\partial \varepsilon} - \frac{\varepsilon}{u_*^2 f}
$$

Each of these terms may be derived from aircraft measurements with the exception of $C$ (see for example Pennell and LeMone 1974; Lenschow et al. 1980). The measurements show that term $Q$ is negligibly small. The shear production terms $S1$ and $S2$ could be derived directly from the results shown in Fig. 4 (since the geostrophic wind shear is small compared with the actual wind shear) but the scatter in the $(U/g - U)/u_*$ measurements precludes an accurate determination of $S1$. Thus it was decided to obtain $S1$ as a residual by assuming $C$ to be negligible. Previous experimental and numerical studies indicate that this assumption can be justified provided levels close to the surface or near the bases of active convective clouds are avoided (Pennell and LeMone 1974; Lenschow et al. 1980; Nicholls et al. 1982). The dissipation rate, term $D$, was estimated from the inertial subrange of the one-dimensional vertical velocity spectra with use of a Kolmogorov constant of $4 \times 0.55$ (see for example Nicholls and Readings 1981). The buoyancy term, $B$, is a product of the normalized virtual heat flux and the stability parameter $(u_*/f L)$.

The variation of each of these terms with height is shown in Fig. 5(a), which shows best-fit curves to data taken from all but the most unstable day, 7 August, in which a significantly different balance was observed. This is shown in Fig. 5(b).

The near-neutral data of Fig. 5(a) show that unlike many previous results (see for example Lenschow et al. 1980; Caughey and Wyngaard 1979) both $B$ and $T$ are small throughout the Ekman layer and the self-cancelling to some extent. This implies that most of the shear production of TKE, the dominant production term throughout much of this
mixed layer, is dissipated locally. (The term S1 shown in Fig. 5(a) was used to deduce values of the velocity shear \((1/\mu_*)\partial U/\partial z\) and thence the profile drawn through the \((U_0^* - U)/\mu_*\) data in Fig. 4.) D decreases by an order of magnitude through the Ekman layer, and at the higher levels all the terms become small and roughly the same size as the uncertainty involved in their determination. With no export of TKE from lower levels and such small production terms, the intensity of turbulence becomes very small and could be suppressed even by very weak adverse density gradients.

By contrast, the slightly more unstable 7 Aug. case shows significant transport of TKE to upper levels and is much more reminiscent of behaviour in strongly unstable conditions. Although D is also markedly increased at upper levels it is likely that some of this energy could result in mixed layer deepening by entrainment. However, since \(h\) is still approximately \(0.2u_*/f\) on this occasion, either the growth rate is very slow because the buoyancy fluxes are still very small (by typical daytime over-land standards) or the observed degree of instability has only recently been established caused, perhaps, by advection over a warmer sea surface (recall that the only significant cold air advection occurred on 7 Aug.).

It is also interesting to observe how sensitive the TKE balance is to the quite small change of stability between 7 Aug. and the other occasions.

(c) The heat and water vapour balance

The sensible heat and water vapour balance of the Ekman layer were also investigated along similar lines to the momentum balance. However, the surface sensible heat fluxes were very small on all occasions (see Table 3) with only 7 August exceeding 10 W m\(^{-2}\). The variation with height was therefore correspondingly small, usually decreasing to small negative values in the upper parts of the Ekman layer.

If the evaporation of precipitation is neglected, which seems a reasonable approximation for the conditions encountered, the Ekman layer heat balance may be written

\[
D\theta/Dt = -\partial (\overline{w'\theta'})/\partial z + H
\]

where \(H\) represents the net horizontally averaged radiative heating rate.
The LHS of Eq. (15) was evaluated for each of the chosen days by a method analogous to that used to determine the acceleration terms in preceding sections. The Ekman layer average flux divergence, $\partial (w^' \theta^') / \partial z$ was obtained on each day by linear regression. A comparison of these two terms is shown in Fig. 6. The errors involved in the determination of each term are estimated to be about $\pm 1-2 \times 10^{-5}$ K s$^{-1}$ by similar reasoning to that used in (b) above. These two terms are approximately in balance although only 7 August has reasonably large values. Theoretical calculations of radiative effects in the JASIN mixed layer (Slingo et al. 1982) suggest that $H$ is about $1-2 \times 10^{-5}$ K s$^{-1}$ beneath stratocumulus at local noon. If this is typical of other occasions, the three terms in Eq. (15) are of comparable size on each occasion except 7 August where the rate of change of $\theta$ (dominated by the horizontal advection terms) is approximately balanced by the turbulent flux divergence.

Figure 6. Terms in the temperature balance equation for each day. The line represents a balance between the measured heating rate, $D \theta / Dt$, and sensible heat flux divergence. Symbols as Table 3.

Unfortunately, a failure of the fast response hygrometer on the Electra significantly reduced the number of water vapour flux measurements. The capability to determine the variation of $w^' q^'$ with height was therefore much reduced; however, some inferences may still be drawn. The measured rates of moistening of the Ekman layer were generally less than would have been anticipated had there been no water vapour transport through the top of the Ekman layer. Estimates of $D Q / Dt$ averaged throughout the depth of the Ekman layer, obtained in a similar manner to those of $D \theta / Dt$, varied between $-2$ and $3 \times 10^{-8}$ s$^{-1}$. Since the surface water vapour fluxes were all reasonably large and positive, a typical value being $1-2 \times 10^{-5}$ m s$^{-1}$, no transport above a 500 m deep mixed layer would
imply $DQ/Dt = 2.4 \times 10^{-8} \text{s}^{-1}$ in the absence of sources or sinks of water vapour. This is on the high side of the range quoted above and suggests, although somewhat indirectly, that relatively large values of $\overline{w'q'}$ may on occasion have been maintained near the Ekman layer top. This is suggested by the direct flux measurements. Figure 7 shows scaled $\overline{w'q'}$ measurements made on individual runs. Although most of the measurements are concentrated at the lower levels, no strong decrease with height is observed as was the case with the $u'w'$ measurements. The $\overline{w'q'}$ measurements scatter quite considerably with occasional points being larger at altitude than those nearer the surface. This suggests that mixing with drier air above $z = h$ is occurring, a discontinuity in the $q$ profiles being a common occurrence at this altitude (e.g. see Fig. 3). Such mixing would also be consistent with the negative $\overline{w'T'}$ values observed there. Further implications of this interpretation are discussed in section 7 below.

![Diagram](image)

**Figure 7.** Scaled humidity flux measurements. Symbols as Table 3, but here represent values from individual runs, not 'box-averages'.

6. **Spectra**

The turbulent fluctuation data from the aircraft were also investigated using spectral analysis (fast Fourier transform) techniques. Figure 8 shows average velocity spectra from aircraft data obtained at low level ($z \approx 0.02$) on eight occasions: the seven discussed above plus the stably stratified case. The data are two-dimensional averages, i.e. they contain data from along- and across-wind sampled runs around box patterns. Although
Figure 8. Average velocity spectra from the lowest non-dimensional height class. The inset table gives values of the stability parameter \(-u_*/fL\) for each day (symbols as Table 3).
differences have been shown to exist at low levels between spectra sampled in this way (Nicholls and Readings 1981), their effects on the averaged spectra presented here are small, especially when compared with the features discussed below. The spectra have been brought into coincidence at high wavenumbers (thus defining a scaling parameter $\Phi_0$) and are plotted on a double logarithmic scale as $kS_r(k)$ against $k$ (after Kaimal et al. 1972). All show an approach towards the expected $-\frac{3}{2}$ power-law behaviour at high wavenumbers although the combined effects of noise and aliasing are visible at the highest wavenumbers.

The spectra generally reveal two ranges of wavenumbers that contain a relatively large proportion of the total variance. In Fig. 8 both the $u$ and the $v$ spectra display minima at $k = 0.4$ km$^{-1}$ although this becomes less distinct with increasing instability (as measured by the stability parameter, $-u_*/fL$). This minimum is even more pronounced in the temperature and humidity spectra (Fig. 9), which are more strongly bimodal. Because the temperature fluctuation levels were generally so small, the high frequency ends of the spectra are noise limited and therefore show no sign of a decrease proportional to $k^{-2.25}$. The temperature spectra in the figure have therefore been normalized to give approximately the same values at low wavenumbers.

In contrast, the $w$ spectra (Fig. 8) are singly peaked so that at high wavenumbers the partition of energy between the three velocity components is approximately equal while the motion at low wavenumber is strongly two-dimensional.

The peaks of the regions of the spectra at high wavenumbers ($k > 0.4$ km$^{-1}$) tend to move to lower wavenumbers as instability increases in agreement with previous findings (see for example Kaimal et al. 1972). The same happens at higher levels in the Ekman layer, although the most noticeable effect is the reduction in amplitude of the high wavenumber part of the $u$, $v$, $T$ and $q$ spectra relative to the contributions from the low wavenumber end. The latter tend to remain unchanged throughout the Ekman layer and therefore dominate the overall variance at higher levels. The behaviour of the velocity variances with height is shown in Fig. 10. These are very similar to those published by Nicholls and Readings (1979).

One aspect of the behaviour of the $w$ spectra which can be investigated more quantitatively is the variation of $\lambda_m(w)$, the wavelength at which the peak of the $w$ spectrum occurs. This has been much investigated in the past because $\lambda_m$ is related to the integral scale of the flow (e.g. see the review by Pasquill (1974), pp 55–61) although there is little consensus, other than the broadest generalizations, for its behaviour at levels above the surface layer. In terms of the Ekman scales used in previous sections, the variation of $\lambda_m(w)$ is predicted to be

$$(f/u_*)\lambda_m(w) = F(zf/u_*, u_*/fL).$$

By analysing the variation of $\lambda_m(w)$ on each day (i.e. with $u_*/fL$ fixed), it was found that variation with $zf/u_*$ was approximately linear throughout the Ekman layer, i.e. $\lambda_m(w) \propto z$, as shown in Fig. 11. The stability dependence is expressed as a function of $u_*/fL$ and is also shown in the figure. The final expression then becomes

$$\lambda_m(w) = \phi(u_*/fL)z.$$ 

Previous investigations into the behaviour of $\lambda_m(w)$ have been in more convective situations. Both Kaimal et al. (1976) and Nicholls and LeMone (1980) showed $\lambda_m \propto z$ at low levels but tended towards a constant value at higher levels ($\lambda_m \approx 1.5$–$2.0z$) though both results were obtained in very unstable conditions. In slightly less unstable conditions, Donelan and Miyake (1973) reported $\lambda_m \propto z^{0.75}$ at all levels, which lies between the results obtained here in near-neutral stability ($\propto z$) and those obtained in very unstable
Figure 9. Average temperature and humidity spectra from the lowest non-dimensional height class. Symbols and stability parameters as Fig. 8. Scaling methods are described in the text.

Figure 10. Scaled velocity variances $\overline{v^2}$ vs. $z$. The symbols represent dimensionless height class mean values for cell days except 7 August of $u^2/u_*^2$ (○), $w^2/u_*^2$ (▲) and $v^2/u_*^2$ (■). ($w^2/u_*^2$ measurements for this day are shown by the dotted line).
conditions (constant with height). These data also support the conclusion by Nicholls and Readings (1981) that $\lambda_m(w)$ scaled with $z$ to quite high levels in the near-neutral boundary layer, although the range of heights considered was more limited in that study.

The value of $\varphi$ appears to be quite sensitive to variations in $u_*/fL$ around neutral. However, this is compatible with the observations made by Kaimal et al. (1972) in the surface layer which showed a similar threefold increase in $\lambda_m(w)/z$ (i.e. equivalent to $\varphi$) as $z/L$ decreased from 0 to $-1.0$. This is a similar range of stability to that encompassed by the JASIN data (see values of $L$ in Table 3) and is possibly one of the reasons for the lack of agreement noted amongst earlier data (e.g. Pasquill 1974) where the conditions were much less well specified. The neutral value of $\varphi$ is observed to be about $1.3$ whereas Kaimal et al. (1972) reported a value nearer $1.6$. This is possibly a consequence of the two-dimensional averaging used here since Kaimal et al. employed purely along-wind sampled data and the results of Nicholls and Readings (1981), while not observing significant differences in $\lambda_m(w)$ between along- and across-wind sampled data, did note relatively more energy at higher across-wind sampled wavenumbers.

The variation of the mean $uw$ and $wq$ cospectra with height is shown in Figs. 12(a) and 12(b) where the cospectra have been grouped into dimensionless height classes. These show that the motion responsible for the turbulent momentum and water vapour transport increases in scale throughout the Ekman layer in a well-defined manner although the fluxes are confined almost entirely to the high wavenumber region. A shift to lower wavenumbers of about one decade is observed between the lowest and highest levels, representing a movement of the peaks of the cospectra from about 200 m to 2 km.
Figure 12(a). Variation of the scaled $uw$ cospectra with non-dimensional height. The height class number and the mean values of $z$ for each are given on the right. Symbols as Table 3.
Figure 12(b). As Figure 12(a) but for \( w_g \) cospectra.
The two sets of cospectra are very similar in the lowest four classes, i.e. through the lowest half of the Ekman layer, but above this the \( wq \) cospectra tend to increase, especially at wavenumbers less than \( 1 \text{ km}^{-1} \) (however, there are only two occasions on which \( wq \) data are available at these levels and both tend to be more unstable than average). This suggests that some mixing with the generally drier air above \( \tilde{z} \approx 0.2 \) is occurring without significantly affecting the momentum transport. The \( wT \) cospectra are generally very small although at the higher levels both the \( wT \) and \( Tq \) cospectra became more negative over a wider range of low wavenumbers, consistent with the occurrence of some exchange at the top of the Ekman layer.

If the lowest-level cospectra shown in Figs. 12(a) and 12(b) are examined in detail small variations in the cospectra between the different days are apparent which appear to be related to the change in stability. The peaks of the \( uw \) cospectra tend to be shifted towards lower wavenumbers with increased instability, but the effect is small and comparable with the scatter in the average estimates. The average \( uw \) cospectrum from the lowest levels \( (\tilde{z} = 0.02) \) is shown in Fig. 13, which is also a good fit to the corresponding \( wq \) cospectrum. Also shown are the average \( uw \) cospectra from Nicholls and Readings (1981) obtained at similar levels over the sea from both across-wind and along-wind runs. The averaged JASIN data is in very good agreement with these results, lying roughly midway between these two curves.

7. DISCUSSION

The previous sections have demonstrated that in the near-neutral, barotropic conditions sampled by the aircraft during JASIN, small-scale turbulent mixing is primarily shear driven and is limited to a depth of approximately \( 0.2u_*/f \) (typically a few hundred metres) in the absence of externally imposed low-level inversions. For this structure to be observed, it is important that the surface buoyancy input is small, i.e. \( -h/L < 2 \). Substituting \( h = 0.2u_*/f \) and assuming that \( C_D = C_E = C_H \) in
Eqs. (3)-(5) yields

\[-h/L = (0.2gk/fT_o) \Delta T_v/U_{10} \approx 20\Delta T_v/U_{10}\]

where \(\Delta T_v = (T_v - T_{10}) + 0.61T(Q_v - Q_{10})\) and has units of K with \(U_{10}\) expressed in m s\(^{-1}\). The criterion may therefore be rewritten as

\[0 < -h/L \approx 20\Delta T_v/U_{10} < 2; \quad \text{or} \quad 0 < \Delta T_v < 0.1U_{10};\]

i.e. if the sea-air virtual temperature difference is less than 1 K when the wind speed is 10 m s\(^{-1}\), it is likely that mixing will be limited to a depth of \(0.2u^*/f\) if there is no inversion at a lower level. Such conditions could be expected to be commonplace over mid-latitude oceans in summer. For example, climatological data from OWS 'I' (59°N 19°W) show the sea-air temperature difference to lie between 0 and 1°C for about half the period from June to August (see Nicholls 1983).

In such situations, the TKE generated by shear production in the lower parts of the Ekman layer appears to be dissipated locally, not transported to upper levels to permit significant deepening by entrainment processes. However, the water vapour budget implies and certain observations reveal that some transport of water out of the Ekman layer does occur. While this could be explained in terms of upward growth of this layer, the growth rates required to maintain a water vapour flux comparable with surface values are \(-50\) m h\(^{-1}\) with the small humidity decreases typically observed near \(z = h\) (see Fig. 3 for example) and seem too large to be consistent with the observations and the interpretation summarized above. An alternative and more convincing explanation lies in the effect of cumulus convection which was observed to be a common feature while the measurements were being made (see Table 1).

Cloud formation depends upon air within the Ekman layer reaching its lifting condensation level (LCL), and cumulus will form if the upper levels are conditionally unstable (i.e. if \(\theta_e\) decreases with height). An example is shown in Fig. 3, where the LCL is about 360 m, the foregoing conditions are satisfied and a field of cumulus extending into the overlying stratocumulus was observed. The release of latent heat leads to vertical velocities within clouds which are significantly greater than those typical of the rest of the upper parts of the Ekman layer so that even a fairly small cloud fraction may transfer a significant amount of water. This has been documented by measurements from tethered balloons also made during JASIN and summarized in RS. In conditions similar to those illustrated in Fig. 3, the onset of cumulus convection is clearly very sensitive to small changes in the temperature and humidity structure. Indeed cumulus cloud was generally observed to be intermittent and irregularly distributed on scales of tens of kilometres, i.e. these cloud processes are associated with very different scales from those of the mixing processes that predominate within the Ekman layer (and may contribute to the variability observed at low wavenumbers in the Ekman layer spectra). It is possible that this larger-scale variability reflects coupling between the cumulus convection and the evolution of the Ekman layer, for example if there is initially no cloud and no removal of water vapour from the Ekman layer, the associated moistening will lower the LCL and increase the low-level \(\theta_e\) making cumulus convection more likely. Subsequent cloud formation may then transport sufficient moisture upwards to lift the LCL again. Alternatively, it may reflect the cumulative effects of changing surface conditions encountered along different trajectories. However, while many plausible physical mechanisms can be postulated which might affect the thermodynamic structure of the boundary layer on these larger scales (and therefore the cloud structure), the data presented here are limited in scale and cannot address such questions directly. Although it is suspected that
clouds are an important component of these types of boundary layer, especially at upper levels, it is unfortunate that even two aircraft could not adequately sample both the Ekman and cloudy layers. This must be a goal of future observational programmes.

The variation of the turbulent fluxes with height within the boundary layer was also deduced by Taylor et al. (1983) for selected periods during JASIN by the indirect method of integrating residuals from budgets mentioned in the introduction, the basic data being obtained from radiosondes launched from ships located at the corners of the triangular JASIN area. Their conclusions agree with those presented here in as much as their water vapour flux values vary little with height and the sensible heat fluxes are very small in the lowest few hundred metres, though their results also imply that significant small-scale turbulent transports, including momentum fluxes, extend up to the main inversion located typically at or above 1 km. The directly sampled aircraft and tethered balloon observations suggest that these become small at much lower levels. However, this may not be a serious disagreement (as a comparison of this sort may not be a profitable exercise) for the following reasons:

Firstly, the limitations of the methods mean that the results refer to non-overlapping areas and time periods. In an area where significant gradients were regularly observed in most quantities across the JASIN array, any difference in the results from different methods might merely reflect the response of the mixed layer to these changes integrated across the whole area over the time period under consideration.

Secondly, the uncertainties inherent in attempting to reconstruct flux profiles from budget residuals may systematically lead to apparently significant fluxes within the boundary layer under conditions such as those encountered during JASIN. As mentioned in the introduction, the budget technique is best suited to quasi-steady regions, e.g. the Trades. The question of adequate representivity is of central importance to this method where small differences in mean quantities are interpreted as fluxes. The degree to which three-dimensional fields of meteorological variables and their spatial and temporal gradients on scales of order 100 km and a few hours can be adequately determined by soundings from three points on similar scales is open to question, especially in an area where significant mesoscale (10–100 km) variability has been shown to exist. Further assumptions concerning the area-averaged cloud cover, liquid water storage, radiative effects, precipitation and pressure gradient corrections are also necessitated by this approach which can only add further to the uncertainties involved.

8. CONCLUSIONS

This study has presented observations made in the mid-latitude atmospheric boundary layer during the JASIN experiment. These measurements were made primarily from aircraft (13 flights on eight days) which provided not only the horizontal sampling capability lacking in previous investigations but also, using improved analysis techniques, standards of accuracy and measurements of quantities hitherto unobtainable. Few previous experimental studies have been made in similar conditions even though such boundary layers are believed to be widespread over the oceans at mid-latitudes. The results are mainly concerned with the structure of the lowest part of the ABL which could be adequately sampled by the aircraft. The main findings are summarized below.

(i) The use of aircraft coupled with carefully designed flight plans enabled horizontally averaged quantities and horizontal gradients to be determined with considerable accuracy, thus avoiding one of the major difficulties encountered in previous experiments. Both
the surface virtual heat flux and horizontal virtual temperature gradients, and therefore the geostrophic wind shear, were found to be small. The conditions encountered were therefore well described as near-neutral and barotropic.

(ii) The low-level flux measurements were found to be quite sensitive to changes in the larger-scale flow across the experimental areas showing that the turbulence responded quickly to local changes. Having extrapolated the results the short distance to the surface, good agreement was found between the aircraft measurements and previous surface-based results expressed in terms of bulk parametrizations. The aircraft data yielded values of $10^2C_D = 1.3 \pm 0.1$ and $10^2C_E = 0.9 \pm 0.1$. Sensible heat flux values were too small for a reliable estimate of $C_H$ to be made. The mean geostrophic drag coefficient was found to be $10^2C_r = 2.6 \pm 0.1$.

(iii) The depth of the layer in which turbulent mixing was observed, $h$, defined from the momentum flux profiles but supported by evidence from the thermodynamic profiles, was found to be approximately $0.2u_*/f$ and termed the Ekman layer. It was found that $h$ was usually considerably less than the height of the lowest inversion, $z_l$. This is in reasonable agreement with some previous experimental findings in near-neutral situations although many theoretical steady, neutral Ekman solutions (e.g. see RS) imply a much deeper layer. This discrepancy appears to be due to the effects of stratification. Even the weakly stabilizing density structure above $0.2u_*/f$ appears to be sufficient to suppress turbulent transfer. As instability increases, this layer would be expected to deepen to be eventually limited by $z_l$ as found in experiments conducted over heated land surfaces (or conversely if $z_l$ could be maintained less than $0.2u_*/f$ e.g. by subsidence). Expressed in terms of non-dimensional parameters, the conditions necessary for the mixed layer depth to be related to $u_*/f$ would appear to be that $-h/L$ be of order 1, i.e. conditions are close to neutral stability and that $z_l/h > 1$, i.e. the Ekman layer is not limited by a low-level inversion. Both conditions were satisfied during the JASIN flights. At these latitudes, the former constraint may be expressed in terms of more easily observed quantities:

$$0 < \Delta T_v (K) < 0.1 \ U_{10} \ (m \ s^{-1}).$$

(iv) Measurements of terms in the turbulent kinetic energy balance in conditions close to neutral ($0 < -h/L < 2$) imply that the major source, the shear production term, is dissipated locally with the buoyancy and turbulent transport terms relatively unimportant. As the shear production term diminishes rapidly with height through the Ekman layer as $0.2u_*/f$ is approached, high turbulence levels cannot be sustained there against the adverse density gradient. Small-scale turbulent mixing is therefore suppressed. However, on 7 Aug., when $-h/L \sim 8$, the balance is reminiscent of that encountered in more unstable situations (e.g. Lenschow 1974) where the turbulent transport term provides a mechanism to ‘export’ turbulent kinetic energy generated at lower levels to the upper part of a mixed layer and thereby promote deepening by entrainment.

(v) Scaling the momentum flux and velocity defect profiles within the Ekman layer ordered the data from different occasions quite successfully. These were found to balance quite closely and were also consistent with the measured terms in the turbulent kinetic energy balance. Estimates of the parameters $A$ and $B$ of neutral, barotropic Rossby number similarity theory from the same data were found to be $A = 4.2 \pm 0.6$ and $B = 1.4 \pm 0.8$, in good agreement with previously reported values.
(vi) Sensible heat fluxes were generally found to be very small throughout the Ekman layer although when the surface values were slightly larger, the resultant flux convergence was approximately balanced by horizontal advection. Conversely, the water vapour budget estimates imply variable water vapour fluxes in the upper Ekman layer with values comparable to those at the surface on some occasions. This is supported by the limited number of observations and is suspected to be primarily associated with cloud-induced transport. This has little effect on sensible heat or momentum transfer, which probably reflects the essentially different forms of the mean profiles of temperature, velocity and humidity. The possible effects of cloud-related mixing are not considered in present boundary layer similarity theories and it is therefore unlikely that these could successfully describe the thermodynamic structure of boundary layers such as those investigated during JASIN.

(vii) The spectra of $u$, $v$, $T$ and $q$ displayed two distinct ranges of wavenumbers which contained relatively large fractions of the total variance. The higher wavenumber region was generally located at values of $k$ greater than about 0-4 km$^{-1}$ or approximately $kh > 0.1$, with the other region at lower wavenumbers. In contrast, the $w$ spectra were singly peaked in the high wavenumber region. The partition of energy between the three velocity components was therefore fairly equal in this part of the spectrum while the low wavenumber motion was strongly two dimensional.

(viii) The high wavenumber parts of the spectra were found to vary consistently with changes in Ekman layer parameters i.e. non-dimensional heights and stability. The spectral peaks in this range of wavenumbers were generally observed to move towards lower wavenumbers as instability and height were increased, in agreement with previous results. Another notable feature was the decrease in amplitude of this part of the $u$, $v$, $T$ and $q$ spectra relative to the low wavenumber values as height increased. Since the low wavenumber spectra remained relatively unchanged throughout the Ekman layer, they contained most of the variance at higher levels. In consequence, the variance showed only a weak height variation. The $w$ spectra tended to broaden and move to lower wavenumbers with increased height and instability. The wavelength at which this spectral peak occurred was found to be reasonably well described at all levels by $\lambda_m(w) = \varphi(u_*/f)L_z$ over the limited range of stabilities encountered. At neutral stability $\lambda_m(w) \approx 1.3z$ although the stability dependence, $\varphi$, does vary quite quickly with $u_*/fL$.

(ix) The measured covariances were found to be confined almost entirely to the high wavenumber region. The cospectra were also found to vary in a similar manner to the spectra with changing height and stability.

(x) The results of this spectral analysis are consistent with the interpretation of the other results given above. The continuous small-scale turbulent mixing driven primarily by shear production is associated with the higher wavenumber regions of the spectra. This motion is also responsible for the measured turbulent fluxes but decreases considerably in intensity with increasing altitude within the Ekman layer. Superimposed on this are larger-scale motions which do not appear to be associated with the same physical processes. These findings imply that the scales on which turbulent transport occur change markedly from continuous small-scale mixing dominant below $z = h$ to much more intermittent processes above, characterized by significantly different and larger scales. These may be partially responsible for the fluctuations observed at low wavenumbers within the Ekman layer and are possibly related to cloud activity.
APPENDIX I

Method used to determine horizontal gradients

Data from each individual run along the side of a box were analysed as follows:

(i) The data were corrected for deviations of the aircraft from the run-average altitude. This was only found to be necessary for pressure and temperature where fluctuations caused by vertical deviations were large compared with typical horizontal variations. A dry adiabatic lapse rate was assumed, a good approximation within the Ekman layer. An example of the effect of these corrections on pressure measurements was shown in RS.

(ii) The change occurring along the run, e.g. a difference $\Delta p$ in the case of pressure measurements, was determined by a linear least squares fit to the corrected data.

(iii) This difference was then corrected for changes occurring during the course of the run (normally 10-15 min), the local rate of change being determined from a weighted average of hourly observations from the nearest ships. This could be accurately determined by ship-borne instruments as it did not involve an absolute pressure calibration. The size of this correction was typically only a few per cent of the larger values of $\Delta p$ and was negligible for all other parameters.

(iv) The distances covered during the run, $\Delta N$ (distance north) and $\Delta E$ (distance east), were calculated.

Each run therefore yielded measurements of $\Delta p$, $\Delta N$ and $\Delta E$. Since the gradient was assumed to be constant across the experimental area, least squares estimates of the components $\partial p/\partial N$, $\partial p/\partial E$ could be obtained by multiple regression in which

$$\Delta p_i' = (\partial p/\partial N)\Delta N_i + (\partial p/\partial E)\Delta E_i$$

and the quantity $\Sigma(\Delta p_i' - \Delta p_i)^2$ was minimized. Estimates of the uncertainties in the components so obtained were also made (e.g. Snedecor and Cochran 1967).

In applying this technique to the JASIN data, it was found in general that no significant changes in the gradients could be detected across the few hundred metres depth of the Ekman layer. This is certainly in agreement with the expected result for the pressure gradient, given the magnitude of the observed horizontal temperature gradients (see Table 2(b)). Therefore, all data from within the Ekman layer were taken together on each day and layer average values computed. The assumption of a constant gradient fits the pressure data very well on all occasions, as illustrated by Fig. 14 which shows $\Delta p'_i$ as a function of $\Delta p_i$ for 7 Aug. In this particular instance, the resultant layer-averaged geostrophic wind was found to be $009^\circ$ at 7.8 m s$^{-1}$ with a standard vector error of $\pm 0.4$ m s$^{-1}$. This figure also shows that the scatter in the individual measurements of $\Delta p$ is very small. Measurements of $\Delta p$ as made on side-by-side intercomparison runs show agreement between the C130 and Electra to within 0.1 mb over 60 km distances, as shown in the inset table in Fig. 14.
Figure 14. $\Delta p'$ plotted against $\Delta p$ for all runs on 7 August from C130 (●) and Electra (×) measurements. The line represents perfect agreement between the measured differences and the constant, linear gradient defined by the regression. The inset table shows measurements of $\Delta p$ made during runs with the aircraft flying side-by-side.

APPENDIX II

Notation

The following lists the meaning of symbols not explicitly defined in the text.

$C_D, C_E, C_H$  Neutral stability transfer coefficients for momentum, water vapour and heat at $z = 10\, m$

$\text{div}$  Horizontal divergence of the wind vector, i.e. $\nabla \cdot \mathbf{V}$

$E$  Distance east—Cartesian coordinate

$f$  Coriolis parameter

$G$  Geostrophic wind speed

$g$  Acceleration due to gravity

$k$  Von Kármán's constant ($= 0.4$) or wave number ($= 1/\lambda$)

$L$  Monin–Obukhov length scale

$N$  Distance north—Cartesian coordinate

$p$  Pressure

$q, Q$  Specific humidity

$q_l$  Specific liquid water content

$q_{sat}$  Saturation specific humidity

$Q_s$  Saturation value of $Q$ at temperature $T_s$
Mean quantities (represented by an overbar) were determined over the length of a measurement run (≈50–70 km) and fluctuations (') by removing the linear trend.

The mean and fluctuating components of single parameters defined above are denoted, e.g., \( \bar{u} = U + u' \), with capitals representing mean values to avoid a proliferation of overbars. When quantities are ordinarily represented by a capital letter (e.g. \( T \)), confusion does not arise since it is always clear from the context which meaning is implied.

\[
\frac{D}{Dt} = U \frac{\partial}{\partial x} + V \frac{\partial}{\partial y} + W \frac{\partial}{\partial z} + \frac{\partial}{\partial t}
\]

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