A field study of nocturnal stratocumulus: 
I. Mean structure and budgets

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SUMMARY

Observations of anticyclonic stratocumulus obtained with ground-based and balloon-borne equipment during the night of 19/20 November 1976 at Cardington, Bedford, UK, are described in relation to the synoptic situation.

The mean structure of the atmosphere from ground to about 300 m above cloud top is discussed. Of particular interest are the large temperature and humidity ‘jumps’ observed within a layer 1–10 m thick at cloud top, and the thermally stable surface layer about 200 m thick in the presence of an overcast sky and surface wind speed of 3–4 m s\(^{-1}\). The mean liquid water content profile is accounted for in terms of turbulence theory, but the cause of observed large variations in total cloud water content is not identified.

The locally steady conditions allowed estimates to be made of the local budgets of heat and water for the boundary layer (cloud top to ground) as a whole. In particular, the shape of the temperature inversion above cloud top allowed a quantitative estimate of cloud top entrainment rate to be made.

Although the local water budget appears to be in approximate balance, only about one-third of the radiative loss from cloud top is accounted for locally. It is not possible to account for this deficit in terms of the observations; it is suggested it could result in a downstream mean cooling of the boundary layer of about 0.5 K per 100 km.

The observed dispersal of cloud downstream cannot be explained by the local budget deficit (which would tend to increase cloud). It is estimated that a local subsidence rate of 2–4 cm s\(^{-1}\) through the cloud would be required to disperse it on the observed time scale. Potential sources of evidence for this are discussed.

1. INTRODUCTION

This paper is the first of three devoted to the description and interpretation of a comprehensive range of observations of nocturnal anticyclonic stratocumulus made during the night of 19/20 November 1976 with ground-based and balloon-borne equipment at Cardington, Bedford, UK (latitude 52°06'N, longitude 0°24'W). It describes observations of the mean structure of the atmosphere from ground to above cloud top and of the mesoscale situation. These observations are used to infer the heat and water budgets of the layer from cloud top to ground, and of the cloud layer within it. Paper II by Caughey, et al. (1982) discusses measurements of turbulence and cloud top structure and relates them to cloud top entrainment. Paper III by Slingo, et al. (1982) describes radiative and microphysical measurements made on this night and on two other nights when conditions were changing too rapidly to allow budget studies to be made. Using the observations presented in this paper and (where necessary) II and III, an attempt is then made to unravel some of the interaction of the various physical and dynamical processes thought to control the development or dispersal of stratocumulus on this occasion in the hope that it may illuminate the general problem of layer cloud. A list of symbols used in this paper is at Annex A.

(a) Background

This study is part of a continuing programme of field and numerical studies of radiation fog which aims to improve understanding of the basic physics in order to assess prospects of forecasting or modifying fog. The work reported so far (Roach et al. 1976; Brown and Roach 1976) concentrated on the local aspects. On a larger scale, particularly in
anticyclonic conditions in temperate latitudes in winter, the uncertainty in forecasting fog is often due to the uncertainty in forecasting stratocumulus cloud cover. Also errors in surface temperature forecasts frequently arise from a failure to forecast the presence (or absence) of fog or layer cloud. However, the study of stratocumulus is of interest in a wider context: the evolution of inversion capped convective boundary layers, with or without cloud, is of central importance to the parametrization of low level heat, moisture and momentum fluxes over land and sea in general circulation and numerical forecasting models of the atmosphere.

An original theoretical study by Ball (1960) has been followed by Lilly (1968), Carson (1973), Deardorff (1976a, b), Schaller and Kraus (1981), Schubert et al. (1977), Tennekes (1973) and Wyngaard and Cote (1974) and others. They have been concerned mainly with the diurnal development of the dry convective boundary layer in which the local heat budget is dominated by the surface and entrained heat fluxes. This may sometimes be an oversimplification, since Roach (1961), Zobel (1966), Moores (1977) and Moores et al. (1979) have shown that direct absorption of solar radiation within the boundary layer may on occasions be a significant contribution to the local heat budget. At night on the other hand, the general level of turbulent exchange in the nocturnal boundary layer is smaller (e.g. Caughey et al., 1979) but may be appreciable in elevated cloud layers with a large radiative cooling at the cloud top. Furthermore, anticyclonic subsidence is important and its role has not been sufficiently emphasized in recent studies but was recognized in an early study by James (1959). The situation is similar in some respects to the modelling of persistent subtropical marine stratocumulus by Schubert et al. (1977) in which surface fluxes and subsidence rates are of similar magnitudes to those observed or inferred in this study. Finally, an area of importance to an understanding of the cloudy or cloudless convective boundary layer is the identification of entrainment mechanisms occurring at the base of the inversion capping the boundary layer (Rayment and Readings 1974, Carson and Smith 1974, Palmer et al. 1979).

The main stream of field studies has reflected the current emphasis on cloud-free convective boundary layers by reports of extensive flux profile measurements (Lenschow 1970, Pennel and LeMone 1974, Kaimal et al. 1976, Caughey and Palmer 1979). Observations of cloudy boundary layers are few (Ivanov 1973). Measurements in layer cloud situations are particularly sparse although James (1956, 1959) made synoptic and aircraft studies in an attempt to identify the factors controlling the dissipation of nocturnal stratocumulus. Some observations of liquid water content and turbulence made in the USSR were reviewed by Cornford (1966). Paltridge (1974) made some radiometric measurements and related them to microphysical parameters. More recently Coulman (1978) attempted observations of the turbulent structure and heat transfer in stratocumulus.

(b) Observations

Tethered balloon observations at Cardington were made from 2345 to 0415 (All times are GMT). The balloon mounted instrument package was distributed (Fig. 1) along about 10m of balloon cable about 40m below the supporting balloon (1300m³). The package consisted of an optical scattering droplet spectrometer (III), two net radiometers (III), a turbulence probe (II) and a pressure sensor for height measurement. The pressure transducer and associated electronics were thermostatically controlled to within 2°C. A level calibration run near the surface at the start and end of the observational period allowed corrections to be made for any drift and the observed pressures are estimated to be accurate to within ±0.5mb. The droplet spectrometer with associated electronics and batteries weighed 150kg and was attached to a boom supported directly by the balloon via separate steel cables. The instruments are described in the papers indicated. The signals from the instrument package were telemetered to the ground using VHF telemetry links.

The ground based equipment consisted of an acoustic sounder (II), a microwave radiometer (III) for measuring total cloud water content, and the surface energy balance installation which is in routine operation at Cardington. The latter comprises sensors
for monitoring soil heat flux (at 5 and 10 cm depths), net evapotranspiration, net radiation flux at 1 m and profiles of mean wind speed and temperature to 16 m. Details of the equipment and the method of data analysis is given by Richards (1978).

2. SYNOPTIC SITUATION

An intense stationary anticyclone was centred over the Irish Sea and maintained a northeast gradient wind of 8–10 ms⁻¹ over SE England during the period. At 0000 England was covered by a sheet of stratocumulus except for a band of mostly clear skies about 100 km wide running from the Thames Estuary WSW to the English Channel. This band persisted for at least 24 hours, and its northern boundary lay within 30–60 km of Cardington until after 0400 when it began to move SW.

(a) Surface Observations

Surface charts (Fig. 2) indicate that conditions in space and time were rather uniform over East Anglia and the Midlands. Dry bulb temperatures were generally 7°C in this area with marked surface cooling in the clear band to the south. Horizontal temperature
gradients were small except across the cloud edge and near the East Coast where surface air was arriving onshore with a temperature of 9–10°C after passing over a sea surface temperature some 2°C higher.

Winds in the cloud layer observed from the tethered balloon (Fig. 4) were combined with 900 m wind observations from radio-sonde stations made at 1800, 0000 and 0600 in an attempt to construct the wind field within the cloud layer. Wind-field vectors (which were markedly ageostrophic) were used to estimate the trajectories of cloudy air downstream from Cardington. Points on Fig. 2 show the positions of air parcels at the chart times ‘released’ from Cardington (C) at 2100, 0000 and 0300 respectively. The 2100 and 0000 parcels reach the clear band some four and one and a half hours respectively after leaving Cardington, while the 0300 parcel is approaching the cloud edge three hours after release. This analysis also showed the cloud was generally dispersing along the northern edge and forming along the southern edge of the cloudless band.

![Figure 2. Observations of screen temperature at 2100, 0000, 0300, 0600 on the night of 19/20 November 1976.](image)

Cardington (c)
Isobars (mb)
4 okta isoneph
7 okta isoneph
900 m wind speed (m s⁻¹) and direction
Location of parcel at 900 m which left Cardington at time indicated (2100 in example) ● 21
(b) Upper Air Observations

The principal feature of the upper air data was the subsidence inversion overlying the cloud. Table 1 shows the pressure at the base of the inversion, and the inversion 'strength' (given by the difference between the maximum temperature in the inversion and the temperature at its base) at UK radio-sonde stations.

<table>
<thead>
<tr>
<th>Station</th>
<th>Time</th>
<th>$p_{b}$</th>
<th>$\theta_{b}$</th>
<th>$\delta T$</th>
<th>$q_{\text{min}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>GMT</td>
<td>mb</td>
<td>°C</td>
<td>°C</td>
<td>g kg$^{-1}$</td>
</tr>
<tr>
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<td>910</td>
<td>6.5</td>
<td>6.5</td>
<td>2.1 m</td>
</tr>
<tr>
<td></td>
<td>2345</td>
<td>906</td>
<td>6.5</td>
<td>6.5</td>
<td>2.4 m</td>
</tr>
<tr>
<td></td>
<td>0415</td>
<td>909</td>
<td>7.0</td>
<td>7.5</td>
<td>3.0 m</td>
</tr>
<tr>
<td>Camborne</td>
<td>2300</td>
<td>925</td>
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<td>7.0</td>
<td>2.2 d</td>
</tr>
<tr>
<td></td>
<td>1100</td>
<td>920</td>
<td>7.0</td>
<td>6.5</td>
<td>2.9 m</td>
</tr>
<tr>
<td>Crawley</td>
<td>2300</td>
<td>899</td>
<td>9.0</td>
<td>6.0</td>
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</tr>
<tr>
<td></td>
<td>1100</td>
<td>894</td>
<td>7.5</td>
<td>6.0</td>
<td>1.5 d</td>
</tr>
<tr>
<td>Larkhill</td>
<td>0800</td>
<td>905</td>
<td>7.5</td>
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<tr>
<td>Hensby</td>
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<tr>
<td></td>
<td>1100</td>
<td>905</td>
<td>6.5</td>
<td>6.5</td>
<td>1.3 m</td>
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<tr>
<td>Aughton</td>
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<td>6.5</td>
<td>8.0</td>
<td>1.3 d</td>
</tr>
<tr>
<td></td>
<td>1100</td>
<td>918</td>
<td>0.8</td>
<td>9.0</td>
<td>1.6 m</td>
</tr>
<tr>
<td>Valentia</td>
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<td>940</td>
<td>0.8</td>
<td>8.0</td>
<td>2.4 d</td>
</tr>
<tr>
<td></td>
<td>1100</td>
<td>910</td>
<td>6.5</td>
<td>10.0</td>
<td>2.1 d</td>
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<td>Long Kesh</td>
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<td>923</td>
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<td>13.0</td>
<td>1.7 d</td>
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</tr>
<tr>
<td></td>
<td>1100</td>
<td>980</td>
<td>6.5</td>
<td>9.5</td>
<td>2.4 d</td>
</tr>
<tr>
<td>Stornoway</td>
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<td>8.5</td>
<td>9.0</td>
<td>1.8 m</td>
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<tr>
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<td>923</td>
<td>8.5</td>
<td>11.0</td>
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</tr>
<tr>
<td>Lerwick</td>
<td>2300</td>
<td>950</td>
<td>8.5</td>
<td>7.0</td>
<td>3.2 d</td>
</tr>
<tr>
<td></td>
<td>1100</td>
<td>912</td>
<td>11.5</td>
<td>9.0</td>
<td>1.6 m</td>
</tr>
</tbody>
</table>

$\theta_{b}$ = potential temperature at inversion base

$p_{b}$ = pressure level of inversion base

$\delta T$ = total increase of temperature across inversion

$q_{\text{min}}$ = lowest humidity mixing ratio within 50mb layer above inversion base

$d$ = continuous decrease

$m$ = minimum within 50mb above inversion

Stations south of a line from the Wash to South Wales (first 5 stations in Table 1) show little variation in time or location in the inversion characteristics. North of the line, the inversion base is lower, $\delta T$ is much larger and the air above the base is on the whole drier. The thickness of the inversion varied between about 15 and 50mb. It was not possible to determine this separation accurately because of the limited resolution of standard radio-sonde ascents, but it seems likely, that 80–90% of $\delta T$ lies within 10mb of the inversion base and more than half within a temperature discontinuity associated with cloud top (See Section 3).

The uniformity of the potential temperature (near 7°C) at the base of the inversion independently of the inversion base height is an interesting feature.

3. Mean Structure

The principal feature was the remarkable steadiness of most of the parameters observed
at Cardington during the night. This has enabled the mean structure and associated heat and water budgets to be estimated.

(a) Surface Observations

Screen temperature showed little variation from 7.5°C throughout the night. Screen dew point decreased slightly during the night and by about 0.7°C from 2345 to 0415; the BALTHUM* ascents (Fig. 4) indicate that this drying out did not extend above about 200 m.

A gradual backing of the wind occurred during the experimental period with little variation of speed from about 4 m s⁻¹ at 37 m apart from a period of much lighter winds near midnight, possibly due to lee effects from the Cardington Balloon Sheds. These are some 70 m high, 200 m long and located about 600 m to the NNE of the observing site.

Hourly mean logarithmic wind profiles between 0.5 and 8 m above ground exhibited an 'unstable' curvature in what was clearly a thermally stable situation and this may also be a lee effect of the Cardington Sheds. A least squares linear fit to the curves yielded a value of $u_*$ of about 0.2 m s⁻¹ and of $z_0$ of about 0.02–0.03 m.

(b) Surface Energy Balance

This is expressed in the form

$$ (R_0)_{0} + E_0 + S_0 = G_0 $$

which follows from the condition that the energy flux has to be continuous across the soil-air interface. Observations made with the surface energy balance equipment at Cardington over the period 1800 on 19 November until 0600 on 20 November are summarized in Table 2. Hourly averages shown are for the 1 h period beginning at the time indicated. Fluxes are averaged 3 h since estimates of $E_0$ cannot be resolved over shorter periods. Upward directed fluxes are positive, and indicate a loss of heat by the earth. The net radiative flux $(R_0)_{0}$ varies little from 8 W m⁻² during the experimental period and agrees well with the balloon borne radiometer when near the ground (Paper III).

The soil heat flux, $G_0$, was estimated by linear extrapolation from soil flux measurements at 5 and 10 cm depth. This is acceptable if there has been no large change in surface energy balance components (e.g. due to cloud clearance) in the previous three hours or so. $G_0$ was found to vary little from 5 W m⁻² and is very nearly in balance with $(R_0)_{0}$.

The latent heat flux, $E_0$, displayed an erratic variation between zero and about 35 W m⁻². Since the lysimeter measurements are subject to an uncertainty of 10 W m⁻², an attempt was made to estimate evaporation using a version of Penman's expression discussed by Monteith (1973) which is essentially a bulk aerodynamic method adapted to use the saturation deficit observed at a given level, $z$, above a canopy. This expression

$$ \frac{E(z) = A((R_0)_{0} - G_0) + \rho c_p [e_s(T(z)) - e(z)]/r_a}{A + \gamma} $$

is where $A$ = slope of saturation vapour pressure curve (≈ 0.67 mb K⁻¹)

$\gamma$ = modified psychrometric constant (≈ 0.63 mb K⁻¹)

$r_a$ = effective resistance of air and canopy below height $z$ to diffusion of water vapour, and given approximately by $u(z)/u_*^2$

Substituting appropriate values from Table 2 and using wind observations at 1 m yielded values of $E_0$ between 15 and 25 W m⁻² over the period 2300–0400. This is in broad agreement with the lysimeter observations, and a representative value of 20 W m⁻² has been used in the budget estimates discussed later.

The sensible heat flux, $S_0$, was not measured directly. The residual term in Eq. (1), $S_0$, is about −25 W m⁻². This is consistent with the observation of a thermally stable lapse in the lowest 200 m and is evidence that the latent heat required to evaporate surface

* Tethered balloon ascents made on a routine basis every 6 hours to measure temperature, humidity and wind speed up to a height of about 1 km. The BALTHUM balloon is not the same balloon as that used for the main experiment.
TABLE 2. SURFACE OBSERVATIONS

<table>
<thead>
<tr>
<th>Time GMT</th>
<th>Screen T</th>
<th>Screen Td</th>
<th>q g kg⁻¹</th>
<th>Wind (37 m tower) m s⁻¹</th>
<th>(R_L)_0</th>
<th>G₀ W m⁻²</th>
<th>E₀ W m⁻²</th>
<th>S₀</th>
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<tr>
<td>1800</td>
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<td>6·1</td>
<td>3·0</td>
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<td>30</td>
<td>17</td>
<td>3</td>
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<tr>
<td>1900</td>
<td>7·0</td>
<td>6·1</td>
<td>5·7</td>
<td>3·5</td>
<td>020</td>
<td>10</td>
<td>12</td>
<td>−26</td>
</tr>
<tr>
<td>2000</td>
<td>7·8</td>
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<td>6·0</td>
<td>4·0</td>
<td>020</td>
<td>10</td>
<td>−26</td>
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<td>8</td>
<td>7</td>
<td>35</td>
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<td>5·1</td>
<td>3·5</td>
<td>360</td>
<td>6</td>
<td>7</td>
<td>35</td>
</tr>
</tbody>
</table>

moisture has to be drawn from the atmosphere since the soil heat flux is nearly balanced by radiative loss. It also implies a Bowen ratio of about −1.

The values adopted for the discussion of the heat and water budget of the boundary layer in Section 4 during the main period of observation are

\[
(R_L)_0 = 10 \pm 2 \text{ W m}^{-2}
\]

\[
G_0 = 5 \pm 3 \text{ W m}^{-2}
\]

\[
E_0 = 20 \pm 10 \text{ W m}^{-2}
\]

\[
S_0 = -25 \pm 10 \text{ W m}^{-2}
\]

The error values are intended to indicate the general level of uncertainty in the figures quoted; they are not standard deviations or confidence limits in a formal sense.

\(c\) Balloon Profile

The pressure sensor recording is shown as a function of time in Fig. 3. Vertical profiles were obtained in the periods indicated by the sloping trace, and eddy flux measurements during level flights. Between profiles 2 and 3, attempts were made to raise and lower the equipment very slowly through the very sharp temperature step found at cloud top.

![Figure 3 The height of the tethered balloon package as a function of time. The numbered arrows identify the profiles discussed in Papers I, II and III.](image_url)
Figure 4. (a) Observations of temperature and humidity (b) Observations of wind speed and direction. The temperature and humidity profiles are labelled with profile number from Fig. 3. Tethered balloon observations
BALTHUM observations at 2300/19 ●
BALTHUM observations at 0500/20 ○
Location of cloud top — — —
Location of cloud base — — —
The temperature scale refers to Profile 0; Profile 6 is displaced 2 °C to the right of Profile 0. The layer acronyms are defined in the text.

(d) Vertical Profiles

Profiles of temperatures, humidity mixing ratio, wind speed and direction during the initial ascent and final descent are shown in Fig. 4. Also shown are BALTHUM observations of temperature, humidity and wind speed from the 2300/19 and 0500/20 ascents.

The quasi-steady conditions are evident from the temperature profiles and show that the depth of atmosphere sampled may be divided into layers as indicated in Fig. 4.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Limits (mb)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SIL</td>
<td>A subsidence inversion layer with a maximum temperature near 6 °C at 870–880 mb decreasing to 4 °C just above cloud top</td>
<td>880–906</td>
</tr>
<tr>
<td>EIL</td>
<td>Entrainment interface layer containing a temperature step of about 5 °C</td>
<td>906</td>
</tr>
</tbody>
</table>
FIELD STUDY OF NOCTURNAL STRATOCUMULUS: I

CRL The upper part of cloud containing the radiative flux divergence region

BL

CLL The remaining (lower) part of the cloud layer

SCL The sub-cloud dry adiabatic layer

SSL The stable surface layer

The lowest four layers are collectively called the boundary layer.

(i) Temperature

The temperature profiles showed two unexpected features: the first was the stable layer in the lowest 20mb of atmosphere, which is shown by the tethered balloon (but not by the BALTHUM due to coarse height resolution) to be capped by a small inversion; the second was the sharpness of the temperature step at cloud top concentrated in a layer (EIL) which varied between 1m and 10m in thickness (see Table 1 in Paper II). Lapse rates are near or slightly steeper than wet adiabatic in cloud and dry adiabatic in the sub-cloud layer – the lapse in the upper part of the sub-cloud layer appears to be slightly super-adiabatic in both ascents.

Finally the reference adiabat indicates an almost imperceptible cooling (~ 0·2°C) of the boundary layer during the five hours between profiles. In contrast, there is a warming of about 0·5°C in the subsidence inversion layer, associated with a decrease of cloud top height by some 50m near the end of the main observing period (also see Fig. 2, Paper II).

(ii) Humidity

There is a hydrolapse from ground to cloud top which is steepest in the surface stable layer. The steepening in the upper part of the cloud disappears if cloud liquid water content is added to the vapour mixing ratio. There is a decrease of 1·0–1·5g per kg immediately above cloud top over a layer, rather thicker than the EIL. There were also one or more marked minima of humidity 10mb or so above cloud top – a feature observed by James (1959) and commented on by Cornford (1966).

A slight decrease of 0·2–0·3g per kg within the surface stable layer (but little change at higher levels) is evident from the BALTHUM observations and is consistent with the screen observations. The tethered balloon hygriostor observations on the descent at the end of the experimental period were discarded due to instrument malfunctions.

(iii) Wind

About 90% of the wind speed increase from the surface is achieved within the surface stable layer. A further slow increase in speed takes place above this, but there appears to be little systematic change across cloud top. There was a backing of about 20° in wind direction throughout the layer during the period. While it may appear that most of the surface stress is taken up within the surface stable layer – which might then be identified with the Eckman layer (depth h) – the winds in the whole layer above the surface stable layer are markedly sub-geostrophic. The ageostrophic wind is about 5 m s⁻¹, corresponding to an acceleration of about 4 x 10⁻⁴ ms⁻². This is comparable with the characteristic magnitude of the stress term, uₚ²/h, within the Ekman layer, and large compared with the value above it. The curvature of trajectories of air leaving Cardington (Section 2a) also indicates a markedly ageostrophic flow.

(iv) Radiation

Profiles of the net radiative flux and heating are shown in Figs. 3 and 4 of Paper III. There is a net radiative loss at cloud top of about 75 W m⁻². Radiative cooling of 5–10 K h⁻¹ is concentrated in the upper 5 mb of cloud, and a level of zero flux and flux divergence is located about 15 mb below cloud top. There is a marked maximum of heating of about 0·3 K h⁻¹ just above cloud base.

Radiative cooling just above cloud top is 0·5–1·0 K h⁻¹ and a change of net flux of
Figure 5. Liquid water content from Profile 1.

- Observation
  A  Adiabatic value
  B  computed from turbulence observation and latent heat flux estimate.

5–10 W m\(^{-2}\) occurs in the first 10–20 mb above cloud top.

(e) Cloud Parameters

Figure 5 shows observations of liquid water content obtained from profile 1. Liquid water content (\(q_L\)) increases approximately linearly with height at 70–80\% of the adiabatic value. Fluctuations in \(q_L\) appear to increase relatively and absolutely to a maximum near cloud top.

Cloud water content (Fig. 1 of Paper III) shows large fluctuations between about 30 and 100 g m\(^{-2}\) but no apparent long term trend about a mean of 56 g m\(^{-2}\). This contrasts with the steadiness of most other parameters.

Time series spectral analysis of the water content using a 'maximum entropy' technique revealed peaks of variance at 3 h, 1 h and 38 min, corresponding to advection scales of about 75, 25 and 15 km respectively. The cloud base records from stations within 100 km indicated that the cloud base fluctuated with a standard deviation of 30–50 m. This is about 10\% of cloud thickness (\(h - z_c\)) and may be fairly representative of cloudbase fluctuations at Cardington. It may also indicate the magnitude of cloud-top fluctuations.

(f) Subsidence Inversion Layer and Entrainment Rate

The entrainment interface layer is maintained in a quasi-equilibrium state against diffusive processes, and the acoustic sounder (Paper II Fig. 2) indicates rather slow excursions of this layer of up to 30 m on either side of a mean level of about 1120 m. All transits through this layer and the main subsidence inversion above indicate that the radiative flux and temperature profiles move 'bodily' with these excursions. Comparison of profiles 0 and 6 in Fig. 4 show that the lateral shift of the temperature profile above cloud by about 0.5°C corresponds approximately to an adiabatic displacement by the observed downward shift of cloud top between profile 0 and 6. The application of similarity theory to the temperature jump at cloud top (Eq. (3) of Paper II) yields an estimate of \(w_e\) of about 0.5–1.0 cm s\(^{-1}\), but it will now be shown that the temperature structure of the main inversion layer above the temperature jump allows an independent estimate of entrainment rate (\(w_e\)) to be made.

Air above cloud experiences radiative cooling as it approaches cloud top; at the same time it is warming by subsidence. Thus
\[
\frac{DT}{Dt} = \frac{\partial T}{\partial t} + V \frac{\partial T}{\partial s} + w \frac{\partial T}{\partial z} = H_R - wT
\]  
(3)

where turbulent diffusion is evidently negligible (see Paper II) Referring Eq. (3) to a height co-ordinate, \( z' \), relative to cloud top, we have

\[
\left( \frac{\partial T}{\partial t} \right)_{z'} + V \left( \frac{\partial T}{\partial s} \right)_{z'} - w_e \left( \frac{\partial T}{\partial z} + \Gamma \right)_{z'} = (H_R)_{z'} - \Gamma \frac{dh}{dt}
\]  
(4)

Following the observations that \( z' \) surfaces are also approximately isentropic surfaces, it is seen that

\[
\left( \frac{\partial T}{\partial t} \right)_{z'} + V \left( \frac{\partial T}{\partial s} \right)_{z'} = - \Gamma \frac{dh}{dt}
\]  
(5)

Eq. (5) expresses the change in temperature of an air parcel in adiabatic motion along an (undulating) \( z' \) surface.

Thus Eq. (4) reduces to

\[- w_e \left( \frac{\partial T}{\partial z} + \Gamma \right)_{z'} = (H_R)_{z'}
\]  
(6)

Eq. (6) yields a method of estimating the entrainment velocity \( w_e \) entirely from local measurements, but contains no information on the subsidence rate.

Using a radiative transfer computational scheme designed to cope with the large gradients of temperature and opacity occurring near cloud top (Roach and Slingo 1979), the observations were used to compute the profile of \( H_R \) in layers about 1 mb thick from cloud top near 905 mb to about 880 mb. This was combined with observations of the temperature profile to estimate entrainment velocity, \( w_e \), using Eq. (6). Most of the values of \( w_e \) lay between \(-0.4\) and \(-0.6\) cm s\(^{-1}\), but the remainder were widely scattered. A plot of the cumulative percentage of estimates less than the abscissa value shown in Fig. 6 yields a heavily skewed and peaked distribution far removed from normal. The scattered values are derived from minor irregularities in the temperature profile which were probably generated by local thin patches of turbulence which are characteristically embedded in stable fluids, but in general make an insignificant contribution to mean energy fluxes. It is therefore reasonable to regard the scattered values as belonging to a different population not conforming to Eq. (6). Those observations which do appear to conform to Eq. (6) are represented by line B which corresponds to a mean of 0.53 cm s\(^{-1}\) and a standard deviation of 0.10 cm s\(^{-1}\). Thus an entrainment rate of 0.5 cm s\(^{-1}\) is used in the budget estimates which follow.

4. BUDGETS

Integration of the thermodynamic equation and the continuity equations for water substance from the surface to cloud top yields sensible and latent heat budget equations for the boundary layer as a whole:

\[
\int_0^h \rho c_p \left( \frac{\partial T}{\partial t} + V \frac{\partial T}{\partial s} + w \left( \frac{\partial T}{\partial z} + \Gamma \right) \right) dz = \int_0^h \rho L C d z
\]

\[
= S_0 - \rho_h c_p (w'T)_h + (R_L)_0 - (R_L)_h
\]  
(7)

\[
\int_0^h \rho L \left( \frac{\partial q_w}{\partial t} + V \frac{\partial q_w}{\partial s} + w \frac{\partial q_w}{\partial z} \right) dz = E_0 - \rho L (w'q_w)_h
\]  
(8)

The RHS of Eqs. (7) and (8) represent the sensible and latent heat fluxes at the upper and lower boundaries, while the LHS sides represent the transport terms. The cloud-top entrainment fluxes can be estimated following Deardorff RV(1976). He defines the height of the mixed layer, \( h \), as the ‘mean height of the lower excursions of cloud top’. This lies between a lower level \( z_1 \), which denotes the lower level of the zone of radiative flux divergence (and corresponds to the base of the cloud radiative layer (CRL) defined in 3d. above),
Figure 6. A plot of the cumulative percentage ($P_w$) of observations of vertical velocity less than the abscissa value of $w$. Line A is the normal distribution corresponding to the mean and standard deviation of all the observations of $w$; line B is the normal distribution approximating to the observations clustered about a central value near $-0.5 \text{ cm s}^{-1}$.

and an upper level $z_2$ as 'the uppermost height to which the diffusive turbulence extends'. The cloud-top inversion extends from $h$ to $z_2$ and $r$ is the fraction of flux divergence between $h$ and $z_2$.

This leads Deardorff to two entrainment relationships

$$w_e \Delta T = r \Delta F - (w'T')_h$$
$$w_e \Delta q_v = -(w'q'_v)_h$$

where $\Delta F$ is the total radiative flux difference between levels $z_1$ and $z_2$. While Deardorff's analysis may be applicable to daytime stratocumulus, our observations of nocturnal stratocumulus indicate some significant differences:

(i) The cloud-top height excursions observed with the acoustic sounder do not appear to be part of the cloud-top entrainment process (see 3f(iii) above).

(ii) The inversion above cloud top extends well above the entrainment interfacial layer in this case study, but not in Deardorff's model. Cloud top is observed to coincide
with the base of the entrainment layer to within 1–2 m; the layer itself is only a few metres thick and contains only a small fraction of the total flux difference $\Delta F$. Thus if cloud top is taken as corresponding to Deardorff’s definition of $h$, the base of the cloud radiative layer is $z_1$ and the top of the entrainment layer is $z_2$, then the corresponding value of $r$ is very much less than 1 and can be neglected, thus

$$w_e \Delta T = -\langle w^{T'} \rangle_h . \quad \quad \quad (11)$$

Substituting Eq. (11) in Eq. (7) and Eq. (10) in Eq. (8), we now have estimates for all the fluxes of sensible and latent heat across the boundaries of the boundary layer in terms of observed quantities as follows:

- Surface sensible heat flux ($S_0$) = $-25 \pm 10$ W m$^{-2}$
- Entrainment sensible heat flux ($\rho_h c_p w_e \Delta T$) = $-30 \pm 10$ W m$^{-2}$
- Surface radiative flux ($R_L)_0$ = $10 \pm 2$ W m$^{-2}$
- Cloud top radiative flux ($R_L)_h$ = $75 \pm 5$ W m$^{-2}$

Thus the RHS of Eq. (7) yields a net heat loss of

$$60 \pm 15 \text{ W m}^{-2}$$

where again the error values imply a general level of uncertainty. This imbalance is dominated by the radiative loss term, with the local surface and entrainment heat fluxes rather small and largely self-cancelling.

For the boundary values of latent heat flux we have an estimate of surface latent heat flux from 3b. $E_0 = 20 \pm 10 \text{ W m}^{-2}$. Eq. (10) yields an estimate of entrainment latent heat flux ($\rho L w_e \Delta q_w$) = $20 \pm 10$ W m$^{-2}$. This implies an approximate balance between surface evaporation and cloud top entrainment. The only term on the LHS of Eq. (8) that can be estimated from the observations is the local rate of change term. The slight drying out of the surface stable layer corresponds to a loss of about 5 W m$^{-2}$.

5. EDDY FLUX PROFILES

Since turbulence measurements were not made below cloud, it is possible to infer only the general character of profiles of eddy fluxes of sensible and latent heat through the boundary layer.

The main feature of the profile of the eddy flux of sensible heat is the large upward flux at the base of the cloud radiative layer required to satisfy the heat budget for the layer. Using Eq. (7) with the appropriate limits of integration, the significant terms are

$$\rho c_p w_e \Delta T = (R_L)_h - \rho c(w^{T'})_{\text{max}} . \quad \quad \quad (12)$$

where ($w^{T'}$)$_{\text{max}}$ is the eddy heat flux at the base of the cloud radiative layer where the net radiative flux is zero. Eq. (12) implies an upward eddy heat flux of 40 W m$^{-2}$ at the base of the cloud radiative layer changing to a downward flux of 30 W m$^{-2}$ at cloud top (implying an eddy flux convergence in the cloud radiative layer equivalent to a heating of about 2 K h$^{-1}$. The eddy heat flux also changes (probably nearly linearly) to a downward flux of 25 W m$^{-2}$ at the surface, implying a mean eddy flux divergence equivalent to a cooling of about 5 K d$^{-1}$ through most of the boundary layer.

Since entrainment and surface evaporation fluxes are about the same, and since there is no sink of water substance in cloud (as there is of heat), it appears that the eddy flux of latent heat does not vary much throughout the boundary layer.

6. DISCUSSION

The principal questions raised by the observations of the structure and local heat and water budgets of the boundary layer are:

(i) Can the downstream dispersal of cloud be reconciled with a local heat loss of 60 W m$^{-2}$ and a local balance of water substance?

(ii) What is the influence of anticyclonic subsidence on the observed behaviour?

(iii) What determines the mean profile of liquid water content in cloud and the large fluctuations in total cloud liquid water content ($Q_L$)?
These questions are interrelated, and the ensuing discussion leads only to inferences rather than definite conclusions, but is of value in identifying aspects of the problem requiring future attention.

(a) The Local Heat and Water Budgets

The simplest approximation to the observed situation is a cloud-topped airstream of constant depth flowing at a uniform speed \((6-7 \text{ m s}^{-1})\) and cooling following the motion in response to the heat loss. Since the total water substance appears to be locally balanced, the resulting decrease in temperature should result in an increase in cloud. Referring to Eq. (7), the downstream cooling is represented by the first two terms on the LHS (of these, the local cooling – see Section 3d(i) – contributes about \(15 \text{ W m}^{-2}\)), and the change in cloud content should appear in the condensation term. The vertical advection term cannot be estimated since the subsidence rate is not known, but is likely to be small below cloud top due to the prevailing near adiabatic lapse rate through most of the boundary layer.

However, this simple model is at variance with the observed downstream cloud dispersal.

(b) Cloud Dispersal

In order to assess what is required to disperse a 300 m thick cloud layer in 2–4 hours at night, it is first necessary to account for the observed profile of liquid water content. Providing the effect of gravitational droplet settling is small compared with the characteristic magnitude of turbulent velocity fluctuations (as in this case), it is reasonable to suppose that water substance in liquid and water vapour forms will be subject to the same turbulent transport mechanism. Thus this mechanism will not differentiate between the liquid and vapour phases, and will therefore determine the profile of total water substance through the boundary layer. This may be tested by applying the flux-gradient relationship to the profile of \(q_w\) to see if this accounts for the observed \(q_L\) profiles. We have

\[
E = - K_q \frac{\partial q_w}{\partial z} \quad . \quad \quad \quad (13)
\]

where \(K_q\) is an eddy diffusivity for water substance. Within cloud, we may also use the identity

\[
\frac{\partial q_w}{\partial z} = \frac{\partial q_s}{\partial z} + \frac{\partial q_L}{\partial z} = - \frac{\partial}{\partial z} (q_{ad} - q_L) \quad . \quad \quad (14)
\]

Integrating Eq. (14) from cloud base to a height \(z'\) above cloud base and using Eq. (13)

\[
\bar{\rho} (q_{ad} - q_L) = \int_0^{z'} \left( E/K_q L \right) \, dz \quad . \quad \quad (15)
\]

where \(\bar{\rho}\) is mean air density over layer of thickness \(z'\). Consideration of the water budget in the previous section suggests that \(E\) does not vary much from \(20 \text{ W m}^{-2}\) throughout the boundary layer.

The turbulence measurements reported in Paper II may be used to infer a profile of \(K_q\) through cloud; from profiles of \(\varepsilon\) and \((\lambda_m)_{\omega}\) in Figs. 8 and 9 using the similarity relation (e.g. Smith 1975)

\[
K_q = e^{\varepsilon/2} (\lambda_m)_{\omega}^{1/2} / 15 \quad . \quad \quad (16)
\]

where it is assumed that \(K_q\) is equal to eddy diffusivity of heat and momentum. \(\varepsilon\) is the turbulent energy dissipation rate and \((\lambda_m)_{\omega}\) is the maximum wavelength energy in the spectrum of vertical velocity fluctuations \(nS(n) \varepsilon \log n^*\). The profile of \(\rho q_L\) derived using Eqs. (13)–(16) agrees well with the observed profile through most of the cloud (Fig. 5), but estimated values are less than those observed near cloud top. Profiles of liquid water content and total water substance can be represented on an expanded tephigram (Fig. 7). At any given level in cloud, represented for example by the isobar \(AA'\), the difference between the mixing ratios at \(A\) and \('A'\) (read from the saturation

*where \(n = \lambda^{-1}\)
mixing ratio domain on the tephigram) represents the liquid water content, while the difference between A and A' represents the adiabatic content. Thus the stippled area OBB' is approximately proportional to the total water content in cloud (not exactly, since the saturation mixing ratio domain is non-linear). If cloud-top height does not change downstream, then some process is required which shifts the temperature profile entirely to the right of the water substance profile. This may be achieved by a warming of the boundary layer by about 2 K or a drying out by about 0.4 g per kg, or a combination of both in a period of 2-4 h. The latter combination could be achieved by a cloud-top entrainment of about an order of magnitude greater than that already inferred and to be balanced by an equally large subsidence rate to keep cloud top height constant. This does not seem tenable in the light of earlier discussion, and it appears that diabatic processes are too weak to disperse a cloud layer at night on the observed time scale. The cloud-top entrainment instability criterion derived by Randall (1980) is nowhere near satisfied at Cardington and is unlikely to occur downstream.

It is thus necessary to turn to dynamical processes to disperse the cloud, the most obvious of which is a locally large subsidence rate extending through the cloud layer and displacing it downwards – the locally steady cloud top observed at Cardington could slope downstream.

(c) Subsidence

The effect of this subsidence may be assessed by reference to the tephigram (Fig. 7b). Suppose the cloud is displaced bodily downwards by an amount Δp. The $q_w$ profile will be bodily displaced to $q_w'$ in a direction parallel to a mixing ratio isopleth. Each air parcel

![Diagram](image)

**Figure 7(a)** Expanded tephigram demonstrating relationship between temperature, total water substance and liquid water content profiles. CB and CT represent cloud base and cloud top. The lines T and $q_w$ represent profiles of temperature and water substance respectively. T has been drawn following a dry adiabatic lapse below cloud and a wet adiabatic lapse through cloud. The profiles of T and $q_w$ intersect at O at cloud base. The dashed line through O represents an isopleth of saturation mixing ratio corresponding to the value of $q_w$ at cloud base.

(b) As (a), but demonstrating the effect of adiabatic downward displacement of the layer as a whole. The profiles of T and $q_w$ prior to downward displacement are as in (a). The dotted line represents an isopleth of water substance mixing ratio corresponding to the value of $q_w$ at cloud top. The other lines are explained in the text.
(e.g. \( P \) in Fig. 7b) on the temperature profile initially in cloud will proceed down a wet adiabat until it meets its corresponding point \( (P') \) on the \( q_w \) profile at \( P'' \). This is the point of disappearance of cloud water within the parcel, following which the parcel will proceed down the dry adiabat. If \( \Delta P \) is chosen so that cloud within a parcel starting at cloud top just evaporates, this to a first approximation will represent the vertical displacement required to evaporate the whole cloud. It appears to be about 25 mb (220 m) or about two-thirds of the original cloud depth. The construction leads to a layer beneath the inversion of lapse rate intermediate between wet and dry adiabatic and of thickness equal to that of the original cloud. A vertical displacement of 220 m in 2–4 h implies a local subsidence rate of 2–4 cm s\(^{-1}\) after allowance for an entrainment rate of about 0.5 cm s\(^{-1}\). The details of the final cloud dispersal are likely to be modified as the opacity of the cloud decreases and some of the radiative loss from cloud top is transferred to the ground. In addition there is no reason to believe that subsidence rate is constant through the cloud layer but this seems plausible. Finally, we note that the cloud edge runs roughly parallel to the Chiltern Escarpment. The clearance of stratocumulus to the lee of high ground is a well-known phenomenon.

From the argument so far, locally enhanced subsidence is largely responsible for dispersing cloud downstream on the time scale observed. We now look for some independent evidence of this from a study of the dynamics of the wind field in the boundary layer.

The most direct method of estimating the subsidence rate is by integrating the divergence of the horizontal wind field. An objective analysis (see Annex B for description) of the surface wind field (Fig. 8) for 0000 and 0600 on the night of this study demonstrates a broad association between a belt of maximum divergence and the cloud-free area over Southern England. Wind observations at 900 m are too sparse for mesoscale analysis to include the divergence field at any other level, but using this analysis alone the maximum of \( 3.4 \times 10^{-5} \) s\(^{-1}\) would need to continue for only about 300 m above the surface to generate subsidence rates of about 1 cm s\(^{-1}\). Thus the observed association of clear skies with large surface divergences may indicate that regions of increased subsidence are more likely than not to lie above regions of large surface divergence.

Although this gives no information on subsidence rates downstream from Cardington, we note that the air leaving Cardington appears to be flowing towards a region of increased surface divergence, and therefore (possibly) increased subsidence.

Another approach is to integrate the vorticity equation through the friction layer to obtain the vertical velocity at its top (see e.g. Sawyer (1959)).

\[
\frac{D}{Dt} \zeta_a + \zeta_a \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) = - \frac{1}{\rho} \frac{\partial}{\partial z} \left( \frac{\partial \tau_x}{\partial y} - \frac{\partial \tau_y}{\partial x} \right)
\]  

(17)

where it is assumed that the solenoidal and twisting terms can be neglected since these involve quantities (horizontal temperature gradient, vertical wind shear) which are generally small in this situation. Using the continuity equation in the form

\[
\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0
\]  

(18)

and integrating Eq. (17) from the surface to some level \( z \) where the stress is small compared with the surface stress leads to

\[
w(z) = \int_0^z \frac{D}{Dt} (\ln \zeta_a) \, dz \quad - \frac{1}{\rho \zeta_a} \left( \frac{\partial (\tau_0)_x}{\partial y} - \frac{\partial (\tau_0)_y}{\partial x} \right)
\]  

(19)

The first term on the right-hand side represents dynamical forcing, while the second represents frictional forcing. Taking the surface stress as

\[
\tau_0 = -\rho C_0 |V_0| V_0
\]  

(20)

the frictional term in Eq. (19) involves the relative vorticity of the surface wind such that subsidence is induced at the top of the layer of large stress divergences if the relative vorticity is anticyclonic. The relative vorticity of the surface wind field in Fig. 8, shows a belt of
maximum anticyclonic vorticity over the English Channel, which is displaced well to the south of the cloudless band and is not therefore obviously associated with it. Thus, the dynamical forcing term may be significant though this cannot be demonstrated quantitatively from the observations available.

Figure 8 demonstrates the presence of mesoscale structure in what is, on the synoptic scale, an area of small horizontal change. There are enough observations over Southern England to suggest that this structure has some reality, although there are problems of representativeness of the wind observations and a variable density of data points.

(d) Fluctuations in Cloud Water Content $(Q_L)$

$Q_L$ is typically 1% of the total water substance in the boundary layer and it can be seen from Fig. 7 that a small shift in the $q_w$ profile relative to the temperature profile causes a large change in $Q_L$. Noting that $q_L$ is approximately proportional to height above cloud base, we have
\[ Q_L \propto (q_L)_h^2 \]

Therefore
\[ \frac{\delta Q_L}{Q_L} = 2 \frac{\delta (q_L)_h}{(q_L)_h} = 2 \frac{\delta (q_w)_h - \delta (q_s)_h}{(q_L)_h} \]

Thus a fractional change in \( q_w \) or \( q_s \) is magnified by a factor \( 2(q_w)_h/(q_L)_h \) i.e. about 20 in this case, and an increase of about 25% in \( Q_L \) would be generated by an increase of about 0.05 g per kg in \( (q_w)_h \) or decrease of 0.1 K in \( T \). However, the mechanism for generating the small changes required has not been identified.

6. Conclusions

The principal inference drawn from this case study is that mesoscale variations of stratocumulus cloud cover relevant to local forecasting may be controlled by variations in the subsidence rate rather than by (slower) diabatic processes. It follows from failure to explain the loss of heat of 60 W m\(^{-2}\) from the boundary layer at Cardington with the observed downstream dispersal of cloud in terms of diabatic processes. The heat loss is dominated by cloud-top radiative loss, and is probably representative of many anticyclonic situations at night over land in which a layer of stratocumulus cloud lies underneath a dry subsidence inversion with no cloud above. Surface and cloud-top entrainment fluxes appear to be small and to some extent self-cancelling, and are unlikely to balance the main heat loss which presumably manifests itself as a slow downstream cooling of the layer from cloud top to ground. It will be difficult to detect this cooling from synoptic observations since it may often be masked by local dynamical effects, as in this case study. Over the sea, the surface fluxes are likely to be comparable with the cloud-top radiative loss and mesoscale dynamical effects are probably less obtrusive; over land by day solar radiation input will usually more than offset any nocturnal cooling.

Other interesting aspects of the mean structure are the entrainment interfacial layer (discussed in Paper II) and the thermally stable surface layer which appears to contain most of the stress divergence. The thermal stability is probably caused by surface evaporation but may be enhanced by strong divergence within the layer (Fig. 8). The presence of this stable layer under an overcast sky at night may be of significance to the dispersal of pollution from low-level sources.

The emphasis of much layer cloud modelling (and of Papers II and III of this series) is on diabatic processes; on the other hand, the emphasis of some mesoscale modelling is on dynamics with little or no account taken of the water substance in the atmosphere. It is hoped that this study will add to the data base required to inform and so increase the realism of all mesoscale modelling studies. In particular there is a need for information on mesoscale variations of subsidence rate through and above the boundary layer, incorporating the effects of local topography.

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FIELD STUDY OF NOCTURNAL STRATOCUMULUS: I

ANNEX A

Symbols

The principal symbols used in the text are listed below, other symbols are defined as they arise. As the independent height variable \((z)\) is positive upwards, fluxes use the same convention. Thus a positive net flux by this convention implies a loss of heat by the earth.

\[c_p\] specific heat of air at constant pressure \((1005 \text{ J kg}^{-1}\text{K}^{-1})\)

\[C\] rate of condensation per unit mass of air

\[\Delta\] incremental change of parameter across the temperature and humidity jumps above cloud top

\[E\] energy equivalent of eddy flux of total water substance \(= \rho L w q_w\)

\[f\] Coriolis parameter \((1.15 \times 10^{-4} \text{ s}^{-1} \text{ at latitude } 52^\circ)\)

\[F\] net radiative flux in dynamic units \(= R_L/\rho c_p\)

\[G_o\] soil heat flux at surface

\[g\] acceleration due to gravity \((9.8 \text{ m s}^{-2})\)

\[\Gamma\] dry adiabatic lapse rate \((0.0098 \text{ K m}^{-1})\)

\[h\] height of cloud top

\[h_e\] Ekman layer depth

\[H_R\] radiative heating rate \(= -\partial F/\partial z\)

\[L\] latent heat of evaporation of water \((2.5 \times 10^6 \text{ J kg}^{-1})\)

\[p\] pressure

\[q\] water vapour mixing ratio

\[q_s\] saturation water vapour mixing ratio

\[q_L\] liquid water mixing ratio

\[q_{ad}\] adiabatic liquid water mixing ratio

\[q_w\] \(q_s + q_L\)

\[Q_L\] \(\int_{z_e}^{h} \rho q_L \, dz = \text{total liquid water content in cloud as measured by microwave radiometer}\)

\[R_L\] net radiative flux

\[\rho\] air density

\[s\] distance along trajectory

\[S\] sensible eddy heat flux \(= \rho c_p \overline{w'T'}\)

\[T\] temperature

\[\tau\] Reynold stress tensor \(= -\rho(\overline{u'w'} + \overline{j^\prime v^\prime}) = \tau_x + j \tau_y\)

\[u_s\] surface friction velocity

\[u', v', w'\] components of velocity fluctuations from mean values

\[V\] horizontal wind vector; \(V - \text{horizontal wind speed}\)

\[w\] vertical velocity

\[w_e\] entrainment velocity

\[w'T'\] \(\overline{w'T'}\)

\[w'q_w\] covariation of fluctuations in \(w\) and \(q_w\)

\[z_c\] height of cloud base

\[z_0\] roughness length

\[\zeta\] relative vorticity \(= \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}\)

\[\zeta_a\] absolute vorticity \(= \zeta + f\)

The bar indicates a mean over a horizontal distance large compared to the thickness of the boundary layer
The analysis of the surface wind fields on which Fig. 8 is based is constructed iteratively by the method of 'successive correction'. At each stage we have, corresponding to each observation, a set of 'residuals', each defined as the difference between the observation and the partially completed analysis - the observations themselves constitute the first set of residuals. These residuals are crudely analysed using simple but efficient spatially-recursive numerical filters that produce a smooth new correction field to be added. The process is repeated several times and at each stage the numerical filters are tuned to a progressively smaller scale as far as is justified by the local observation density, which is automatically estimated as a by-product of the scheme. The nature of the compromise between filtering observations and achieving a smooth analysis can be adjusted by choosing the asymptotic scale of structure to which the scale of successive corrections is permitted to tend. Full details of this scheme form the subject of an unpublished Meteorological Office (Met. O. 11) Technical Note.

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