Precipitation in marine cumulus and stratocumulus. 
Part I: Thermodynamic and dynamic observations of closed cell circulations and cumulus bands

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Abstract

A case study of vigorous drizzle development in marine boundary layer clouds is presented. The clouds were observed using a research aircraft west of Tasmania during the Southern Ocean Cloud Experiment. A clean marine boundary layer contained both cumulus clouds (some oriented in bands) and an upper level stratus deck organised in a closed cell circulation.

“Cold pools,’’ that is regions of evaporatively cooled air under mature cumulus clouds, were observed on several occasions. Air in the cold pool centre showed a divergent component of the wind perpendicular to the mature cumulus bands. Ahead of this spreading precipitation downdraft, the wind component perpendicular to the cumulus band was convergent, and new cumulus clouds were subsequently formed over these regions.

This dynamical situation is likened to an actively developing squall line of the propagating type. The development of drizzle, and its evaporation in the mixed-layer below the base of mature cumulus bands, is critical to the cloud evolution. It is demonstrated here, that processes resembling “deep convection’’ may also be responsible for the cloud development in a marine boundary layer of only 1500 m depth. © 2000 Elsevier Science B.V. All rights reserved.

Keywords: Stratocumulus; Precipitation; Cumulus; Boundary layer meteorology

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1. Introduction

Boundary-layer clouds span the entire spectrum between well-mixed stratocumulus clouds and isolated cumulus clouds. Often, stratocumulus and cumulus occur simultaneously. Measurements from the North Sea (Nicholls, 1984), from the sub-tropical Atlantic Ocean (Martin et al., 1995; Rogers et al., 1995; Wang and Lenschow, 1995; Roode and Duynkerke, 1996), from the Southern Ocean (Boers et al., 1997), as well as modelling studies (e.g. Krueger et al., 1995; Wyant et al., 1997) have all examined cumulus clouds rising into overlying stratocumulus.

It is now well recognised that precipitation fluxes from stratocumulus and cumulus in the marine boundary layer are, in some cases, a significant factor in the boundary layer evolution. Yet, observations of drizzle fluxes and the effects on boundary layer structure remain relatively scarce (e.g. Brost et al., 1982; Austin et al., 1995; Bretherton et al., 1995; Boers et al., 1997). The effects of drizzle from boundary layer clouds on the vertical thermodynamic profiles are likely to be comparatively stronger in clean environments, since the development of drizzle is strongest in areas with low cloud condensation nuclei (CCN) concentration, all else being equal. Low concentrations of CCN are a very common feature in the mid latitudes of the Southern Hemisphere. Here, anthropogenic emissions of sulphates are low, and natural sources of CCN will therefore be important in determining the total concentration of CCN.

Measurements of CCN have been made continuously at the Cape Grim Baseline Air Pollution Station on the northwest (NW) coast of Tasmania since 1981 (Gras, 1995). The measurements in clean background air show a seasonal variation with low concentrations in the winter and higher concentrations in summer. The seasonal variation in CCN concentration is thought to be primarily due to the seasonal variation in emissions of sulphur gases from biological production in the ocean (Ayers and Gras, 1991; Ayers et al., 1997). The Southern Ocean Cloud Experiment (SOCEX) was designed to study the seasonal variability in marine boundary-layer clouds.

This paper is the first in a series that focuses on the importance of precipitation in marine boundary layer clouds for a case where significant drizzle from warm stratocumulus and cumulus clouds was observed. Here (Part I), we investigate the importance of the precipitation on the thermodynamic and dynamic evolution of both stratiform clouds and cumulus cloud bands. The analysis of this particular case shows that precipitation is extremely important for the organisation and maintenance of the cloud dynamics.

The synoptic setting and aircraft flight plan are described in Section 2. The observations from two flight stacks, that is sets of horizontal flight legs made at different altitudes, are described in Sections 3 and 4. Section 5 contains a discussion of the convective organisation, the development of cold air pools created by evaporation of precipitation, and the effects of widespread stratiform drizzle on the mixed-layer development. Appendix A contains a discussion of instrumental issues.

Part II of the series (Jensen et al., 2000) examines the microphysical characteristics of both the stratiform and cumulus clouds. That paper contains an analysis of CCN spectra, entrainment, and droplet spectra to evaluate whether giant and ultra-giant nuclei (Johnson, 1982) are essential to the development of the observed drizzle.
Part III (Jensen and Lee, 2000) contains a study of the development of drizzle using a simple combined condensation and stochastic coalescence model. The stochastic model employed is the Gillespie (1973) Monte Carlo model, which keeps track of both the size of drops and their CCN mass during the coalescence development.

2. General observations

The SOCEXs, SOCEX-1 and SOCEX-2, took place over the ocean west (W) of Tasmania in the winter of 1993 and the summer of 1995, respectively. The principal platform was the Australian Flight Test Services F27 research aircraft (formerly CSIRO F27), instrumented by CSIRO and collaborating institutions, with an extensive suite of thermodynamic, liquid water, wind, radiation, CCN, and particle probes. Of particular interest here are the CSIRO cloud liquid water probe (King et al., 1978), a PMS FSSP-100 for cloud drops, and a PMS 260X for measurement of drizzle drops. The aircraft had a variety of temperature sensors installed. Measurement of cloud temperature using immersion sensors is difficult due to possible wetting of the sensing element (e.g. Lenschow and Pennell, 1974; Lawson and Cooper, 1990). This issue is critical for the present analysis. Appendix A details why we feel confident that in-cloud temperature can be measured with little error due to wetting of the Rosemount sensor element on the F27.

2.1. Synoptic setting

On February 1, 1995, Tasmania was under the influence of a large high-pressure system located over the southeast (SE) of Australia (see Fig. 1). This led to a light northwesterly airflow in the area W of the Tasmanian coast and an extensive region of stratiform clouds. The study area, 100 km W of Tasmania, was near the centre of the high-pressure system, and the F27 was located in this area during the period 1030–1300 h; all times are eastern daylight savings time (EDST = UTC + 11 h). Boers et al. (1997) have described a cumulus band that was penetrated during the second half of the flight. The focus of the present study is the interaction between several such cumulus bands and precipitation areas.

The Australian Bureau of Meteorology’s 75-km operational analysis (Mills and Logan, 1994) has been used to calculate 72-h back trajectories for air which, at 1200 h on February 1, was W of Tasmania (see Fig. 1). The air below 850 hPa on February 1 originated from the periphery of Antarctica 3 days prior, at the trailing edge of a deep cold air outbreak. The air at 900 hPa had subsided significantly throughout the previous 3 days (see Fig. 2). The air parcel arriving at 800 hPa had also descended significantly in the preceding 2 days, but had an origin further W in the mid latitudes. Thus, the air from above the boundary layer had a different origin from the air within the boundary layer. Note that none of the trajectories shows any influence of the Australian continent in the 3 days prior to observation (see Fig. 1).

The air closest to the sea surface moved at very high speed (15–18 m s⁻¹) in the 72–36 h prior to observations; this is apparent from the large distance between the 12-h
Fig. 1. Map of 1100 EDST February 1, 1995 mean sea-level isobars (hPa, solid lines) and average weekly sea surface temperature (°C, dashed lines). Also shown are the 72-h back trajectories for air that was W of Tasmania at noon on February 1, 1995. The curves are labelled 800, 900 and 1000 to denote the pressure levels of the trajectories at noon during the flight. Position marks are given for each 12 h along the trajectories. The labels along the curves (24, 48 and 72 h) denote hours prior to noon February 1, 1995.

Fig. 2. Pressure levels of air along the three trajectories in the 72 h prior to noon on February 1, 1995.
labels far W of Tasmania. After that, the air subsided under the influence of the large high-pressure system, and the wind speed was reduced. Twelve hours prior to the flight, the mean boundary-layer air speed was 10 m s\(^{-1}\), but during the flight, the wind in the boundary layer was light (4–6.5 m s\(^{-1}\)) from W–NW. During the 72-h period, the air close to the surface experienced an increase in sea surface temperature of as much as 11°C. In the last 12 h, the change in sea surface temperature was quite modest, probably only a fraction of a degree.

Fig. 3 is a NOAA-9 visual satellite image from 1007 h showing Tasmania and the Southern Ocean. The image shows numerous mesoscale cloud structures — closed cell convection — with typical horizontal scales of about 30 km. The cells are separated by subsidence lines, which contain thin cloud or clear air on much smaller scales. All observations in this paper will be presented in a coordinate system advecting towards the SE at 4 m s\(^{-1}\). A distance scale has been added to Fig. 3. The F27 observations were taken along this scale and it will be used as a reference throughout the paper. The 4 m s\(^{-1}\) advection speed was determined by iteration to ensure that cumulus clouds are nearly vertical in the advecting coordinate system. The 4 m s\(^{-1}\) speed was also the mean along-track wind speed near the surface. Using this advective coordinate system, observations taken at different times can be related to one another; this is important for studying vertical coherence and also for relating aircraft observations to positions on the satellite image.

A number of linear cloud features were observed both from above and below the cloud deck, some of these features being visible on the satellite image. One example is the band apparent in the closed circulation beyond the NW end of the distance scale (see the small arrows in Fig. 3). A number of cumulus bands were penetrated to the SE of this one during the flight; the first one was located in the closed cell circulation centred on the 10-km mark on the distance scale in Fig. 3.

The F27 entered the area at 1027 h that was 20 min after the satellite pass. Over the next 2.5 h, the stratiform decks and cumulus clouds persisted and developed. A second flight later in the afternoon showed that the pattern of stratus decks and cumulus continued throughout the day.

2.2. Terminology

Albrecht et al. (1979) described a model of cumulus clouds in the marine boundary layer, and we will adopt a similar terminology (see Fig. 4). They examined a case in which there were positive heat and moisture fluxes from the sea surface. Close to the sea surface is a mixed-layer in which both water vapour mixing ratio \(q_v\) and dry static energy \(s\) is assumed well mixed. The mixed-layer is assumed saturated at the top and capped by a small inversion, referred to as the transition layer. The top of the mixed-layer is also the base of cumulus clouds. These clouds define a cloud layer to the marine boundary-layer inversion (see Fig. 4). It is possible to have profiles of vapour mixing ratio and dry static energy such that the top of the cloud layer is saturated everywhere, i.e. an upper level stratus deck may exist. The subsaturated layer below the stratiform layer will be called the intermediate layer. The mixed-layer and cloud layer together constitute the marine boundary layer (or, boundary layer for short). These
Fig. 3. NOAA-9 visible satellite image of stratocumulus cloud features for 1007 h. The flight area was approximately 42°S; the location of a cumulus band is highlighted with the two small arrows. A distance scale has also been added with labels in kilometers. The flight tracks were mainly along the 10–60 km part of the distance scale.
profiles are convenient idealisations of actual boundary layers. In the present case, a more complex boundary layer structure will be documented.

2.3. Flight plan

Fig. 5a and b shows the flight tracks for two stacks from the first flight on February 1 projected onto a vertical plane. A flight stack consists of horizontal legs of about 30–40 km length and a sounding from above cloud top to near sea surface. The stack commenced with a horizontal leg about 700 or 1000 m above cloud top for radiation measurements (the radiation leg), followed by a turn and descent to 60 m above sea level (the descent sounding). Then, a horizontal leg was made at 60-m height (the surface leg), after which an ascent to 800 m was made (the ascent sounding). This was followed by four horizontal legs, which were partly or fully in cloud (the four cloud legs), the lowest three of which were through cumulus clouds at various times. The highest cloud leg was generally in solid cloud for 30–40 km, mostly in a stratus deck, but at times in the upper portions of the cumulus clouds penetrated into the stratus cloud.

The track line in Fig. 5a and b has been made bold where cloud liquid water is present. The first stack (Fig. 5a) shows extensive cloud liquid water at the ends of the track, particularly in the NW end, which are associated with relatively deep cumulus clouds. The descent sounding in the middle goes through a shallow stratiform cloud deck, which is also penetrated by the aircraft in the upper of the four cloud legs. The second stack (Fig. 5b) shows the descent sounding going through a shallow stratiform cloud deck, but the four subsequent cloud legs show that a cumulus cloud in the centre later rose up into the stratiform cloud.
Fig. 5. Flight tracks projected onto a vertical plane (NW–SE) for the two stacks, (a) and (b), respectively. Crosses are shown for each 10 min of flying time. The tracks are shown in bold where cloud liquid water was observed.
Wind hodographs were created from average wind direction and speed for each of the horizontal legs, with the near surface wind slightly less than 4 m s\(^{-1}\) and from W–WNW. The cloud legs have winds nearer NW and speeds of 4.5–6.5 m s\(^{-1}\), while the radiation legs, 700–1000 m above the inversion, showed wind speeds increasing to 9 m s\(^{-1}\) with a northwesterly direction.

3. Observations from stack 1

3.1. Cloud top

The top of the marine boundary-layer cloud was monitored by forward and downward looking video cameras. The cloud top was visually quite uniform along the first radiation leg, but a band of cumulus was visible a few kilometers NW of the end of the flight track (near 0 km on the distance scale; see Fig. 3). This cumulus band was roughly perpendicular to the flight track and was estimated to be tens of kilometres long. The band was visible as several cumulus domes, each of which was estimated to overshoot the stratiform cloud top by no more than 100 m.

3.2. Vertical profiles

After completing the radiation leg above cloud, the aircraft made a descent almost to the sea surface towards SE (see Fig. 5a). After a surface leg, the aircraft ascended to 700 m at the NW end of the stack. Profiles from these two soundings are shown in Fig. 6a–f. During the descent, the top of the solid stratiform cloud deck was at 862 hPa (1350 m), and the less well-defined base at 892 hPa (1060 m). Some smaller fractus clouds were observed below the main stratiform base.

Fig. 6a shows a profile of the virtual potential temperature (\(\theta_v\)), here defined as:

\[
\theta_v = \theta (1 + 0.61 q_v - q_l)
\]

where \(q_v\) and \(q_l\) are the vapour mixing ratio and the cloud liquid water mixing ratio, respectively. Fig. 6b shows the vertical profile of \(q_v\), and Fig. 6c shows the vertical profiles of \(q_l\) from the King probe and of drizzle mixing ratio (\(q_{dl}\)) from the 260X probe. The profiles of \(\theta_v\) and \(q_v\) have been made bold whenever the aircraft measured solid cloud in a 1-s sample. This was done by requiring that all 64 samples within a second contained cloud drops. The mixed-layer (below 918 hPa, as measured during the descent sounding) is not entirely neutral in stability; the lowest part of the descent sounding below 300 m altitude reveals cooler air. After the completion of the descent sounding, the aircraft penetrated a shower and the cool air observed in the descent sounding may be caused by evaporation of precipitation from this shower.

The profile of \(\theta_v\) for the ascent sounding to 700 m, at the NW end of the flight track, appears to show a slightly warmer and neutrally stratified mixed-layer. However, the corresponding profiles of \(q_v\) from the two soundings are not at all well mixed. There is a decrease of almost 1 g kg\(^{-1}\) from the surface to the top of the mixed-layer, and considerable variability is also apparent.
Fig. 6a shows virtual moist adiabats from a lifting condensation level (LCL) determined from surface leg measurements (to be described later) and from the base of the stratiform deck; these are labelled ‘‘Cu’’ and ‘‘Sr’’ in Fig. 6a, respectively. The pressure and temperature of the LCL are 945 hPa and 9.7°C, and for the stratiform base, they are 892 hPa and 6.5°C. Notice that the LCL level is as much as 25 hPa below the apparent top of the mixed-layer.

The observed profile of $\theta_v$ in the stratiform cloud deck appears to be stable with regard to moist adiabatic processes. Other results, presented in Appendix A, suggest that the cloud deck is nearly neutrally stratified; the apparent cloud layer stability in Fig. 6a is likely the result of a slow response of the Rosemount temperature sensor housing, such as described by Rodi and Spyers-Duran (1971). The implication is that the temperature, and thus $\theta_v$, in the upper part of the stratiform deck is recorded as being
too high during the descent sounding. We suspect that the real gradient of $\theta_v$ in the stratiform cloud is, in fact, much closer to the gradient predicted by adiabatic ascent (see Appendix A).

The descent sounding thus shows four de-coupled layers below the marine boundary layer inversion: (i) the $\theta_v$ gradient is likely close to moist adiabatic in the stratiform cloud, (ii) it is very stable with regard to dry adiabatic motions in the intermediate layer below the stratiform cloud, (iii) it is nearly neutral in most of the mixed-layer and (iv) quite stable in the lowest 300 m above sea surface. The stable layer lowest to sea surface (the cold pool) is of limited horizontal extent.

Fig. 6c shows the observed values of cloud liquid water mixing ratio, and the corresponding values based on adiabatic ascent, from both the LCL-determined cloud base of the cumulus clouds and from the observed base of the stratiform clouds. The stratiform cloud deck has strong vertical variations in $q_l$, and both entrainment and drizzle production may contribute to this variability. The stratiform cloud base is quite ill-defined due to the fractus clouds below, and it is therefore difficult to say to what extent the stratiform cloud has less water than the value based on adiabatic ascent. Drizzle, as measured by the 260X probe, constitutes a significant fraction of the total condensed water in the stratiform cloud, and drizzle from the stratiform cloud deck extends to below the top of the mixed-layer.

Relative humidity (Fig. 6d) decreases from the stratiform cloud base towards the sea surface, although the value is always greater than 75% below the stratiform cloud.

Wind direction and speed are shown in Fig. 6e–f, with wind direction being generally westerly at low levels and changing to more northwesterly with height. Wind speed was in general greatest in the air above the inversion. In the stratiform cloud deck, the wind speed decreased substantially, before increasing closer to cloud base. Throughout the marine boundary layer, the wind speeds were quite small with values between 3 and 7 m s$^{-1}$. The high variability, in both wind direction and speed, is probably caused by mesoscale circulations setup by the closed cell structure of the convection.

A measure of the turbulence intensity has been calculated by filtering the 64 samples per second vertical velocity ($w$) series from the descent sounding such that only frequencies between 1 and 10 Hz remain. The resulting series was then used to calculate a 1-Hz root-mean-square velocity, $w_{rms}$. This series was subsequently smoothed with an 11-s running mean, and the result is shown in Fig. 6f. It can be seen that the turbulence intensity is highest in the middle of the stratiform cloud, that it is generally low in the mixed-layer, before again increasing as the aircraft enters the cold pool 300 m above the surface.

### 3.3. Mixed-layer properties

A surface leg towards NW was made after the completion of the descent sounding. A photo, taken at the 23-km mark on the surface leg looking NW, shows a very pronounced band of cumulus convection (see large arrows in Fig. 7). This cumulus feature will be referred to as Band 1. It is located at the 12–16-km mark on the distance scale (Figs. 3 and 5a), i.e. in the turn, after the completion of the surface leg. Another band of cumulus clouds is visible much further to the NW (see small arrows in Fig. 7).
This is towards the region where the satellite image shows a cloud band NW of the flight area (see small arrows in Fig. 3).

Measurements along the surface leg (60 m altitude) show considerable variability in thermodynamic parameters (see Fig. 8a–e), for example, vapour mixing ratio \( q \); Fig. 8a has variations of up to 1 g kg\(^{-1} \) along the track. This is not surprising given the large vertical gradients in \( q \), observed during the descent sounding (Fig. 6b). The virtual potential temperature (Fig. 8b) shows a marked minimum in the SE end of the track. This is where Fig. 5a shows a cloud with high liquid water content immediately to the SE of the track; the cold pool is likely the result of evaporating precipitation from this cloud. The virtual temperature trace increases slowly towards the NW to reach a value 1\(^\circ\)C higher than in the centre of the cold pool. The descent sounding showed that this cold pool was less than 300 m deep. The pressure of the LCL \( (p_{\text{LCL}}) \) shows variations within a 25-hPa range (see Fig. 8c). The sea surface temperature, as measured from a downward looking infrared thermometer (Barnes PRT-5, not shown), has values of 16.2 \pm 0.2\(^\circ\)C over the 40-km long surface leg. The sea surface temperature is in general 1.3\(^\circ\)C higher than the air temperature immediately above the sea surface, which implies a positive heat flux from the sea surface to the air. The cold pool is even colder than the rest of the mixed-layer and hence much colder than the sea surface. Therefore, the cold pool must have been created by evaporation of falling precipitation.

The along-track horizontal wind component \( u \) in the advecting coordinate system is shown in Fig. 8d. The flight track was nearly perpendicular to the cumulus band in the NW end of the flight track, and the along-track wind shows significant variations. In the NW end of the track, \( u \) is towards SE. In the SE end of the track, it is generally towards
NW, and the centre of the track shows no net along-track wind component relative to the advecting coordinate system. The along-track wind is used to calculate the along-track convergence \( C = -d u_t/dx \); Fig. 8e, where \( x \) is the distance along the track. This series has been filtered with a 31-s (2.5 km) running mean, and the first and last segments of the series have been omitted. The along-track air flow is divergent at the far NW end of the track (NW of 19 km), and in the subsequent turn (not shown), there are traces of drizzle. Between 19 and 26 km, there is a strong, and near solid, region of along-track convergence, and the region between 26 and 40 km is devoid of systematic along-track convergence or divergence. Further to the SE, there is primarily light along-track convergence. The region with the lowest virtual potential temperature (54 km and beyond; see Fig. 8b) is divergent in the along-track direction. Later, a cumulus band was observed to grow over the 19–26 km along-track convergence zone, and the thermodynamic parameters for this segment have been used to calculate the LCL (+) shown in Fig. 6a and b.

3.4. The cloud legs

The four cloud legs were all made between the LCL (945 hPa) and the inversion. The lowest cloud leg was at 927 hPa; i.e. below the top of the mixed-layer as determined.
from the descent sounding (918 hPa; Fig. 6a). Analysis, to be presented later, shows that the mixed-layer height varies along the track. The lowest cloud leg appears to be partially in the mixed-layer and partially above the top of the mixed-layer.

Observations of cloud and drizzle water mixing ratio, for the four cloud legs and the surface leg, are shown in Fig. 9a–e. The surface leg is virtually free from drizzle, but small traces are apparent in the far SE end of the track (56–57 km). Cloud liquid water in the lowest cloud leg (Fig. 9d, thin line) is only apparent NW of the 15-km mark; this is part of the cumulus band visible in Fig. 7 (Band 1).

In the NW end of the three upper cloud legs, there is a region with high cloud liquid water content extending progressively towards the SE at high altitudes. In the top cloud leg, it may be as far towards the SE as the 27-km mark. This indicates that a strong

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Fig. 9. Cloud and drizzle liquid water mixing ratio ($q_c$ and $q_d$, respectively) for the four cloud legs and the surface leg in stack 1. The top box shows the values for the highest flight leg, the bottom box for the surface leg. Cloud liquid water is measured using the King probe (thin line) and drizzle water is from the 260X probe (bold line), class 5 and larger drops; i.e. $r > 22.5$ nominally. (A brief data system error is apparent in the centre of (c)).
cumulus cloud grew over the main surface along-track convergence (19–26 km) while the F27 executed the three upper cloud legs. This cloud is directly on the SE side of cumulus Band 1. Due to low visibility, evident in video recordings, it could not be determined if this cumulus feature was band-shaped; it will therefore be called cumulus Cloud 2. This cumulus is the first of the cloud features that is not readily identifiable in the satellite image (Fig. 3) obtained an hour earlier.

A solid stratiform deck is also apparent SE of the 30-km mark in the cloud top leg and to a lesser extent in the cloud leg below. Both of these cloud legs also show what appears to be a small cumulus cloud around the 45-km mark.

The drizzle mixing ratio is very high, particularly in and below the stratiform cloud region. In the highest cloud leg, close to half of the condensed water in the stratiform cloud is contained in drizzle-sized drops. The cloud liquid water is much higher than the drizzle water content in the solid cumulus clouds, the young age of the cumulus clouds being a possible explanation for this difference. Drizzle from the stratiform cloud is apparent even in the lowest cloud leg (28–50 km; Fig. 9d).

The buoyancy field is of considerable interest for this cumulus and stratocumulus system, though, due to horizontal variability, it is difficult to define a representative sounding for such a complicated system. Table 1 shows a simplified sounding that is based on selected measurements from stack 2. This sounding is solely used as a simple reference, so that buoyancy can be calculated relative to it. Fig. 10 shows the virtual temperature excess $\Delta \theta_v = \theta_v - \theta_v(p)$, where $\theta_v$ is the virtual potential temperature from the aircraft during the flight legs, and $\theta_v(p)$ is the value for the reference sounding at pressure $p$ (see Table 1).

In the lowest cloud leg (Fig. 10d), a cold air region is apparent from 30 to 52 km. This region is almost identical to the region with light drizzle shown in Fig. 9d. The $\Delta \theta_v$ field shows a pronounced maximum between 20 and 30 km. This coincides with little or no trace of drizzle in Fig. 9d. This cloud-free region shows high $\theta_v$ suggesting that the top of the mixed-layer dips below the flight level between 20 and 30 km; this results in high $\theta_v$ air being observed at the flight level. The region from 30 to 52 km is essentially at the temperature of the mixed-layer, i.e. the top of the mixed-layer is above the flight level in this region.

Table 1
Simplified reference sounding
Values from below 918 hPa are loosely based on the stack 2 ascent sounding; values from above 918 hPa are from the stack 2 descent sounding. Note that no transition layer jump is included in this table.

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<th>$p$ (hPa)</th>
<th>$z$ (m)</th>
<th>$T$ (°C)</th>
<th>$T_v$ (°C)</th>
<th>$q_v$ (g kg$^{-1}$)</th>
<th>$\theta_v$ (K)</th>
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The virtual temperature patterns in the top two cloud legs are quite similar. There are maxima at the two ends of the flight legs and a broad minimum at the centre. The region of maximum cloud liquid water at the NW end (Fig. 9a and b) coincides with one of the most buoyant regions in Fig. 10a and b.

The vertical wind speed ($w$, bold line) and along-track wind speed ($u_r$, thin line) are shown for the same five legs in Fig. 11a–e. The surface leg has quite modest and fairly small-scale vertical velocities. The strong along-track convergence region in the surface leg between 19 and 26 km is not readily apparent in the vertical velocity at the same level. This may be because the surface leg is made as low as 60 m above the sea surface; a convergence of $5 \times 10^{-4}$ m s$^{-1}$ (see Fig. 8e) between 0 and 60 m results in an updraft of just 0.03 m s$^{-1}$ at 60 m altitude. If the same convergence is present over the depth between surface and the LCL (0–600 m), then the resulting mean updraft at the LCL in the 19–26 km region would be 0.3 m s$^{-1}$. The along-track convergence pattern is thus consistent with the subsequent formation of a cumulus cloud aloft. The highest updrafts
and downdrafts in the cloud legs are found in the NW end of the track and appear to be related to cumulus Band 1 and cumulus Cloud 2.

The along-track wind component \( u \) in the advecting coordinate system is shown in Fig. 11 with the thin lines. The lowest two cloud legs show very low values of \( u \) NW of the 25–27-km mark. It is not apparent why the strong cloud layer gradients in \( u \) are located in this position; the location does not coincide with the cumulus edge or the stratiform drizzle region.

4. Observations from stack 2

At the completion of stack 1, there was a mature cumulus band in the far NW end of the track (Band 1), a new strong cumulus cloud on the SE side of it (Cloud 2), a more stratiform cloud in the centre of the flight track, and another mature cumulus region in
the far SE end of the flight track. Cloud 2 matured during the time it took to fly stack 2, and the cumulus cloud in the far SE is not apparent at later times. During the time stack 2 was flown, a new cumulus band (Band 3) developed in the centre of the stack (Fig. 5b) where previously only a stratiform cloud existed. The manner in which this band appeared to be forced by the maturation of Cloud 2 will now be described.

4.1. Vertical profiles

Fig. 12a–f shows vertical profiles of thermodynamic, microphysical, and wind parameters for the descent sounding, which began in the SE end of the stack and finished near the surface in the far NW end (see Fig. 5b). Some of the profiles also show values for the ascent sounding to 800 m, which was made in the SE end of the stack.

The virtual potential temperature profile (Fig. 12a) shows an inversion of about 2.5°C at 848 hPa, and a stratiform cloud layer between 848 and 880 hPa. Below this is a layer with, on average, stable stratification, but with strong variability (880–925 hPa). Video recordings show that the aircraft, at 920 hPa, skimmed the top of a small cumulus band, which later grew to dominate the cloud field. The small decrease in $\theta_e$ at 920 hPa may well be due to mixing between this cumulus band and clear air. At 928–926 hPa, there

![Fig. 12. Same as Fig. 6, but for stack 2.](image-url)
is a small transition layer, marking the top of the mixed-layer found in the descent sounding.

The descent sounding shows this ‘‘mixed-layer’’ as having an almost neutral stratification in the upper portion. Lower down, the virtual potential temperature decreases to about 1±1.5°C less than at higher levels. This cold pool in the NW end of the descent sounding may have resulted from the evaporation of 0.4–0.6 g kg$^{-1}$ of precipitation in the mixed-layer. Sea surface pressure is calculated to be 1012 hPa.

After reaching 60 m altitude, the aircraft turned around, made a 40-km surface leg, and ascended to 800 m in the SE end of the flight stack. Fig. 12a also shows the virtual temperature from this ascent. It can be seen that a neutrally stratified mixed-layer extends all the way to 918 hPa (800 m) at this end of the stack, i.e. at this stage, the precipitation-cooled air is confined to the NW end of the stack. The $\theta_v$ profiles for the descent and ascent soundings match one another closely in the narrow range of 930–940 hPa, but the ascent sounding is neutrally stratified to a higher level (918 hPa).

Fig. 12a also shows (with a bold +) the pressure and virtual temperature for a LCL determined from surface leg measurements; the pressure and temperature at the LCL are 939 hPa and 9.2°C, respectively. The $\theta_v$ profile for a parcel ascending adiabatically from this LCL is shown with a dashed curve, labelled Cu in Fig. 12a. Air ascending adiabatically from the LCL has an alternating small positive or negative buoyancy in the first 40 hPa above the LCL. Higher up, the buoyancy is systematically positive for an adiabatically ascending parcel, but the excess virtual potential temperature is always less than 0.5°C. The instability in the cloud layer is quite modest, with a convective available potential energy (CAPE) of about 3 m$^2$s$^{-2}$.

The figure also shows the virtual potential temperature of a parcel ascending moist adiabatically from the base of the stratiform cloud deck, labelled Sr. The virtual potential temperature gradient in the stratiform cloud is shown to be more stable than a moist adiabat. As for stack 1, we believe that this apparent stability in the stratiform deck is largely an instrumental artifact caused by the slow time response of the Rosemount temperature sensor housing (see Appendix A).

Fig. 12b shows the vapour mixing ratio, $q_v$, for both the descent sounding and for the 800-m ascent made after the completion of the surface leg. The ‘‘mixed-layer’’ is well mixed in $\theta_v$ (Fig. 12a) away from the cold pool, but it is not well mixed in $q_v$ (Fig. 12b). There is both a strong mean vertical gradient in $q_v$ and local variability, the latter presumably due to the localised nature of evaporating drizzle and to individual thermals rising from the sea surface. The vapour mixing ratio in the descent sounding is, in general, lower than that in the ascent sounding. This is consistent with evaporation of precipitation in the cold pool part of the descent sounding. The cooling of air due to evaporation may lead to downdrafts and thus bring down air that is comparatively low in vapour mixing ratio. The vapour mixing ratio of the LCL (determined from the surface leg) is about 1 g kg$^{-1}$ higher than the mixed-layer air at the same level (see the bold + in Fig. 12b).

The vertical profile of cloud liquid water mixing ratio determined from the King probe is shown in Fig. 12c. Also shown in this figure is the drizzle water mixing ratio from the PMS 260X probe for drops larger than 23 μm radius. Drizzle observed during the descent sounding reaches a peak at the stratiform cloud base, but is apparent at all
levels from the cloud top to about 30 hPa below the stratiform cloud base. Some drizzle is also observed in the lower part of the descent sounding, both in the top part of the mixed-layer and in the cold pool. The maximum drizzle water mixing ratio is about half the maximum cloud water mixing ratio, though at different altitudes. The figure also shows the calculated liquid water mixing values assuming moist adiabatic ascent from both the LCL (curve labelled Cu) and for the base of the stratiform cloud (curve labelled Sr). The reduction in cloud liquid water from the adiabatic values in the stratiform cloud could be due to drizzle and entrainment, but also to uncertainty related to the determination of the stratiform cloud base.

The relative humidity in clear air has important effects on the evaporation of drizzle drops below the cloud and on entrainment. Fig. 12d shows the vertical profile of relative humidity for the descent sounding. The air between the LCL and the stratiform cloud is very moist with relative humidity larger than 85%; this is partly caused by evaporation of drizzle. The air in the first 50 m in and above the marine boundary-layer inversion is also comparatively moist, with a relative humidity of 75%. A further 50 m higher, the relative humidity decreases to about 50%. Wind direction and speed for the descent sounding are shown in Fig. 12e–f; as was the case for stack 1, the winds are light and

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![Figure 13](image_url)

Fig. 13. Surface leg along-track variation of (a) vapour mixing ratio, $q_v$; (b) virtual potential temperature, $\theta_v$; (c) LCL pressure, $p_{LCL}$; (d) along-track wind speed, $u$; and (e) along-track convergence, $C$. The measurements are for stack 2, and three thermodynamically distinct regions have been identified (see text).
variable. The turbulence level ($w_{max}$; Fig. 12f) is again high in the centre of the stratiform cloud, somewhat lower below the stratiform cloud, low in the upper part of the mixed-layer, and high further down as the aircraft enters the cold pool in the lower mixed-layer.

4.2. Mixed-layer

Measurements of thermodynamic parameters from the surface leg (60-m altitude) show considerable variability over the 32-km flight distance from NW towards SE. Three segments can be identified from the thermodynamic and dynamic parameters shown in Fig. 13a–e.

The surface leg commenced in a region of light, intermittent drizzle (segment A, 18–28 km in Fig. 13a).

This segment shows the highest values of vapour mixing ratio ($q_v$), the highest LCL pressure (i.e. LCL is closest to the sea surface), and the lowest virtual potential temperature ($\theta_v$). The low $\theta_v$ implies that this air is unlikely to ascend, and so defines the cold pool.

The next segment (B, 28–34 km in Fig. 13a) shows the lowest values of $q_v$ but a considerably higher $\theta_v$. The air can readily ascend if forced, but the dryness of the air ensures that the LCL pressure is very low; i.e. the air in segment B must be forced high up before condensation begins.

Further towards the SE is segment C (SE of the 34-km mark in Fig. 13), which is characterised by warmer and slightly dryer air than segment A, but which is more moist than segment B. Segment C has a LCL that is intermediate between those of segments A

![Photo of cloud base from the second surface leg (1207 h) at the 26-km mark. Note the band of shallow cumuli in the foreground and the mature band in the background (see text for further details).](image)

Fig. 14.
and B. Segment C is by far the most likely source for air entering the cumulus cloud bases, as the air is both warm and moist.

The LCL pressure, $p_{LCL}$, varies by 30 hPa along the flight track (see Fig. 13c). The cloud base, calculated from the average LCL parameters for segment C, is 942 hPa and 9.3. The altitude of the LCL is 615 m. The virtual moist adiabat ("Cu" in Fig. 12a) is based on the LCL determined from segment C.

The along-track horizontal wind component in the advecting coordinate system ($u_t$) is shown in Fig. 13d. The flight track was nearly at right angles to the cumulus band. In region A, $u_t$ increases from $-1$ to $3$ m s$^{-1}$ before settling back to $1$ m s$^{-1}$ in the SE end of region A. This pattern is consistent with a downdraft spreading out in region A. In region B, the value of $u_t$ decreases gradually towards the SE, and a few kilometres into region C, it takes on a fairly uniform value of $-1$ to $-2$ m s$^{-1}$.

Fig. 13e shows the along-track convergence, $C$. NW of the 23-km mark, there is strong along-track divergence. This divergence is located immediately below the centre of the cumulus cloud.

Fig. 15. Cloud and drizzle liquid water mixing ratio ($q_l$ and $q_v$, respectively) for the four cloud legs and surface leg in stack 2. Legend as in Fig. 9.
of Cloud 2 as determined from Fig. 9. From 23 to 35 km, there is a fairly persistent along-track convergence pattern, and further towards the SE, there is no net along-track convergence or divergence. This pattern is consistent with a downdraft in region A spreading out and forcing the lowest part of the mixed-layer for a distance of 15 km further to the SE. Note how the along-track convergence extends into the far NW part of region C, which the thermodynamic analysis suggests as the source of air most likely to ascend into cumulus clouds.

The sea-surface temperature as determined from the Barnes PRT-5 (not shown) has values of 16.0 ± 0.2°C along the surface leg. The air immediately above the surface is, on average, 1.1°C colder than the sea surface.

A photo looking towards SE was taken at the 26-km mark (1207 h) during the surface leg (see Fig. 14), and showed two bands of cumulus clouds visible below the stratiform cloud deck. The nearest band is visible as a string of very shallow cumulus, estimated at no more than 100–200 m deep; video recordings show that these are located at the

![Buoyancy diagram](image_url)

Fig. 16. Buoyancy, expressed as deviation of $\theta_v$ from the reference $\theta_v$ sounding (Table 1), for surface leg and four cloud legs from stack 2.
33-km mark. This shallow cumulus band was also observed during the descent sounding at the same location 7 min earlier. We will now show how this cumulus line developed and ended up dominating the cloud field.

4.3. The cloud legs

Cloud and drizzle liquid water mixing ratio, from the surface leg and four cloud legs, are shown in Fig. 15. A very pronounced cumulus cloud is evident from the 34–40-km region at low levels and in a somewhat wider region at the highest level. Video recordings from the lowest cloud leg show that it is a cumulus band located where the very shallow band of cumuli were observed in Fig. 14. This band has been described by Boers et al. (1997) and will be referred to here as Band 3.

The drizzle water contents in the four cloud legs are very large, in some regions even exceeding the cloud liquid water content. It is noted that the drizzle water content is very

![Graph showing vertical velocity (w, bold line) and along-track wind speed (u, thin line) for the surface leg and four cloud legs in stack 2.](image-url)
low in the new cumulus band (Band 3) and high in the mature cumulus cloud in the NW end (Cloud 2).

The virtual temperature difference between observations and the reference sounding (Table 1), expressed as $\Delta \theta_v$, along the surface leg and the four cloud legs is shown in Fig. 16. For the four cloud legs, the figure shows that $\Delta \theta_v$ is high at the two ends of the track and relatively low in the centre. The new cumulus band is positively buoyant relative to its immediate environment in the two lower cloud legs (Fig. 16c and d). The lowest cloud leg is close to the top of the mixed-layer. As for stack 1, there is a possibility that a small segment of this leg is above the top of the mixed-layer, e.g. the relatively warm segment from 25 to 29 km.

The vertical velocity for stack 2 is shown in Fig. 17. The new cumulus band (Band 3) dominates the velocity variations in the lower three cloud legs (Fig. 17b–d). In the top cloud leg, the cumulus shows very modest updrafts and downdrafts, which are indistinguishable from the remainder of the top cloud leg.

The along-track wind ($u_t$, thin line in Fig. 17) shows Band 3 to be in a region of broad along-track divergence (high $u_t$ on the SE side, low $u_t$ on the NW side); this is particularly clear in the lower two cloud legs. The two upper flight legs show along-track divergence within Band 3.

5. Discussion

In this section, we discuss the interaction between the three “generations” of cumulus clouds. The existence of the cold pools in the mixed-layer is shown to be essential to the cloud organisation. Therefore, we will use a simple model to compare the mixed-layer development inside and outside a cold pool.

5.1. Convective organisation

The cloud fields developed considerably over the 2.5 h of flying time. At least two, possibly three, cumulus bands were observed in various stages of inception, new growth, maturation, and dissipation.

The existence of cumulus bands is based on visual observations. The band shape could only be determined from the cloud appearance during the surface and the lowest cloud legs. At higher altitudes, the same cloud formations were observed, but they were obscured by falling drizzle or they were entirely embedded in the stratiform cloud. Photographs from the flight legs above the boundary layer show some linear features but the orientation and length cannot be determined.

Cumulus Band 1 was clearly visible from both the video and a photograph on the first surface flight leg. The length of the band can be determined using the known distance from the aircraft to the band and known viewing angles of the video and photographic cameras. Observations show that Band 1 had a length of about 22 km, or more, as the cumulus band extended to the edge of the camera field-of-view. Cumulus Band 3, determined from the end of the stack 2 ascent sounding, had a length of 10 km, possibly more, as it also extended beyond the camera viewing angle. As with other cumulus
bands, there may be cellular structure along the bands. This is particularly clear in Fig. 14.

Cumulus Band 1 had a width of 2–3 km as measured close to the NE end of the band. Band 3 was penetrated closer to its centre and had a width of 5 km at low levels. It is uncertain if Cloud 2 was band-shaped. This cloud grew rapidly and was only observed during the three upper cloud legs where visibility was limited.

The bands appeared to be oriented in a SSW–NNE direction, which is roughly perpendicular to the near-surface wind (W–NW). The bands were not penetrated strictly at right angles and their apparent width (2–3 and 5 km, respectively) could be overestimates. The closed cell circulation evident in the satellite image (Fig. 3) suggests that the bands were of limited length. The ratio of length-to-width of a new band appears to be about 3:1. It would be unlikely that the bands were as long as the scale of the closed cells apparent in the satellite image, since this would imply abrupt transitions at the end of the cumulus bands. It is unclear if cumulus bands are a common feature of closed cell circulation, and, if they are, whether the orientation is commonly perpendicular to the low-level wind. As a cold air downdraft spreads out, it may have a tendency to form an arc of increasing radius. The new convection may form where a spreading cold pool “collides” with a moist and warm part of the mixed-layer. This could in principle lead to arc-shaped or near-linear cumulus bands. If this mechanism is taking place, then the orientation of the bands would not depend on the mean wind direction or mean vertical shear. The satellite image shows reflectivity maxima that in some cases are linear, but the lines appear to have a variety of orientations. The aircraft observations show cumulus clouds with high droplet concentrations, high albedo, and low effective radius $r_e = \left( \int_0^\infty n(r)r^3dr \right) / \int_0^\infty n(r)r^2dr$, Hansen and Travis, 1974. This may be a useful signature for satellite determinations of the linearity of cumulus bands embedded in stratocumulus.

Wang and Lenschow (1995) flew through a series of cumulus clouds that grew into a stratocumulus deck. Their stacked flight legs were roughly perpendicular to the mean surface wind; i.e. turned 90° relative to our measurements. They found that the cumulus clouds at higher levels were visible for 36 km along the track. They used an upwards pointing lidar to determine cloud base, and these measurements show several cumulus cells along the track. However, no photographs of the clouds were shown and band structure was not discussed.

The closed cell structure apparent in the satellite image pertains to the cloud layer. The cell structure may be different in the mixed-layer due to the presence of up to four decoupled layers within the marine boundary layer (see Sections 3.2 and 4.2). The closed cell circulation appears to be sustained by the drizzle formation, the evaporation in the mixed-layer, and the ensuing forcing from spreading of evaporatively cooled downdrafts. It is therefore likely to be a cloud type more typical of clean air where precipitation by coalescence processes are more likely. The air trajectory analysis (Fig. 1) shows no continental influence in the 72 h prior to the flight. No sample with adiabatic liquid water content was observed from the aircraft in the cumulus, but an entrainment analysis (Part II) shows that cloud droplet concentration immediately above cloud base in the new and solid cumulus bands must be in the range $155–235 \times 10^6$ kg$^{-1}$. This droplet concentration is consistent with observations of CCN concentrations from Cape Grim during summer months (Gras, 1995).
All of the aircraft observations are projected onto a coordinate system advecting at 4 m s$^{-1}$ towards SE. This was found by trial and error to give the best vertical coherence of cumulus clouds. The 4 m s$^{-1}$ is also the average along-track wind speed in the surface flight legs, and closer to the inversion, the along-track wind speed increased to 4.5–6.5 m s$^{-1}$. The advecting coordinate system is not perfect; however, it is a reasonable and necessary way of relating observations taken at different locations and times to one another.

From the aircraft wind measurements, it is not possible to calculate the complete horizontal convergence and divergence. Instead we use the aircraft wind measurements to calculate the along-track convergence and divergence components. For a boundary layer dominated by near-linear cumulus convection, this simplification yields a good agreement between the variations in along-track convergence/divergence components and the cloud development aloft.

5.2. Synthesis of cloud evolution

The schematic evolution of the boundary layer is detailed in Fig. 18a–d. Each of these shows the flight tracks for a specific time period, roughly corresponding to half the time it took to execute a flight stack. As before, the flight track line is made bolder where cloud liquid water is observed. The figure also shows the outline of the cold pools observed during each of the two stacks and of the along-track convergence/divergence field during the surface legs. The degree of cooling in the cold pools is shown with a cold front signature where the cold pool became apparent, and also with contours of reduction in $\theta_e$ compared to other air at the same level. Arrows show where the highest updrafts and downdrafts were observed; the size of the arrows is roughly proportional to the updraft speed. The top of the mixed-layer (transition layer) is shown with a bold dashed line. Above cloud top, the variation in short-wave albedo is shown. Some of the figures also show the variation of the effective radius as measured during the highest cloud leg. Drizzle mixing ratio is shown with shading. Some extrapolation from the flight tracks has been made, given that drizzle normally originates from a higher level and falls to a lower level, relative to the observation point.

Fig. 18a shows cumulus Band 1 in the NW end while apparently still in an early stage of development. The aircraft flew in the NE end of this band and the observations showed very little drizzle below the band. There is nevertheless a distinct along-track divergence and convergence pattern SE of Band 1 in the surface flight leg. It is possible that this is caused by precipitation falling from parts of Band 1 further to the SW. The cold pool and low-level circulations in the SE end of the surface flight leg are caused by precipitation from the mature cumulus above.

A strong updraft is observed in Band 1 (Fig. 18a). The mixed-layer top is lowest in the drizzle free region (15–28 km; Fig. 18a), whereas the mixed-layer top is much higher in the region of stratiform drizzle (28–50 km). A small cumulus is apparent at the 30-km mark during cloud leg 1. This is nearly above the region of maximum temperature and humidity observed during the surface leg. After the completion of cloud leg 1, a field of scud clouds was observed below the aircraft during cloud leg 2. This may be related to the spreading of the cold pool to the SE.
The cloud top was below a height of 1400 m during the descent sounding (Fig. 18a). It may have been higher above the cumuli, but this cannot be substantiated by measurements. The cloud albedo is low near the 20-km mark along the cloud top where a downward looking Barnes PRT-5 remote temperature sensor measured a very high cloud-top temperature. This feature may be a subsidence line between closed cell stratocumulus. The albedo is high over cumulus Band 1 and over the stratiform cloud closest to the cumulus to the SE.

Fig. 18b shows that a substantial cumulus (Cloud 2) grew up on the SE side of Band 1. This cloud formed over the along-track convergence region observed during the surface leg 20 min earlier. The proximity of this region to the along-track divergence region, observed further to the NW, suggests that Cloud 2 was forced by outflow created by evaporatively cooled air below Band 1, not from the cold pool much further to the SE. The new Cloud 2 was not observed during cloud leg 1 but was apparent as a solid cumulus when penetrated during cloud leg 2. The cloud subsequently ascended and spread to form a massive region with high liquid water content in the NW end of the track. At low levels, Band 1 and Cloud 2 were separated; at high levels, the distinction between them was not apparent. The inversion altitude had increased from 1400 to 1500 m when the aircraft finished flight stack 1. A small isolated cumulus was apparent during the two upper cloud legs near the 45-km mark. The difference between cumulus and stratiform cloud was very apparent in the albedo measurements.

At the beginning of stack 2, the stratiform cloud top height was also at 1500 m. Fig. 18c shows the descent through the stratiform cloud into the extensive drizzle below and through the tops of a band of small cumuli at the 35-km mark. Further to the NW, the aircraft flew into the cold pool generated by extensive evaporation of drizzle falling from Band 1 and Cloud 2. The cold pool setup divergence in the along-track direction, and an extensive region of along-track convergence occurred ahead of it, even further out than the limit of the cold pool. The band of small cumuli at 35 km was at the edge of the along-track convergence zone, and above a region of warm and moist air observed at the surface.

Fig. 18d shows that cumulus Band 3 subsequently grew and dominated the upwards motion in the area. During the previous surface leg, it was only visible as shallow cumulus humilis. During the first cloud leg, it was very fragmented at the penetration level, although some of the cloud elements may have reached the stratiform cloud. During cloud legs 2, 3, and 4, it was solid, indicating that strong growth took place after the aircraft had penetrated the cloud in the lowest cloud leg. The older convection in Cloud 2 was almost in the mature to dissipating stages at this point. The drizzle flux in the lowest cloud leg below Cloud 2 was 0.4 mm h\(^{-1}\), but the cloud base was now raised to a level between cloud leg 1 and 2. The strong drizzle flux from Cloud 2 suggests that the cold pool below continued to develop in the 20 min after the cold pool was first

Fig. 18. Schematic development of the marine boundary layer during the 2.5-h observation period. The flight track is thicker where cloud liquid water occurs. The shading is approximately proportional to drizzle water mixing ratio. The bold dashed line shows the top of the mixed-layer. The divergence and convergence labels refer only to the along-track component of the wind. Relative values of albedo and effective droplet radius \((r_e)\) are also shown.
Fig. 18 (continued).
observed during the descent sounding of stack 2. The outline of the cold pool shown in Fig. 18d is shown as observed 15–20 min earlier than cloud leg 1. It is highly likely that the cold air pool has spread out during this time towards SE and that it forced the strong growth of cumulus Band 3.

With time, one would assume that Band 3 would spread out, and thus be the latest of a series of cumulus that provide moisture for the preservation of the stratiform shield. During the 2.5 h of observations, the boundary layer grew from 1350 to 1500 m, yet the stratiform cloud was still only about 300 m deep. This depth may be limited by both entrainment and drizzle formation. If the stratiform cloud deck was much deeper, then the drizzle formation would be even stronger. This would deplete liquid water and result in a thinner cloud deck.

The time evolution depicted in Fig. 18a–d shows an apparently persistent development of a convective system: cumulus bands form, precipitate, decay, and force new cumulus ahead of them. The convective system moves faster than the ambient wind, leaving processed cloud elements behind. This is a remarkable similarity to squall line systems in the free atmosphere, yet the present cumulus clouds are only 900 m deep in a 1500-m deep marine boundary layer. The observations show that clouds behaving as ‘deep convection’ may occur in the boundary layer: evaporation of precipitation drives mesoscale circulations, which forces new convective elements. The video recordings and photographs show that some of these are band-shaped; for lack of a better term, we will call them boundary-layer squall lines. These bands may have lengths of at least 10 km and occur within cloud fields that the satellite image reveals as closed cell circulations.

The present boundary-layer squall line system is different in a number of ways from the deep convection types. Firstly, the energetics are vastly different. The CAPE in these clouds is only 13 and 3 m$^2$ s$^{-2}$, as determined from the descent soundings of stacks 1 and 2. Deep convection squall lines may be two orders of magnitude more energetic. Secondly, the aspect ratio is high in the present marine boundary layer: the cloud bands measured at low levels are typically 3–5 km wide in a cloud layer of only 900-m depth; i.e. an aspect ratio of about 3:1 to 6:1. Thirdly, the convection appears to move faster through the air than predicted by the theoretical study of Moncrieff (1981). For a propagating system, Moncrieff predicts that the system speed through the air should be in excess of the ambient wind speed by 0.32 $\sqrt{\text{CAPE}}$. For an average CAPE of 8 m$^2$ s$^{-2}$, Moncrieff’s relationship predicts a 1 m s$^{-1}$ speed relative to the surrounding air. The observations shown here (using positions from the lowest penetration of Bands 1 and 3) indicate a speed of 4.5 m s$^{-1}$ relative to the surrounding air.

In a recent paper, Wyant et al. (1997) use a two-dimensional eddy-resolving model to simulate the marine boundary-layer development as air is moving SW in the Pacific Ocean towards the equator. Their simulation shows the cloud development from a solid low level stratus deck to isolated and deeper cumuli closer to the equator over a 10-day period. In the middle of their run, they show a 1250-m deep boundary layer with a combination of cumulus and upper level stratiform cloud. Wyant et al. use a 4-km wide domain for their main model, resulting in one or two cumulus cells; for a 12-km wide domain, three to four cumulus cells were found.

Although the Wyant et al. model study shows many similarities to the present observational study, there are some significant differences. The horizontal scales ob-
served in the SOCEX case study were nearly a factor of 10 larger than in Wyant et al. (1997). This applies to both the cumulus band width (SOCEX 3–5 km; Wyant et al. 0.5 km) and to the distance between cumuli (SOCEX 30–50 km, Wyant et al. < 4 km). This is apparent from both the aircraft observations and from the satellite image. The drizzle rate near the top of the mixed-layer is 5 mm day$^{-1}$ as averaged over cloud leg 1 in stack 2. Wyant et al. show average drizzle rates for boundary layers of comparable depths that are only a few tenths of a millimeter per day. Our observations are only snapshots, and their representativeness is therefore somewhat uncertain, but the differences between our study and the one of Wyant et al. are nevertheless substantial.

5.3. Mixed-layer and drizzle from cumulus

Classical marine boundary-layer models with cumulus convection (e.g. Albrecht et al., 1979) assume a mixed-layer where the top coincides with the LCL. The mixed-layer is assumed well mixed in terms of both energy and vapour mixing ratio. In the present Southern Ocean case, it is also convenient to consider a “mixed-layer” up to the LCL; however, this mixed-layer is at best only well mixed in terms of $\theta_v$, not $q_v$. The localised occurrence of evaporating drizzle in the mixed-layer generates cold air masses, which in turn drive mesoscale circulations. Evaporative cooling is the significant effect that separates the present mixed-layer from classical theory. Locally, where cold pools occur, the “mixed-layer” is stable; in other parts, the mixed-layer is unstable and allows for cumulus mass fluxes out through the top of the mixed-layer.

In a classical cumulus-topped mixed-layer model, there is a small inversion or stable layer at the top of the sub-cloud mixed-layer; Albrecht et al. (1979) called this the transition layer. Positive buoyancy fluxes from the sea surface may drive entrainment fluxes down into the mixed-layer from above this transition layer. When strong drizzle from the cumuli is present, the picture may be considerably more complicated. The observed mixed-layer, outside the cumulus-generated cold pools, has a nearly constant $\theta_v$, but a marked decrease in $q_v$ with height (Figs. 6b and 12b). The figures show an average decrease in $q_v$ of 0.7–1.0 g kg$^{-1}$ over 800 m. Some decrease in $q_v$ with height in the mixed-layer is common due to the primary source of moisture being the sea surface. Another observation is that the LCL, as calculated from the surface leg observations, is about 20–25 hPa below the top of the mixed-layer (Figs. 6a and 12a). The LCL varies along the surface legs; the LCL values chosen here are for segments that are both high in $\theta_v$ and $q_v$, as these are most likely to create the new cumulus clouds.

The cold pools, created by evaporating precipitation in the mixed-layer, are about 1°C colder than the remainder of the mixed-layer. The precipitation flux through the top of the mixed-layer, required to cool the mixed-layer by 1°C, will now be calculated. The mass of an 825-m deep column of air is approximately 900 kg m$^{-2}$. To increase the vapour mixing ratio by 0.4 g kg$^{-1}$ requires a water amount of 0.36 kg m$^{-2}$ or 0.36 mm of rain. The cooling of 1°C can thus result from evaporation of a 1-mm h$^{-1}$ drizzle shower in 20 min.

How rapidly can the temperature in cold pools recover to the values outside cold pools? This can be estimated using a simple mixed-layer model with bulk parameterisa-
tion of surface fluxes, and a simple entrainment assumption at the top of the cold pool based on buoyancy flux arguments. We assume a cold pool that is initially 100 m deep and created by evaporating precipitation. It has an increase in mixing ratio of 0.4 g kg\(^{-1}\) relative to the surrounding mixed-layer, and consistent with this increase in \(q_v\), it is initially 1°C colder than the mixed-layer (see Fig. 19a and b). The sea surface is assumed to be 2°C warmer than the air in the cold pool at the start of the model calculation, and surface fluxes are driven by a 4 m s\(^{-1}\) wind speed. The choice of these parameters is based on the aircraft observations.

Over 3.5 h, the cold pool grows in depth from 100 to 150 m (see Fig. 19). It also increases the vapour mixing ratio by almost 1 g kg\(^{-1}\), such that it is far more moist than the remainder of the mixed-layer. It is thus possible, that cold pools can exist for hours before the temperature recovers to values typical of the rest of the mixed-layer, and during this time, they may force air motions in the mixed-layer.

On February 1, 1995, there were W of Tasmania regions with deep mixed-layers, new cold pool regions, and stages in between the two. In cold pool areas, the surface fluxes were used to recover the temperature profile and to dramatically increase the moisture content close to sea surface. The LCL for the nearly recovered cold pools is often significantly (20–25 hPa) below the top of the mixed-layer. When convergence is

![Fig. 19. Profiles of temperature (\(s/c_v\), where \(s\) is dry static energy (a)) and vapour mixing ratio (\(q_v\), (b)). At the initial time, a cold pool of 100 m depth and 1°C temperature decrease is assumed. After 3.5 h, the temperature reduction is almost eliminated, but the water vapour mixing ratio is vastly increased in the near surface air during this time.](image)
imposed by a precipitation downdraft, then this forcing may be sufficient to allow the air in almost-recovered cold pools to ascend and form new cumuli.

Once significant precipitation falls into the mixed-layer and evaporates, then subsequent precipitation is much more likely. This is because the evaporation in the mixed-layer provides both a dynamical forcing for mesoscale motions in the mixed-layer and a thermodynamic profile that results in the lowering of the LCL in some parts of the mixed-layer, thus making cumuli deeper, more buoyant, and more likely to precipitate.

6. Summary

Aircraft observations of a cloudy marine boundary layer show a closed cell circulation of cumulus growing into, and feeding, a stratiform cloud deck. Three generations of cumulus clouds were observed from a research aircraft during 2.5 h of flight. At least two of these were organised as cumulus bands.

- Variations in the wind speed along the flight track is used to determine one component (the along-track component) of divergence. The along-track convergence/divergence pattern (perpendicular to the bands) appeared closely related to the state of cumulus clouds aloft. Along-track divergence was observed in cold pools under mature cumulus clouds. On the SE side of the cold pools, along track convergence was observed. New cumulus convection was subsequently observed above these regions of along-track convergence. This pattern of triggering new convection on the SE side of mature convection was repeated twice during 2.5 h of observation. In this way, the system of cumulus clouds move through the airmass at a speed faster than the ambient wind speed. Thus, processes resembling “deep convection” may also be responsible for the cloud development in a marine boundary layer of only 1500 m depth.

- The cold pools — created by evaporation of drizzle from mature cumuli aloft — was about 1°C colder and with a mixing ratio 0.4 g kg⁻¹ higher than the remainder of the sub-cloud mixed-layer. These could have been created by a mature cumulus precipitating at 1 mm h⁻¹ for 20 min.

- A simple mixed-layer model is used to calculate the recovery of cold pools. A 100-m deep cold pool, cooled by 1°C and moistened by 0.4 g kg⁻¹, may recover in about 3.5 h. Heat and moisture fluxes from the sea surface results in a further increase of water vapour mixing ratio of about 0.8 g kg⁻¹. This dramatic increase in water vapour in nearly recovered cold pools may partially explain the observation of a decrease in qᵥ of about 1 g kg⁻¹ from the sea surface to near the top of the sub-cloud mixed-layer.

- Nearly recovered cold pools, with their high vapour content, are prime candidates for formation of new cumulus clouds. This is consistent with the observation that the cumulus cloud bases must have been 10–25 hPa below the top of the mixed-layer. The low cloud bases results in higher CAPE for the cumulus clouds. Forcing for the ascent of nearly recovered cold pools may come from other precipitating cumulus clouds, which create a spreading cold pool below them.

- The upper level stratiform cloud was also drizzling strongly, but most of the drizzle evaporated in the layer between the base of the stratiform cloud and the top of the mixed-layer. One effect of this evaporating drizzle was to cool the environment for the cumulus clouds, thus increasing the potential buoyancy of the cumulus clouds.
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Appendix A. Instrumental issues

The accurate measurement of temperature and humidity, both in and outside cloud, is a critical issue for the present study. The F27 has a number of temperature sensors, three of which will be discussed here. The Rosemount sensor is considered the primary sensor. The two others are a NCAR-type reverse flow temperature sensor (Rodi and Spyers-Duran, 1971) with a special fast response resistance wire, and a CSIRO built ‘‘reference wet- and dry-bulb’’ temperature sensor. The wet- and dry-bulb sensors consist of two thin brass tubes, each with a Pt-100 element embedded in a heat conducting compound. One of the brass tubes (the wet-bulb sensor) has a cotton wick on the outside, and this wick is fed water from the cabin. The wet- and dry-bulb reference sensors have a slow response of more than 1 s.

The Rosemount sensor is mounted on the skin of the aircraft, 9 m behind the front of the aircraft. This is under the wing, which could be considered an unfortunate location from an aerodynamic point of view. Yet, it appears that this location effectively keeps the sensor shielded from cloud and drizzle. The reverse flow sensor is mounted near the front of the aircraft and 40 cm off the skin. The wet- and dry-bulb reference sensors are also mounted near the front of the aircraft but only 20 cm off the skin. The brass tubes are mounted perpendicular to the skin of the aircraft and fully exposed, i.e. a housing is not used to slow down the airflow around the brass tubes.

All the temperature sensors have CSIRO-built electronics and no de-icing heaters were used. The temperature sensors were calibrated in a turpentine bath, washed in detergent and rinsed before mounting on the aircraft. The temperature sensors were calibrated against a HP 2801A crystal temperature sensor.

Humidity was measured with two EG&G sensors and the wet- and dry-bulb sensors. The EG&G sensors use the manufacturer’s control circuit and have a CSIRO-built air intake with a suction pump. After prolonged in-cloud flying, they may show evidence of liquid water ingestion; therefore, the EG&G sensors are considered secondary sensors.

All the temperature sensors were corrected for dynamic heating (Lenschow, 1986) using data from speed runs on February 10, 1995. On this day, the wet-bulb sensor was
run dry, such that its recovery factor could be determined along with the recovery factors for the remainder of the temperature sensors. After application of the dynamic heating correction, there was an offset of 0.04°C between the wet-bulb sensor (in dry mode) and the Rosemount sensor. This offset was subsequently incorporated in the data processing, such that the two sensors are aligned.

Evaporation from the wet-bulb sensor is also affected by dynamic heating. The procedures of Lenschow and Pennell (1974) have been used to calculate the true ambient wet-bulb temperature. We have used the Rosemount sensor for the ambient temperature, and assumed that it is not affected by cooling due to wetting. The correction of Lawson and Cooper (1990), which uses pressure ratios instead of air density ratios, was also implemented.

Wetting of immersion temperature sensors is potentially a significant error when measuring temperature inside cloud (Lenschow and Pennell, 1974; LeMone, 1980; Lawson and Cooper, 1990). Cooling errors of 2°C are possible at the F27 flight speed of 80 m s⁻¹ and are thus potentially so large that the thermodynamic analysis in the present paper could be invalid. If the Rosemount sensor is not wet, and if the dynamic corrections to the wet-bulb temperature sensor are correct, then they should both measure the same temperature in solid cloud. The highest cloud legs in stack 1 and 2 are both fully in cloud in the 30–40 km range (see Figs. 9a and 15a). Both of these also contain significant amounts of water in the form of drizzle-sized drops. The average temperature from the Rosemount sensor is 0.03°C higher than the temperature from the wet-bulb sensor for the stratiform leg in stack 1, and the difference is −0.08°C for the stratiform segment in stack 2. These small differences indicate no significant wetting of the Rosemount sensor and a correct dynamic recovery of the wet-bulb sensor.

Although the Rosemount sensor on the F27 does not appear to be subject to wetting errors, it nevertheless still has limitations due to time response. In Sections 3.2 and 4.2, it was noted that the descent soundings showed a stable temperature gradient in the stratiform cloud. Just at the entry to the top of the stratiform cloud, the measured temperature decreased by 2.5–3°C within a few seconds. The Rosemount temperature sensor wire has a time constant less than 0.1 s and can therefore easily resolve the inversion temperature jump. However, the temperature sensor wire is also affected by heat conduction from its supporting construction, i.e. the temperature sensor housing. This metal housing has a much slower time constant, which will also affect the recorded temperature. The dual time constants of immersion temperature sensors have been examined by Rodi and Spyers-Duran (1971). In the present case, with a significant temperature jump across the sharp inversion (3°C), we suspect that the slow temperature response of the housing may have a significant effect right below the inversion, and that the effect gradually becomes smaller as the aircraft descends through the stratiform cloud. Fig. 20a and b shows the details of the two descent sounding θₑ profiles. Also shown are the average θₑ values for parts of the two highest cloud legs for each stack (see Fig. 20a and b). The two selected segments are in the stratiform cloud deck, 32–40 km (stack 1) and 39–56 km (stack 2) on the distance scale. The time lag of the temperature sensor is insignificant during these long level segments, such that the selected leg average values should give a better estimate of the true stability in the stratiform cloud.
The observed gradient of $\theta_v$ from the descent sounding in stack 1 (Fig. 20a) is 5.6°C km$^{-1}$. The corresponding gradient based on the two-leg averages is 3.5°C km$^{-1}$, which is essentially equal to the gradient calculated from adiabatic ascent from the stratiform base (3.1°C km$^{-1}$). It is therefore likely that the stability apparent from the descent sounding is due to instrumental lag and consequently that the true average gradient in the stratiform cloud is close to neutral with respect to moist adiabatic motions. The picture for stack 2 (Fig. 20b) is quite similar. Here, the gradient in $\theta_v$ from the two cloud legs is 4.2°C km$^{-1}$, which again is much closer to the calculated adiabatic gradient than the observed gradient from the descent sounding. Hence, there is considerable evidence to suggest that the actual $\theta_v$ gradient is close to neutral; the apparent stability is most likely due to a slow temperature response of the Rosemount sensor housing.

We have used the manufacturer’s nominal size bins for both the FSSP-100 and 260X data. A check of the FSSP pulse-height voltages showed these to be reasonably accurate. The FSSP data have been corrected for coincidence and dead-time losses (Baumgardner et al., 1985) using the iterative method. No laser beam inhomogeneity corrections (Baumgardner and Spowart, 1990) have been made. The FSSP has consistently 20–30% lower liquid water content than the King and PVM-100A (Gerber et al., 1994) probes. The bias also occurred in samples without drizzle, and is thus not caused by a difference in measurement range. The two other probes were found to show nearly adiabatic liquid water contents in a number of cloud profiles. This suggests that the FSSP has a measurement error, either in the concentration measurement or in the sizing of drops. The analysis of Baumgardner and Spowart (1990) suggests that time response and laser beam inhomogeneities may lead to undersizing by the FSSP probe. A 20–30% error in liquid water content is equivalent to an undersizing error of just 9–11% in drop radius.
The 260X data have been corrected only for depth of field effects using the manufacturer’s values. This probe has been used for calculating drizzle water mixing ratio. The CSIRO-King probe has been used for calculating the cloud liquid water mixing ratio. This probe is also sensitive to small drizzle drops. Biter et al. (1987) have shown that the CSIRO-King probe has full response for drops up to 20 μm radius, and approximately 50% response for drops with 50 μm radius.

References


