The structure of radiatively driven convection in stratocumulus

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(Received 5 May 1988; revised 18 November 1988)

SUMMARY

Data from a series of five research flights in marine stratocumulus are used to investigate the structure of radiatively driven, free convective layers. Conditional sampling methods are used to determine the properties of the primary convective elements: negatively buoyant downdraughts. The turbulent velocity data are also subjected to spectral analysis and the turbulent kinetic energy (TKE) balance is evaluated for each case. Results from each case are found to be very similar if scaled using appropriate mixed layer quantities.

Downdraughts are selected using a criterion based on vertical velocity. The distributions of their intersected widths at cloud top, and the velocity spectra, are consistent with downdraughts occupying narrow regions (~0.1-0.15 h wide, h being the mean mixed layer depth) around the periphery of larger, regular (~0.5 h-0.75 h diameter) updraughts as suggested by cellular patterns observed in the cloud tops. Downdraughts are found to carry over half of the total turbulent fluxes of heat and water substance within cloud, and also transport variance down the mean gradients. Their thermodynamic properties are consistent with their containing a small fraction of air from above cloud top although their negative buoyancy is almost entirely due to the incorporation of radiatively cooled air. The properties of the downdraughts near cloud top suggest they are formed primarily as a result of the local horizontal convergence of upwelling motions constrained by the inversion. While being cooled radiatively, these quasi-horizontal motions scour the base of the inversion, incorporating some drier air before being forced downwards in relatively narrow zones. Comparisons with other convective layers over land or sea suggest that convection in stratocumulus proceeds with a greater mass flux and correspondingly reduced differences between convective elements and the environment, possibly reflecting the improved ventilation possible at an inversion interface compared with much rougher surfaces.

The TKE balance divides into three main regions and is consistent with the interpretations given above. The pressure transport is implied to be the largest source term close to cloud top. Comparisons with model results reveal some important differences.

1. INTRODUCTION

In a recent series of papers, data from a number of research flights have been used to study various aspects of stratocumulus-topped boundary layers. Nicholls (1984) and Nicholls and Leighton (1986, subsequently referred to as NL) were primarily concerned with describing the mean structure in terms of layer averages, identifying the main physical transport processes and assessing the effect of these processes on the mean structure. In particular, the consequences of the distribution of internal energy sources within the boundary layer, especially those associated with radiation, were emphasized. This was extended by Turton and Nicholls (1987, TN) to a more general study of the diurnal cycle exhibited by stratocumulus.

In all of these studies, quantities were generally defined in terms of area-averages, on scales of some tens of kilometres, i.e. using a bulk description of the boundary layer. While this approach does have many advantages, the small-scale detail is unavoidably lost. The purpose of this paper is to use the same data, but to concentrate on the small-scale features of the convective motion fields associated with these cloud layers. While cellular patterns are readily seen in the tops of layer clouds from above (e.g. see NL), there have been no reported in situ measurements of the corresponding convective motions. An understanding of this turbulence structure is essential if the evolution of layer clouds is to be successfully modelled since it controls both vertical transport through the boundary layer and entrainment across mixed layer boundaries.

* Deceased, 26 December 1988 (obituary, p. 713).
Increasing use is also being made of high resolution numerical models to investigate the structure of cloudy boundary layers. A number of higher-order closure models (e.g. Duysherke and Driedonks 1987; Chen and Cotton 1987; Wai 1987) have appeared recently. These make various assumptions concerning the turbulent kinetic energy (TKE) balance which have received little observational verification. Furthermore, large-eddy simulations are now capable of producing details down to scales $\sim 10$ m, which are comparable with the scales measured from aircraft and will increasingly be used to study entrainment processes. It is hoped that the results presented here will be useful in assessing the realism of these simulations in a particularly difficult area for both numerical modelling and observational work.

The data were all obtained in situations where cloud top radiative cooling was the major source of buoyancy driving a convectively mixed layer. While there have been numerous classical laboratory studies of free convection between heated parallel plates, the situation where convection drives fluid away from a free surface has received much less attention. Even so, most of these have involved an interface between essentially immiscible fluids, e.g. air/water (Katsaros et al. 1977), although McEwan and Paltridge (1976) is an exception. In the atmosphere, where the Rayleigh number is typically very large, convective layers heated from below by a warm land or ocean surface are invariably turbulent and have received considerable attention (e.g. Kaimal et al. 1976; Lenschow and Stephens 1980), but again very few observations have been reported in conditions where radiative cooling drives convection away from a free density interface. Cloud-topped boundary layers are such a case, although these are further complicated by the possible effects of buoyancy fluctuations which accompany water phase changes.

Three different but complementary analysis techniques are employed in this paper: conditional sampling, spectral analysis and the evaluation of terms in the TKE budget. Because no other directly comparable results are available, comparisons are made wherever possible with different types of convective layers and numerical simulations. Definitions of quantities and notation not explicitly included in the text are contained in an appendix.

2. THE DATA

(a) Data selection

The data were collected during five flights by the Meteorological Research Flight Hercules aircraft in horizontally uniform, unbroken sheets of stratocumulus over United Kingdom sea areas. These selected flights all display a common structure. In each case, the cloud layer occupied the upper part of a mixed layer within which vertical gradients of $\theta$, and $q_T$ were found to be small. Furthermore, turbulent mixing within this layer was being maintained by buoyant convection with negligible production due to wind shear (see section 5). These cases have all been described in some detail in NL and are here referred to by the same flight numbers (see also Table 1). Further information concerning instrumentation, flight procedures, the definition of significant levels and other relevant data may also be found in NL. Symbols are defined in the appendix.

Some of the main features encountered on each flight are summarized in Fig. 1, which shows the mean cloud base and cloud top levels together with the adiabatic liquid water content profile. The mean cloud thickness was found to vary from 190 m to 540 m. The lower mixed layer boundary was located at various depths beneath cloud base as shown in the figure, the reasons behind this variation having been discussed in NL. Throughout this paper, the mean mixed layer depth is referred to as $h$. Data have been selected from all those straight and level runs within or just outside the mixed layers.
TABLE 1. DETAILS OF CASES CHOSEN FOR ANALYSIS

<table>
<thead>
<tr>
<th>Flight No.</th>
<th>Date</th>
<th>$h$ (m)</th>
<th>$\Delta q_T$ (g/kg)</th>
<th>$\Delta \theta_e$ (K)</th>
<th>$w_*$ (m/s)</th>
<th>$T_{V*}$ (K)</th>
<th>$q_{T*}$ (g/kg)</th>
<th>Plot symbol</th>
</tr>
</thead>
<tbody>
<tr>
<td>511</td>
<td>27 Apr. 1982</td>
<td>370</td>
<td>-3.0</td>
<td>0.9</td>
<td>0.57</td>
<td>0.026</td>
<td>0.028</td>
<td>●</td>
</tr>
<tr>
<td>526</td>
<td>22 Jul. 1982</td>
<td>480</td>
<td>-2.1</td>
<td>2.1</td>
<td>0.57</td>
<td>0.020</td>
<td>0.044</td>
<td>△</td>
</tr>
<tr>
<td>528</td>
<td>29 Jul. 1982</td>
<td>480</td>
<td>-4.7</td>
<td>-2.9</td>
<td>0.63</td>
<td>0.024</td>
<td>0.021</td>
<td>■</td>
</tr>
<tr>
<td>620</td>
<td>2 Nov. 1983</td>
<td>700</td>
<td>-3.0</td>
<td>1.7</td>
<td>0.78</td>
<td>0.025</td>
<td>0.028</td>
<td>▼</td>
</tr>
<tr>
<td>624</td>
<td>16 Nov. 1983</td>
<td>1120</td>
<td>-3.6</td>
<td>0.8</td>
<td>1.18</td>
<td>0.036</td>
<td>(0.028)</td>
<td>+</td>
</tr>
</tbody>
</table>

The derivation of the value in parenthesis assumes saturation in cloud.

The heights of these flight legs are also indicated in Fig. 1 together with the mean liquid water content measured at that level.

(b) Scaling

It is unlikely that any single method of scaling data obtained in convective layers containing stratocumulus will yield results which have universal applicability since the TKE balance is potentially much too complicated. The presence of additional internal energy sources associated with radiative effects and phase changes, whose size and location depend strongly on the cloud structure itself, make the situation much more complex than the corresponding cloud-free case. Nevertheless, the common features mentioned above suggest that some form of mixed layer scaling is the most appropriate choice. This has already been shown by NL to be useful when considering the vertical velocity variance, and a similar scheme was also employed by Moeng (1986) to good effect. The scaling parameters are defined as follows:

$$w_* = \left\{2.5 (g/T_V) \int_h^0 wT_V \, dz' \right\}^{1/3} \tag{1}$$

$$T_{V*} = w_*^2 T_V / gh \tag{2}$$

$$q_{T*} = (wq + wq_t)_{z'=0} / w_* \tag{3}$$

Figure 1. Mixed layer boundary levels, adiabatic and mean observed liquid water contents for the five selected flights.
Because of the coarse vertical spacing of the measurement levels and the relatively large vertical gradients anticipated, the integral in Eq. (1) was derived from results obtained with a diagnostic mixed-layer model, as described in NL. This was found to work well in that paper. Values of each of these scaling parameters are listed in Table 1.

The total water fluxes ($wq + wq_1$) and buoyancy fluxes measured on each run are shown scaled in this way in Fig. 2. The maximum buoyancy flux is observed in the upper part of the cloud layer in every case, implying that cloud top cooling was the dominant buoyancy production mechanism (see also section 3(c)). At lower levels, the buoyancy fluxes decrease downwards to small or slightly negative values in the bottom part of the mixed layer. The scatter observed in the data serves as a reminder not only of the difficulty of making these measurements, but also that the precise forms of these buoyancy flux profiles are expected to differ from case to case as conditions vary (see NL). Only the gross characteristics are similar. However, case 624 is an exception. Here, the mixed layer extended down to the sea surface and there were significant heat and water vapour fluxes across the lower boundary (see Fig. 2). Consequently, $\overline{wT_v}$ increases downwards in the lower half of the mixed layer, although this difference is somewhat concealed in this form of plot by the scatter in the other data. Substantial contributions to the total buoyancy production were made both at the surface and by latent heat release at cloud base. Some differences in the observed small-scale structure might therefore be anticipated in this case.

The largest values of total water flux were usually found near cloud top (see Fig. 2), so $q_{T*}$ was defined in these terms. While this does not take account of any other sources of $q_T$ fluctuations, for example fluxes across the lower mixed layer boundary, Fig. 2 shows that the fluxes generally decreased downwards and were significantly smaller than $w_*q_{T*}$ in the lower part of the mixed layer in all cases except 624, where they were comparable owing to the aforementioned surface transfers (NB. an instrument failure led to the loss of all $q$ data in cloud on flight 624. A value of $q_{T*}$ was therefore obtained by assuming that $\overline{wq}$ could be related to $\overline{wT}$ in cloud using the saturation condition, as described in NL, which also showed this to be a reasonable assumption on the other flights. The same assumption was also used when calculating $\overline{wT_v}$, although the water vapour term is small in this case.)

![Figure 2. Scaled total water and virtual temperature fluxes v. $z'/\delta$. The plotting symbol for each flight is given in Table 1.](image-url)
Figure 3. Scaled velocity variances and TKE, $\overline{E}$, $\nu$, $z'/h$. Symbols as Fig. 2.

(c) General characteristics: measured variances

The variances measured on each of the selected runs, scaled with the quantities listed above, are shown in Figs. 3 and 4. For conformity with the conditionally sampled data presented in the next section, low frequency variations in all time series have been reduced by removing a 30 s (3 km) running mean.

The vertical velocity variance reaches a maximum of about $0.4w_*^2$ in the upper part of the mixed layer, around $z' = 0.2h$. Above this, the proximity of the inversion causes a strong decrease. Values above cloud are about an order of magnitude smaller. Significant internal wave activity above cloud top was detected only on flight 511 and the variance is correspondingly higher for this reason. The decrease observed below $z' = 0.2h$ and through the rest of the mixed layer is more scattered, reflecting differences in the TKE balance between cases in this region. Case 624 shows no significant variation of $w'^2$ with height. A similar plot of the same data in NL showed fair agreement with the results of Lenschow et al. (1980), obtained in surface-heated convective boundary layers, although the data here are slightly smaller owing to the low frequency filtering.

The $\overline{u'^2}$ and $\nu'^2$ data are very similar and have approximately the same values as $w'^2$ throughout most of the mixed layer. However, an increase is often observed close to cloud top, in the region $0 < z' < 0.2h$, where the inhibiting influence of the inversion on vertical motion and the requirements of continuity cause the local horizontal components to increase. The values outside the mixed layers do not generally show the same degree of reduction as $w'^2$ since horizontal motions are not constrained to the same extent by the stable stratification.
Figure 4. Scaled temperature and total water variance \( \sigma^{2}/h \). Symbols as Fig. 2.

The TKE is shown in Fig. 3. It is found to increase steadily upwards through the mixed layer from a value of about 0.5 at the bottom to a peak value around 1 near \( z' = 0.1h \). There is a sharp reduction by an order of magnitude in the inversion. Case 624 shows little variation with height in the mixed layer about a value of 0.5 (instrumental problems on this flight caused data losses above cloud top).

The scatter in the temperature and water data shown in Fig. 4 is much greater. The \( T^{2} \) data for case 624 are significantly smaller than the others at all mixed layer levels, suggesting perhaps that the more energetic convection associated with buoyancy input at both boundaries could result in reduced temperature fluctuation levels. In the other cases, the minimum values tend to be found in the middle of the mixed layer. This minimum is a characteristic of convective layers (e.g. Lenschow et al. 1980; Deardorff 1974), as are the extremely large values found very close to cloud top \( (z' < 0.1h) \) where increases of 100 or greater may be observed as the inversion is approached. The smaller increases seen as \( z' = h \) is approached also reflect the presence of the slight stratification at that level.

The \( q_{1}^{2} \) data are broadly similar, except that they continue to decrease from cloud top down towards the lower boundary, suggesting that on these flights, mixing at cloud top is the main source of water substance fluctuations throughout the whole layer. The few measurements which were made in case 624 are somewhat larger, implying that the cloud-top-only scaling employed here might be insufficient in this case.

The numerical values of the scaled temperature and water variances are both similar to those reported by Lenschow et al. (1980) at a comparable altitude, e.g. at \( z' = 0.15h \), \( T^{2}/T_{v_{*}}^{2} \sim 10 \) and \( q^{2}/q_{v_{*}}^{2} \sim 15. \)

3. Conditional Sampling

Even a casual visual inspection of in-cloud time series data reveals that coherent downdraughts are the most striking feature, as illustrated in Fig. 5. Subsequent analysis confirms that these structures contribute an important part of the total variance and a major fraction of the fluxes. Previous authors (e.g. Caughey et al. 1982) have also commented on their importance. This section analyses the properties of these downdraughts by conditional sampling. The use of criteria based on vertical velocity ensures that these structures must have some degree of vertical coherence, and by assuming that their ensemble-averaged properties vary smoothly with height, information about the
vertical structure of these downdraughts can be inferred from the original, horizontally sampled, data.

(a) Method

A preliminary investigation showed that none of the downdraughts were intersected for longer than $13\text{s} (1.3\text{km})$, so a $30\text{s} (3\text{km})$ running mean was first removed from all time series to aid automatic detection. (Increasing the length of this mean was found to have no further effect on the results.) Prime superscripts denote the residual fluctuation data.

During preliminary analyses of these data, a number of slightly different criteria and thresholds were tried before settling on the method described below, which was partly chosen for its simplicity and ease of application. However, these early results also enable some estimate of the sensitivity of the event size and frequency statistics to variations of the chosen thresholds to be made. These are discussed in section 3(b). Although the criteria for the selection of downdraughts or ‘events’ are necessarily subjective, the following procedure, illustrated in Fig. 5(a), was developed, which appears to detect most of the features which an observer would pick out by eye.

(i) If $w' < w_{\text{thres}}$, the nearest zero crossing points on either side define the extent of the downdraught and its intersected width, $d$. On all flights, $w_{\text{thres}}$ was set to $-0.5w_s$.
(ii) If $d < h/20$ (this varied from $19\text{m}$ to $56\text{m}$ between cases) the event was rejected as too small. Events rejected for this reason were counted and typically make up only a few per cent of the total sample.
(iii) For each accepted event, data values, $s'$, were interpolated onto a regular grid as shown in Fig. 5(b). Ten points lie within an event ($\pm i = 1$ to 5) and up to five points on either side in ‘tails’ extending up to $+d/2$ outside the event boundaries ($\pm i = 6$ to 10).

![Figure 5(a)](image)

Figure 5.(a) An example of events detected by the conditional sampling algorithm from part of a run at $z = 1140\text{m}$ during flight 511. The threshold ($-0.5w_s$) is indicated.

![Figure 5(b)](image)

Figure 5(b). Definition sketch of the notation employed.
If a ‘tail’ section would have extended into a region occupied by another accepted event, data from this section were excluded and the ‘tail’ shortened. A minimum gap of 10 m between successive event boundaries was required, or they were combined and treated as one.

The algorithm was applied to the whole data set. Figure 5(a) illustrates the event boundaries defined in a short section of \( w' \) data.

Statistics from the whole ensemble of \( n \) events detected during a run were then calculated:

\[
\bar{s}^i = \sum_{\text{all events}} s^i / n \quad \text{for } \pm i = 1 \text{ to } 10
\]

\[
\bar{d} = \sum_{\text{all events}} d / n.
\]  

Examples of the ensemble-averaged results from runs at three different levels on flight 511 are shown in Fig. 6. On average, the downdraughts are relatively cool and dry 40 m below cloud top, although there are considerable fluctuations within particular events, as shown by the vertical bars (which represent the typical standard error in the mean of \( s^i \) within a downdraught). At the next level down, the temperature and water contrasts have almost disappeared, and the average downward velocity has increased. At \( z' = 385 \) m, below cloud base and close to the lower mixed layer boundary, the downdraughts have decelerated and are now slightly positively buoyant on average. These features are discussed in more detail below in terms of event-mean quantities, which are representative of the run-averaged downdraught properties.

Conditionally sampled means, i.e. those calculated using data from within events only, are denoted \( \langle \cdot \rangle \). These include only data from event cores (where spatial gradients
are least, as shown in Fig. 6) by excluding the two points adjacent to the event boundaries, i.e.

\[ \langle s' \rangle = \sum_{\text{all events}} \sum_{i=1}^{4} s'_i / 8. \] (6)

The conditionally sampled variance or covariance uses data at full resolution between the same event boundaries.

Note that the \( \bar{q}'_i \) data in Fig. 6 are slightly out of phase with the rest. This is a general problem related to the limited response time of the Johnson–Williams liquid water sensor. A clearer demonstration of this problem is shown in Fig. 7, which shows normalized conditional mean values (\( \sum \langle s_i' / \langle s' \rangle \rangle / m \)) calculated from the \( m \) runs on which the largest values of \( \langle s' \rangle \) were detected (\( m \approx 20 \)). The lag in the \( q'_i \) data determined from the correlation with the \( w' \) data shown in the same figure corresponds to a mean delay time of a few tenths of a second. This is consistent with the known response characteristics of this probe (e.g. Personne et al. 1982). The definition of \( \langle q'_i \rangle \) (see Eq. (6)) is therefore lagged by two points to take some account of this. A lag–correlation analysis of the \( w' \) and \( q'_i \) data shown in Fig. 7 suggests that this deficiency may cause measurements of the \( \bar{wq}_i \) flux due to downdraughts obtained by direct correlation to be underestimated by about 10% on average. Using results presented in section 3(e) below, this translates to an underestimation of about 6% in the total liquid water flux.

Figure 7. Scaled, composite conditionally sampled data. The standard deviation of the mean is shaded (it is too small to detect for \( w' \)).
Figure 8. Variation of run-mean intersected event width, $\bar{d}$, number, $N$, and area covered by downdraughts, $a$, vs. $z'/h$. Symbols as in Fig. 2.

(b) Event size and frequency statistics

Figure 8 shows how the mean intersected event width, $\bar{d}$, the mean number of events detected per unit length, $N$, and the fractional area occupied by events, $a$, vary with altitude when scaled with mixed layer depth. The events are assumed to be randomly distributed in space, so that $a = N\bar{d}$. The cellular patterns observed at cloud top, and the absence of any differences between runs oriented perpendicular and parallel to the wind direction, suggest this assumption is reasonable.

Four of the flights display remarkably similar behaviour, although differences in case 624 again become apparent as depth increases. In all cases, events occupy their maximum area of 37% close to cloud top (at $z' = 0.1h$), where their mean width is about 0.2h and where the largest number of intersections per unit distance is found. At lower levels, the area occupied decreases steadily to about 20% near the mixed layer base, while the mean event width increases by a factor of two. Thus, the encounter frequency decreases strongly, by a factor of 3–4. In contrast, $\bar{d}$ decreases towards the lower boundary in case 624 and the reduction in the numbers encountered is less marked.

These statistics were found to be relatively little affected by variations in $w_{\text{thres}}$, probably because $d$ is defined in terms of zero-crossing points, and does not depend on the precise value of $w_{\text{thres}}$. However, $N$, and therefore also $\bar{d}$, are fairly sensitive to the choice of minimum acceptable event size. This is especially true near cloud top where there are many small events per unit length (see e.g. Fig. 9 below). The fractional area occupied by events, $a$, and the conditionally averaged thermodynamic statistics, are largely unaffected by either change.

By way of comparison, both Lenschow and Stephens (LS, 1980) and Khalsa and Greenhut (KG, 1985) have carried out similar analyses in convective boundary layers heated from below by a relatively warm ocean, although the criteria adopted to define an event were somewhat different in each case. Unfortunately, given the sensitivity of certain results to particular details of the criteria employed, only a limited comparison is possible. Nevertheless, some broad features do stand out.

LS used specific humidity to define events which were accepted if a threshold ($q' > \sigma_q/2$) was exceeded and provided $d > 25$ m. KG used a procedure more like that used here, where events were accepted if $w'$ exceeded a threshold set at approximately $0.5w_*$, provided $d > 41$ m. In both investigations, the fraction of the total record occupied by updraughts was found to be significantly less than that seen in Fig. 8. KG reported 15% while LS gave values of 20–30%. The mean event widths were also much smaller
in both cases at all levels, e.g. at \( z = 0.2h \), \( \bar{d} \sim 0.15h \) (KG) or \( \sim 0.09h \) (LS), while Fig. 8 shows \( d \) at the equivalent level to be \( \sim 0.25h \). The numbers intersected per unit distance lie between those given by KG and LS. If these data are typical, this would imply that convection produced by radiative cooling in stratocumulus has larger convective elements, which occupy a greater fractional area than that driven by surface heating.

Viewed from high level, these clouds invariably showed regular, closed cellular patterns in the cloud tops with an aspect ratio (longest-diameter/shortest-diameter) near one (e.g. see NL), suggesting broad updraughts in the centres of cells with descent in narrow regions around the edges. If downdraught regions of width \( D \) are located around the periphery of quasi-circular cells of radius \( R \), then \( a \approx D/(R + D/2) \), by simple geometry, and an array of cells of diameter \( 5D \) separated by zones of width \( D \) would be needed to give \( a \approx 1/3 \), as observed. The measured distributions of event widths encountered near cloud top (0.05 < \( z'/h \) < 0.2) shown in Fig. 9 are in fact consistent with random intersections of narrow downdraughts. Monte Carlo calculations show that many distributions of randomly oriented, elongated elements give a reasonable approximation to the variation seen in Fig. 9. As an example, the expected distribution of intersected widths for the regular hexagonal array shown in the inset is also a reasonable match if \( y_1 = 0.6h \) and \( y_1/y_2 = 5 \). The numbers of larger events encountered in the measurements may thus be largely explained by oblique intersections of relatively narrow areas, although the perfect regularity of the chosen example precludes agreement for the smallest sizes. This example also yields \( a = 0.35 \) and \( \bar{d} = 0.23h \), which are very close to the observed values. While it is impossible to be conclusive about the geometry of convective patterns without simultaneous observations from above cloud top, this does show that the results in Figs. 8 and 9 are at least consistent with the existence of a cellular structure with mean diameter of \( \sim 0.5h \) to \( 0.75h \), in which broad updraughts are separated by downdraughts in zones 0.1h to 0.15h wide.

On any flight except 624, the frequency with which the smaller (\( d < 0.5h \)) events are detected progressively decreases at lower levels, while the number of larger events increases. It is these changes which lie behind the behaviour of the statistics shown in Fig. 8. These variations with height are compatible with a number of different interpretations:

![Figure 9](image)

**Figure 9.** Distributions of intersected event width measured near cloud top (0.05 < \( z'/h \) < 0.2). The curve is the expected distribution for random intersections through the regular hexagonal array shown in the inset, with \( y_1 = 0.6h \) and \( y_2 = 0.48h \).
downward moving convective elements may expand with only a fraction of them retaining
the necessary event velocity signature after moving through a given depth; elements may
merge; or there may be a change in their horizontal cross-sectional plan-form. It is likely
that all three occur. In case 624 the findings are reversed and there are far fewer large
events at lower levels than above. This is reflected in the different statistics in Fig. 8 and
is perhaps a result of the more active updraughts in this case breaking up downdraughts
into smaller areas.

(c) Conditionally averaged statistics: means

The conditionally sampled event values \( \langle w' \rangle \), \( \langle T'_v \rangle \) and \( \langle q'_r \rangle \) are shown in Fig. 10. Only a few \( T'_v \) and \( q'_r \) measurements were possible on flight 624, all located below cloud base.

In four cases, the maximum values of \( \langle w' \rangle/w_* \) are found near \( z' = 0.2h \), where \( \langle w' \rangle = -0.75w_* \), but in case 624, there is little variation with height about a value of
\( -0.6w_* \). The possible range through which \( \langle w' \rangle \) can vary is constrained because the
method used to define events includes a threshold at \(-0.5w_* \), so most of the vertical
changes are seen in the event size and frequency statistics. The mean value of \( w' \) in the
environment outside events is governed by continuity. If downdraughts occupy a frac-
tional area \( a \), the environment must be moving upwards with a mean velocity given by
\( a\langle w' \rangle/(a-1) \), or about 0.4\( \langle w' \rangle \). This is close to the values found in the ‘tail’ regions (e.g.
see Fig. 7), suggesting that these regions are not untypical of the remainder of the
environment and showing that the total velocity difference across event boundaries is
significantly greater than \( \langle w' \rangle \) and may exceed \( w_* \).

The thermals detected by IS were also found to have the maximum updraught velocity at an equivalent level \( (z = 0.2h) \), although the value was less than that found here \( (0.45w_*) \). In contrast, KG reported little variation from a value of \( 1.1w_* \) albeit through a much more restricted height range \( (0.1h < z < 0.3h) \). These differences may
reflect the different event selection criteria employed, although the updraught mass flux (defined as the product of the fractional area covered and \( \langle w' \rangle/w_* \)) was similar in both cases, \( 0.12-0.16 \) near \( z = 0.1h \). The corresponding downdraught mass flux found here is
significantly larger, about 0.25 at \( z' = 0.1h \). This could indicate that a cloud top interface
is better ventilated than a much rougher ground or sea surface because of the greatly
reduced stress at the boundary.

On average, the downdraughts are driest and most negatively buoyant close to cloud
top (Fig. 10). These differences decrease with depth, becoming zero approximately
midway through the mixed layer. Below this, events tend to be slightly positively buoyant

Figure 10. Run-averaged conditional mean data v. \( z'/h \). Symbols as in Fig. 2.
TABLE 2. VALUES OF TERMS USED IN DISCUSSING DOWNDRAUGHT BUOYANCY

<table>
<thead>
<tr>
<th>Flight No.</th>
<th>(-\langle q' T' \rangle / q_{*T} )</th>
<th>(-100 \Delta q_T / q_{*T} )</th>
<th>(\delta ) (%)</th>
<th>(\delta T_{V*} ) (K)</th>
<th>(-2T_{V*} ) (K)</th>
</tr>
</thead>
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<td>511</td>
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</tr>
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<td>0.45</td>
<td>0.7</td>
<td>-0.002</td>
<td>-0.048</td>
</tr>
<tr>
<td>620</td>
<td>1.5</td>
<td>1.1</td>
<td>1.6</td>
<td>0.031</td>
<td>-0.050</td>
</tr>
<tr>
<td>624</td>
<td>1.5</td>
<td>1.3</td>
<td>1.9</td>
<td>0.031</td>
<td>-0.072</td>
</tr>
</tbody>
</table>

although there are no discernible differences in water content. Both \(\langle T'_{V*} \rangle / T_{V*} \) and \(\langle q' T' \rangle / q_{*T} \) are smaller than the corresponding values reported by LS and KG, by up to 50% at \(z' = 0.1h \). This is consistent with the larger mass flux: increased ventilation near cloud top also implies that equivalent heat or water fluxes may be carried by convective elements with reduced temperature or water deficits.

Since \(q_{*T} \) is conserved during mixing with air from above cloud top (the effects of gravitational settling are negligible on small timescales), the ratio \(\delta = \langle q' T' \rangle / \Delta q_{T} \) is a measure of the average mixing fraction of inversion air inside downdraughts. If \(\langle q' T' \rangle / q_{*T} = -1.5 \) at \(z' = 0.1h \) (see Fig. 10), then \(\delta \) varies between 0.7% and 3%, as shown in Table 2. Mass conservation implies an entrainment velocity \(w_{e} = a\delta(w') \). Evaluating this expression using data from Figs. 8 and 10 shows that 0.2% < \(w_{e}/w_{*} \) < 0.8%, which is consistent with the range published in NL which was obtained using a different method. The downdraughts therefore contain, on average, only a small proportion of air originating from above cloud top. If all this entrained air has been brought to water saturation, the effect on the net buoyancy of the downdraughts may be calculated, since \(\delta \) is known (see TN). Evaporative cooling of the entrained air will reduce the net potential temperature increase. It is found that completely mixing the fractions of inversion air (\(\delta \) quoted in Table 2) into the downdraughts would increase their mean virtual temperature excess, \(\delta T_{V} \), by amounts varying from -0.002 K to 0.05 K (Table 2). However, the observed values of \(\langle T'_{V*} \rangle \) (see Fig. 10) are all significantly negative at this level, about -2\(T_{V*} \), or -0.04 K to -0.07 K (Table 2). The difference must be due to the incorporation of radiatively cooled, cloudy air. Inspection of time series data (e.g. see Fig. 1(a) in TN) shows parcels of air within downdraughts that have been cooled by at least twice the 0.1 K required. This could be achieved by a few minutes exposure to the typical cooling rates of several degrees per hour found within a few tens of metres of cloud top (NL). Using the simple geometry shown in Fig. 9, the mean, area-weighted, horizontal path between up- and downdraught regions is found to be ~0.2h. A radial velocity of \(w_{*} \) (consistent with Figs. 3 and 10) would imply a typical timescale ~150 s in the strongly cooled region, sufficient to produce the required effect.

The incorporation of radiatively cooled, cloudy air into the downdraughts is the primary mechanism by which an overall negative buoyancy is produced. Entrainment, mixing and evaporative cooling can produce only positive buoyancy fluctuations in all cases except one, where a small additional negative contribution is implied.

The level at which the downdraught velocities are largest does not coincide with the level at which the buoyancy excess becomes zero, but occurs somewhat nearer cloud top. Figure 11 compares two of the downward acceleration terms ((A) may be rewritten as shown in the figure)

\[
\frac{-h}{w_{*}^2} \frac{d(w')}{dz'} \quad \text{(A)} \quad \quad \quad \frac{h g}{w_{*} T_{V*} \langle T'_{V} \rangle} = \frac{\langle T'_{V} \rangle}{T_{V*}} \quad \text{(B)}
\]
calculated from curves fitted to the data shown in Fig. 10 (excluding case 624). In the upper half of the mixed layer, the observed accelerations (term (A)) are smaller than would be expected if buoyancy (term (B)) was acting alone. LS also found very similar behaviour for thermals above a heated surface. Non-hydrostatic vertical pressure gradients and momentum exchange by entrainment of air from outside the downdraughts could both account for this difference. The entrainment of cloudy air into downdraughts is also consistent with the observed increase of $\tilde{d}$ with depth (Fig. 8).

The in-cloud event-mean specific humidity and temperature values are derived from independent measurements and are consistent with the air being close to saturation, as shown in Fig. 12. This is to be expected in these conditions, where droplet growth is sufficiently responsive to prevent the development of significant mean supersaturations. However, as discussed in the next section, there are uncorrelated temperature and humidity fluctuations on smaller scales within events which are large compared with the mean (hence the scatter in the values in Fig. 12), implying that transient supersaturation fluctuations are always present, although these are always likely to remain small given the rapid response time of small droplets to changes in saturation in maritime conditions.

Figure 12. As Fig. 10. The approximate supersaturation scale assumes that $q = 5$ g/kg.
(d) Conditionally averaged statistics: variances

A measure of the variance attributable to downdraughts relative to that for the run as a whole is given by the 'effectiveness ratio':

\[ r(s) = \frac{\langle s'^2 \rangle}{\bar{s}'^2}. \]  

(7)

A value of \( r = 1 \) means that downdraughts contain on average no greater variance than the run as a whole, while if \( r > 1 \), the downdraughts contribute proportionally higher levels. The data plotted in Fig. 13 show that \( r > 1 \) on most runs, so downdraughts are generally identified with higher variance levels than the run average, especially for \( w' \).

If the fractional area occupied by downdraughts is \( a \), then the ratio of variance contributed per unit length inside and outside events is given by

\[ R(s) = \frac{\text{variance inside events}}{\text{variance outside events}} = \frac{r(s)(1-a)}{r(s)(1-ar(s))}. \]  

(8)

At \( z' = 0.2h \), where \( a \approx 0.35 \) (see Fig. 8), \( r(w) \approx r(T) \approx r(q) \approx 1.5 \) and thus \( R \approx 2 \). Downdraughts therefore contribute approximately twice as much variance per unit length as the remainder of the run for \( w' \), \( T' \) and \( q' \) at this level. In their study, LS found similar values, with \( R(w) \approx 1.8 \) near the surface (\( z < 0.15h \)). Figure 13 also shows that \( r(w) \) has a tendency to increase with depth (except perhaps on flight 624), whereas \( r(q, T \text{ and } q) \) all show a reversed trend with maximum values near cloud top. Neither \( r(u) \) nor \( r(v) \) vary significantly with \( z'/h \). For an explanation of these differences, it is instructive to look more closely at the conditionally sampled variance.

That fraction of \( \langle s'^2 \rangle \) attributable to the distribution of conditionally sampled data having a mean value different from the run as a whole is given by

\[ g(s) = \frac{\langle s' \rangle^2}{\langle s'^2 \rangle}. \]  

(9)

If \( g \approx 1 \), most of the variance within events arises because the mean of the conditionally sampled data, \( \langle s' \rangle \), is significantly different from the mean found on the run as a whole (which is zero by definition). If \( g \approx 0 \), most of the variance within events arises from fluctuations about \( \langle s' \rangle \). Thus \( \langle s'^2 \rangle = \langle s' \rangle^2 + \sigma_s^2 \), where \( \sigma_s^2 \) is the conditionally sampled variance about a mean value of \( \langle s' \rangle \), and \( g(s) \) is directly related to the dispersion \( \langle \sigma_s/\langle s' \rangle \rangle \) of this distribution:

\[ \sigma_s/\langle s' \rangle = (1/g(s) - 1)^{1/2}. \]  

(10)

The values of \( g \) plotted in Fig. 13 reveal a big difference between \( g(w) \) and the other quantities. Almost 80% of \( \langle w'^2 \rangle \) is due to the conditionally sampled data having a different mean, so the dispersion (from Eq. (10)) is about 0.6, i.e. within events, the standard deviation is only about half the mean value, as shown by the error bars in Figs. 6 and 7. This is of course largely forced by the event selection criteria, but shows again that the event signals are well defined. However, because such a large proportion of \( \langle w'^2 \rangle \) is contributed by \( \langle w' \rangle \), vertical velocity fluctuations about \( \langle w' \rangle \) within downdraughts are actually considerably smaller than those outside. The ratio of \( \sigma_w^2 \) to the level of variance outside events is simply given by \( (1 - g(w))R(w) \), which Figs. 8 and 13 shows is about 0.4 at \( z' = 0.2h \). (It should be noted that instrumental noise may also influence the values of \( r \) and \( g \), although filtering applied during processing has avoided any significant effect being introduced into these data.)

In contrast, \( g \) is found to be much smaller for the other quantities. Here, the increased variance associated with events is not due to the different mean of the conditionally sampled data, but to larger fluctuations encountered within events. Although the conditionally sampled mean values of these quantities are significant and
Figure 13. Quantities $r(x)$ and $g(x)$ v. $z'/h$. $r(q_i)$ is plotted only for in-cloud runs. Data from different flights are not distinguished, for clarity.
vary in a repeatable manner (see Fig. 10), they are small in comparison with the fluctuations, so the dispersions are very large. For example, the largest value of \( g(T) \) occurs near cloud top and is \( \sim 0 \cdot 3 \), so the dispersion is \( \sim 1 \cdot 5 \). The ratio of variance levels within and outside events, given by \( \{1 - g(T)\}R(T) \), is about 1·4. For \( u \) and \( v \), where the conditional mean values are no different from the run averages, the increased fluctuation levels account for all of the increased variance.

Downdraught properties are therefore not so simple to describe as the conditional mean data shown in Fig. 10 might imply. While downdraught near cloud top are relatively cool and dry on average compared with their surroundings, they also contain regions which may be significantly warmer or wetter than the average value prevailing at that level. Indeed, fluctuations within downdraughts are larger than those elsewhere at the same level for all quantities except \( w' \), although this contrast decreases with distance downwards from cloud top (Fig. 13). Downdraughts therefore contain both positively and negatively buoyant pockets (and probably also saturated and sub-saturated pockets), especially near cloud top. Further down into the cloud layer, mixing within downdraughts and entrainment across event boundaries progressively reduce fluctuation levels and diminish downdraught–environment differences.

The relatively large size of the buoyancy fluctuations within events compared with the event-mean buoyancy deficit seems less consistent with the idea that downdraughts are formed simply as a result of the most negatively buoyant regions sinking, than as a consequence of the horizontal convergence of motions close to the density interface sweeping up air from the vicinity of cloud top. Whatever the mechanism, radiative cooling and mixing with drier air above cloud will cause air from this region to be slightly cooler and drier than average if transported downwards. Furthermore, Figs. 3 and 4 show that increased (or decreased in the case of \( w \)) fluctuation levels observed within events are consistent with down-gradient variance transport.

\((e)\) Conditionally averaged statistics: covariances

The fraction of the total run-averaged covariance contributed by downdraughts is \( \hat{a}(w's')/\hat{w}'s' \) and the proportion of this attributable to the event means being different from the run average (see Eq. (7)) is given by \( \langle w' \rangle (s') \langle w's' \rangle \).

Table 3 gives the average values obtained when all the runs in the interval \( 0 < z'/h < 0 \cdot 3 \) are taken together. The figures in brackets are standard deviations. The covariances at this level tend to be relatively large (see Fig. 2) and all runs are in cloud. Lower down, many of the covariances are near zero (see Fig. 2) and the ratios become indeterminate.

The downdraughts are found to contribute over half of the total fluxes in the upper part of the cloud layer despite occupying only an average 0·35 of the horizontal area (an ‘effectiveness ratio’ of 1·6). However, unlike the variance, Table 3 shows that the major part of this contribution is due to the data within events having different mean values from their surroundings. The covariance due to correlations between fluctuations within events is small in comparison.

<table>
<thead>
<tr>
<th>( \hat{a}(w's')/\hat{w}'s' )</th>
<th>( \hat{T} )</th>
<th>( w' )</th>
<th>( \hat{T} )</th>
<th>( \hat{w}' )</th>
<th>( \hat{T} )</th>
<th>( \hat{T} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \langle w' \rangle (s') \langle w's' \rangle )</td>
<td>1·6</td>
<td>1·6</td>
<td>1·6</td>
<td>1·6</td>
<td>1·6</td>
<td></td>
</tr>
<tr>
<td>Effectiveness ratio</td>
<td>1·6</td>
<td>1·6</td>
<td>1·6</td>
<td>1·6</td>
<td>1·6</td>
<td></td>
</tr>
</tbody>
</table>

TABLE 3. Covariance statistics in the upper cloud layer
4. Velocity Spectra

To investigate further the variations seen in the preceding sections, Fourier spectral estimates were calculated from the velocity fluctuation data using successive 4096-point records. The spectral estimates from each record were band-averaged in six logarithmically spaced intervals per decade and the final results were obtained by averaging all such spectra from every run made at the same level.

The vertical velocity spectra, shown in Fig. 14, show the greatest range of variation. They have been grouped into three non-dimensional height bands and have been scaled to bring the high frequency parts into coincidence. Estimates of viscous dissipation, \( \varepsilon \), (see section 5) were derived from this scaling factor using inertial subrange theory by assuming a value for the Kolmogorov constant, as described in Nicholls and Readings (1981) (see also the appendix).

The mixed layer scaling renders the spectra from the different flights quite similar and the reduction of the dominant length scale (represented by the wavelength \( \lambda_m \) at which the spectral peak is located) towards cloud top is clearly evident. These spectra are very similar in appearance to many previously published vertical velocity spectra obtained above a heated land surface (e.g. Kaimal et al. 1976), provided \( z' \) is replaced by \( z' \). The expression given by Kaimal et al. for the vertical variation of \( \lambda_m \) also fits these data well, as shown in Fig. 15.

Corresponding spectra from levels outside or very close to the mixed layer boundaries (0.03 < \( z'/h \) or \( z'/h > 1 \)) are quite different. The peaked shape disappears and there is no well defined maximum nor any \( -2/3 \) slope evident at the shorter wavelengths. This reflects the damping influence of the stable density stratification outside the mixed layer on vertical motion.

![Figure 14. Scaled vertical velocity spectra \( n \) dimensionless wavelength, \( h/\lambda \), for three dimensionless height bands.](image)
The horizontal velocity spectra show much less variation with height and display considerably more energy at longer wavelengths than the vertical velocity spectra. Again, this is typical of results from bottom-heated convective boundary layers (e.g. Kaimal et al.). However, one interesting feature is found in these data close to cloud top, where an increase in the horizontal velocity variances is often observed (see Fig. 3). While there is a small increase at $h/\lambda = 2$, where the vertical velocity spectral peak is located, most of the additional horizontal variance occurs on much larger scales, in the interval $0.1 < h/\lambda < 1.0$. This probably reflects the cellular nature of the convection referred to earlier. In a simple closed cellular flow of the type envisaged in section 3(a), the horizontal velocities near the upper boundary will vary on the scale of the cell diameter (in the absence of significant mean shear) rather than on the scale of the main vertical velocity fluctuations, which are concentrated in narrow regions around cell boundaries. Since the horizontal and vertical velocity spectral peaks referred to above are separated, on average, by a factor of 5–6 in wavelength, this is consistent with the cell and downdraught dimensions suggested in section 3(a).

5. THE TURBULENT KINETIC ENERGY BALANCE

The balance between the production and dissipation of turbulent kinetic energy as a function of height may be expressed in the form

$$\frac{h}{w^3} \frac{\partial E}{\partial t} = \frac{gh}{T_v w^3} + h \frac{\partial}{\partial z} \left( \frac{wE}{w^3} + \frac{wp}{w^3 \rho} \right) + \frac{uw}{w^3} h \frac{\partial U}{\partial z'} + \frac{vw}{w^3} h \frac{\partial V}{\partial z'} - \frac{h e}{w^3}$$

(11)

\begin{align*}
Q & B & T & P & S & D
\end{align*}

(e.g. Nicholls 1984) where $B$, $S$, $T$, $P$ and $D$ are commonly known as the buoyancy and shear production terms, turbulent and pressure transport terms and the dissipation respectively. All except $P$ can, in principle, be estimated from the observations and have been evaluated as described in Nicholls (1985). The nature of this balance is an important controlling influence on internal mixing processes and reveals how TKE is transported from source regions which possess excess TKE to regions with a net deficit.

Figure 16 shows the vertical distribution of $wE/w^3$, $B$ and $D$ for each of the cases (data from similar levels have been averaged together). The variation of each of these terms is broadly repeatable from case to case, although some variations are apparent.
First, the similarities are considered. Ensemble-average values of \( B, D \) and \( T \) were obtained by interpolating the curves in Fig. 16 at regular height intervals, combining all the cases together and differentiating \( \frac{wE}{w_2} \). These are shown in Fig. 17(a). The remaining terms \( P + S - Q \) and any measurement errors combine to produce the residual \( I \). The measured momentum fluxes and the mean velocity gradients were used to estimate \( S \) but the values were found to be relatively small \((-0.2 < S < 0.2)\) at all levels and no systematic vertical variation was detected. The rate-of-change terms, \( Q \), determined from repeated measurements at the same height (NB. these were only performed at the upper levels, \( z' < 0.5h \)) were even smaller. Consequently, if systematic measurement errors are small, it is reasonable to associate the larger values of \( I \) with a significant pressure correlation term (at least in the upper half of the layer).

The size of measurement errors likely to be incurred in this analysis is difficult to assess, although there are reasons for believing that such errors are not serious. The data shown in Fig. 16 appear to be good enough to detect consistent differences between cases (see also below), and the imbalance term is similar in each case taken separately. The method of interpolating linearly between neighbouring flight levels is more likely to underestimate the gradients in \( B \) and \( \frac{wE}{w_2} \) at cloud top (see Fig. 16), both of which would cause \( I \) (i.e. \( P \)) to be underestimated. Thus \( P \) might be even larger at cloud top than is shown in Fig. 17(a).

Figure 17(a) shows three regions, each with a broadly different balance. In the upper part of the layer, roughly where \( z' < 0.25h \), \( B \) is large and positive while \( T \) and \( D \) are negative: TKE generated by buoyancy produced by radiative cooling is partially exported downwards by accelerating downdraughts, and partially dissipated locally. However, \( B \) alone is insufficient to balance the losses represented by \( T \) and \( D \), so \( I \) is positive, especially near cloud top. This suggests that \( P \) is a major source term close to the cloud top interface \((0 < z'/h < 0.1h)\) where vertical motion is constrained by the inversion, and energy is redistributed between the vertical and horizontal components by the pressure field. The transport term \( T \) also removes TKE from this region. This relationship between \( P \) and \( T \) is consistent with the description of motion in this region given earlier, where horizontal motions result from the spreading out of updraughts forming active downdraughts at convergence zones. \( B \) is of secondary importance in this region and reaches its maximum slightly lower down, where \( z' = 0.1h \) to 0.2h.
All four terms are roughly equivalent in size in the interval $0.25 < z'/h < 0.5$. Although $B$ decreases, transport of TKE (from the layer above) also becomes a major source term. This implies that $P$ changes sign. This general behaviour is observed in every case, as can be deduced from the data plotted in Fig. 16, and is related to the observation that downdraughts are decelerating through this region (see Fig. 10).

When $z' > 0.5h$, all the terms are considerably smaller on average, making interpretation less reliable, and there are considerable differences between cases. The balance here is clearly sensitive to residual transport from above and to fairly small variations in $B$.

On two occasions (511 and 526) $B$ is negative (for reasons discussed in NL) and the transport term remains small. These are also the cases with the smallest dissipation. However, either $P$ or $S$ must provide a source term or $Q$ must be negative. As the values needed for balance are relatively small, it is not possible to discriminate between these alternatives using the observations. If the imbalance was entirely accounted for by $E$ decreasing with time (i.e. assuming $P = S = 0$) then $Q = -0.5$. Since $(kE/w^3) \sim 500$ s, Eq. (11) shows that such a situation would imply large reductions in $E$ occurring in only a few tens of minutes. This is essentially the mechanism envisaged by Nicholls (1984) which ultimately results in the decoupling or shallowing of the original mixed layer.

Conversely, in case 528, which NL characterized as a turbulent layer deepening by entrainment at both the upper and lower boundaries, both $B$ and $T$ are significantly larger where they were measured below $z' = 0.5h$. Although $D$ is also correspondingly greater at these levels, there does appear to be a slight excess of production over dissipation in this region in this case, which could support entrainment at the lower boundary.

Unfortunately, there are no other observations currently available with which to compare these results. The closest relevant data are those from convective layers over heated surfaces (e.g. Lenschow et al. 1980) and large-eddy simulations (Moeng 1986), although they are not directly equivalent and only limited comparisons are possible. Lenschow et al. obtained their data in cold air outbreaks over a relatively warm ocean surface (plotted upside down in Fig. 17(b) for ease of comparison). The residual term,
is again associated with the unmeasured term, \( P \), since their results are also derived from observations. The resolved-scale pressure correlation term is calculated explicitly in Moeng’s large-eddy simulation of a stratus layer with strong cloud top radiative cooling and a near-zero surface buoyancy flux (Fig. 17(c)). The scaling used in both sets of results is equivalent to that used in Fig. 17(a). (NB. the definition employed by Moeng does not make a significant difference in this case.)

A comparison of Figs. 17(a) and (b) shows many points of similarity. Although the proximity of the surface in Lenschow’s data causes major differences when \( z' < 0.1h \) where \( S \) is large, \( T, B \) and \( D \) are all similar in the region \( 0.1 < z'/h < 0.3 \). \( B \) reduces more quickly once \( z' > 0.3h \) in the cloudy case owing to radiative and phase change effects, although this is compensated by an increase in \( T \) (provided \( z' < 0.5h \)). At lower levels in the cloud-topped case, \( B \) becomes negative and consequently \( D \) is reduced to approximately half that seen in Fig. 17(b). The imbalance term changes sign in the same way in both sets of results, but appears to be larger in the cloudy case, especially close to cloud top.

Comparing \( P \) from Moeng’s results in Fig. 17(c) with \( I \) in Fig. 17(a) shows they are fairly similar in the region \( z' < 0.5h \), especially near cloud top. However, the behaviour of \( T \) is quite different. The simulated values are small near cloud top (large negative in Fig. 17(a)) and have different signs when \( 0.3 < z'/h < 0.5 \). As a result \( D \) in Fig. 17(c) is much larger than observed below cloud top in Fig. 17(a), while \( B \) is larger beneath. The comparison is invalid below \( z' = 0.6h \) because the presence of the sea surface at the model lower boundary becomes increasingly important. The differences between the modelled variation of \( T \) and the observations appear to be related to the behaviour of \( wT^3 \) below cloud top. The measurements are strongly negative but the simulated values are not. While this discrepancy might reflect differences between the conditions under which the observations were made and those used in the simulation, e.g. the ratio of buoyancy generated by radiative cloud top cooling to latent heating at cloud base, further comparisons between the structure of the simulated convective motion and observations would appear to be desirable. In particular, the sensitivity of the modelled convection to changes in the boundary conditions within an envelope encompassing the observed conditions would be of interest, especially as this aspect is very difficult to assess from a limited series of case studies.

6. Conclusions

Data from five research flights in stratocumulus-capped boundary layers have been examined to reveal the structure of the convective motions which result from buoyancy fluctuations caused (primarily) by cloud top radiative cooling. In four cases, the convectively mixed layer was detached from the surface, with no buoyancy input through the bottom boundary. The largest buoyancy fluxes were located close to cloud top, the net latent heat release at cloud base being small by comparison. In the other case (624) the surface was the bottom boundary and the buoyancy and water vapour fluxes were significant. Consequently, the buoyancy flux maximum was broader and occupied a larger fraction of the upper mixed layer. However, all the cases can be categorized as free convection (shear production was negligible) driven primarily from cloud top.

Mixed layer scaling shows that the overall levels of variance are very similar to those found in other convective layers when account is taken of the different geometry. The same is true of the velocity spectra.
The primary convective elements in the cloud layer are negatively buoyant downdraughts. These are well defined in terms of their vertical velocity signature, even very close to cloud top \((z' < 0.1h)\), which was used to define a conditional sampling criterion. The properties of these downdraughts or 'events' are very similar from case to case when appropriately scaled. They occupy a maximum fractional area of about 0.37 just below cloud top. The average intersected event size (width) increases and the encounter frequency decreases with distance below cloud top \((z')\). The distribution of intersected widths near cloud top is consistent with observed cellular patterns of convection if downdraughts occupy relatively narrow regions \((\sim 0.1h-0.15h \text{ wide})\) around the periphery of larger-area updraughts \((\text{diameter } \sim 0.5h-0.75h)\). The scale separation of peaks in the horizontal and vertical velocity spectra measured close to cloud top is also consistent with this arrangement.

On average, the conditionally sampled downdraughts are both cooler and drier than their surroundings in the upper part of the mixed layer. These contrasts are greatest near cloud top, decreasing to around zero near \(z' = 0.5h\), as downdraughts mix with surrounding cloud. At lower levels, the downdraughts have slightly positive buoyancy. The differences show that downdraughts near cloud top contain both air which has been radiatively cooled and dry air entrained from above cloud top which has been cooled and moistened by evaporation. However, evaporative cooling alone can produce negative buoyancy in only one case. Radiative cooling must dominate in every case if the observed mean buoyancy deficit of downdraughts is to be explained. The degree of cooling required is consistent with radiative transfer calculations and simple circulation models.

The differences between events and their surroundings are smaller than corresponding results previously reported over heated land or sea surfaces, but this is compensated by a mass flux which is significantly greater. This is largely due to the greater fractional area occupied by the downdraughts and suggests that the cloud top interface is ventilated more effectively than a land or sea surface. The downdraughts are efficient transport mechanisms and are responsible for over half the heat, water vapour and liquid fluxes in cloud, despite occupying well under half the area.

The downdraughts transport variance down the mean gradients, so levels of horizontal velocity, temperature and total humidity mixing ratio are all enhanced within events, while \(w'^2\) is reduced. Fluctuation levels within events are large compared with the mean deviation for all quantities except \(w\), so, although the events are relatively cool and dry on average, they may still contain regions which are relatively moist and positively buoyant. The mean buoyancy deficit of the downdraughts is more than sufficient to explain their downward acceleration, but the poor correlation between \(w'\) and \(T'V\) within events suggests that the well defined event vertical velocity signature (especially near cloud top) results from the horizontal convergence of circulations constrained by the overlying inversion rather than direct buoyant instability at cloud top. In this interpretation, horizontal circulations formed by updraughts spreading out under the inversion scour the cloud top interface, incorporating air which has been radiatively cooled and/or partially mixed with inversion air before being forced down in well defined, narrow convergence zones. These downdraughts subsequently accelerate because of their net negative buoyancy, mixing internally and with the surrounding cloud. This downward mass flux causes compensating updraughts to rise until they near the inversion, where they are forced to spread out.

This interpretation is consistent with the TKE balance. Measurements close to cloud top \((0 < z'/h < 0.1)\) imply that the pressure correlation must be a significant source term since the only other positive contribution, from buoyancy generation, is insufficient to match the losses by transport and dissipation. Buoyancy provides the main source term
beneath this \((0.1 < z'/h < 0.5)\) and the pressure and transport terms change sign. These variations are consistent with the observed properties of downdraughts and the associated circulations. In the lower halves of the mixed layers, all terms are smaller and the balance is much more sensitive to the particular circumstances of each case.

**ACKNOWLEDGEMENTS**

I would like to thank Jon Leighton, who wrote much of the analysis software and ran many of the programs, as did Julie Sweetingham and Stephanie Wolohan. The comments of Dr D. H. Lenschow and another referee also improved the final version of the paper.

**APPENDIX**

*Notation not explicitly defined in the text*

\(g\) Acceleration due to gravity
\(h\) Mixed layer depth (see section 2(a))
\(p\) Pressure
\(q\) Specific humidity
\(q_l\) Specific liquid water content
\(q_T\) Total water content \((=q+q_l)\)
\(T_v\) Virtual temperature
\(U, V\) Mean streamwise and lateral horizontal wind components
\(u, v, w\) Fluctuating streamwise, lateral and vertical wind components
\(z, z'\) Height above mean sea level, height measured down from cloud top (see NL for details)
\(\varepsilon\) Rate of dissipation of E
\(\phi_w\) A spectral scaling parameter. By assuming a value \((=0.5)\) for the Kolmogorov constant, \(\varepsilon\) may be determined (e.g. see Nicholls and Readings 1981)
\(E\) Turbulent kinetic energy \((=\frac{1}{2}(u^2+v^2+w^2))\)
\(\lambda\) Wavelength, defined as aircraft true airspeed divided by \(f\), the measurement frequency
\(\rho\) Air density
\(\theta_e\) Equivalent potential temperature (defined as in NL)
\(\Delta s\) Cloud top jump, i.e. the value of \(s\) in undisturbed inversion air just above cloud top, minus the value just below cloud top

An overbar represents an average taken over a (nominal) 60 km measurement run. Fluctuation data are derived by removing the mean and linear trend. Where these data have additionally undergone low frequency filtering (see section 5(a)), this is denoted by a prime.

**REFERENCES**


Nicholls, S. and Leighton, J. R. NL 1986 Aircraft observations of the Ekman layer made during JASIN. *ibid.*, 111, 391–426


