A case-study of a partially cloudy boundary layer

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SUMMARY

A case-study of a partially cloudy boundary layer consisting of cumulus and stratocumulus is presented, and the main dynamical processes responsible for its evolution are identified. Measurements were made with the UK Meteorological Office tethered-balloon facility which was located in Hampshire, southern England, during the Doppler Radar Observation Project, September 1998.

Results show that fair weather cumulus formed during the morning of the third and quickly overdeveloped into patches of stratocumulus. A characteristic of this day was a relatively weak capping inversion which allowed a significant amount of entrainment to occur. Between 1100 and 1400 GMT this was calculated at 153 kg m\(^{-2}\). The ratio of entrainment to capping was found to be 0.6.

Data analysis, together with a simple model, shows that capping layer was always significantly increased the stability there and caused a reduction in the cloud area. The data were used to assess the UK Meteorological Office cloud parametrization, and show that during the middle of the day cloud amount was underestimated.

KEYWORDS: Cumulus Stratocumulus Tethered-balloon observations

1. INTRODUCTION

Current interest in boundary-layer clouds, particularly stratiform types, stems from the desire to improve their representation in computer models, both for short-period operational forecasts, and for assessing their role in climate change. To facilitate this, various experimental studies have been conducted. To date, the majority of these have concentrated on marine stratocumulus (Sc) clouds, with a view to assessing their role in climatology. Examples of such experiments include the Joint Air–Sea Interaction Experiment (JASIN; see, for example, Nicholls (1985)), the First ISCCP (International Satellite Cloud Climatology Project) Regional Experiment (FIRE; see Randall et al. (1984)), the Atlantic Stratocumulus Transition Experiment (ASTEX; see Bretherton and Pincus (1995) and the Journal of Atmospheric Science special ASTEX issue (1995, 52, No. 16)), and the Beaufort and Arctic Seas Experiment (BASE; see Curry et al. (1997)). There appear to be fewer studies of stratiform cloud over land than ocean. Examples include Roach et al. (1982), Caughey et al. (1982) and Slingo et al. (1982) who present results from a nocturnal field experiment where a Sc sheet was seen to disperse downwind of the measurement site. Such studies are important for increasing our knowledge of cloud behaviour over land, since that is where most of the short-period forecasting requirements exist (e.g. Clark and Wilson 1997; Keller 1997).

The objective of this paper is to present analyses of a mixed stratocumulus/cumulus (Cu) cloud field measured over land during the Doppler Radar Observation Project (DROP) during September 1998. The situation was characterized by a particularly weak capping inversion which makes this a likely case for observing significant amounts of entrainment (Stull 1988), and its effect on the cloud field. Sc formed from the spreading out of Cu, and typical of this case, is a relatively common boundary-layer type, particularly during cold outbreaks. It can form rapidly in clear skies as observed here, and in some instances can persist for some time (hereafter Cu spreading into Sc will be termed CuSc). Studies of marine-based CuSc were made during ASTEX (Bretherton and Pincus 1995; Martin et al. 1995; Wang and Lenschow 1995; deRoode

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and Duynkerke 1997). In that case the penetration of Cu into a layer of Sc was seen to cause local thickening by increasing the liquid-water content and offsetting the effects of encroachment/entrainment. In the case studied here Cu was also responsible for the initial formation of the Sc layer. The heat budget for a cloud deck over land is also likely to be different to that over ocean during late summer, due to the increased heat input from the surface. This point was addressed by Price (1999), who presented three case-studies of land-based Sc. One of these (a summer case) showed a significant positive heat-feedback loop, where an initially modest sensible-heat input from the surface grew rapidly as warming thinned the cloud, allowing more radiation through, which increased the heating, and so on. A winter cloud deck was also presented; this showed the same effect but to a lesser degree. In this way cloud over land can ‘burn-off’ rapidly. Clearly, identifying the point at which the boundary-layer heat balance becomes positive is important when determining whether a cloud deck is likely to evaporate. The ratio of sensible to latent heat is also likely to be important in this respect. Conversely, negative heat balance is likely to imply cloud will thicken if entrainment and subsidence do not dominate.

Measurements were made with a combination of equipment including the UK Meteorological Office (UKMO) tethered-balloon system, radiosondes and surface measurements. Details of the tethered-balloon system are given by Lapworth and Mason (1988) and Price et al. (1998), though aspects relevant to this paper are discussed below. Section 2 contains a description of the synoptic situation together with a brief description of the cloud evolution and experimental set-up. Section 3 presents measurements of entrainment and encroachment during the episode, and in section 4 the cloud morphology is described in more detail. Section 5 presents a brief comparison of the data with the UKMO cloud scheme, and section 6 contains a discussion and conclusions.

2. Synoptic situation and experimental set-up

(a) Synoptic development

Measurements were made on 3 September 1998 in Hampshire, southern England, during the passage of a weak ridge of high pressure that was sandwiched between two small cyclones. The surface pressure chart for 00 GMT (not shown) shows that the measuring site was on the north-western edge of one of the cyclones. Figure 1 shows the surface pressure at 12 GMT, by which time the site (located at 51.6°N, 1.3°W) was more under the influence of the ridge. Initially, winds were north-westerly and 4–5 m s⁻¹, later decreasing to 3–4 m s⁻¹ and backing westerly by evening. Cloud reports for 06 GMT showed cloudy conditions in the east associated with the cyclone, with brighter conditions further west. Heavy rain fell at the site during the night, but by 08 GMT, when measurements started, conditions were dry. Cloud cover was 8 oktas, mostly altostratus and cirrus, which decreased to 1–2 oktas by mid-day. During the morning the amount of Cu and Sc increased fairly rapidly to 7–8 oktas, and remained so up to 13 GMT, after which it broke to 6 oktas by 14 GMT, and then decayed further to 2–4 oktas of fair weather Cu. The Sc observed was that characteristically formed by the spreading of Cu under an inversion (i.e. CuSc). By 18 GMT only isolated regions of Cu remained which appeared to be convecting as a result of conditional instability. Cloud observations were made directly upwind of the site at 11 GMT and these showed mainly 4–6 oktas of fair weather Cu with uneven bases and tops characteristic of morning development. These were forming under an inversion at 500 m. At this time a few larger clouds existed that had penetrated the inversion and developed up to a second inversion at 910 m. By the time this air reached the measurement site at 14 GMT the lower inversion had warmed
out and most cloud tops were reaching above 1000 m. The cloud amount had increased slightly and showed cloud base and tops that were more even, characteristic of a well developed cloud field.

(b) Experimental procedures

The tethered kite balloon, normally used at Cardington, Bedfordshire, was a 600 m$^3$ helium filled plastic envelope capable of 200 kg net static lift that is typically capable of lifting six or seven turbulence probes to between 1 and 2 km altitude. The turbulence probes are described in detail by Lapworth and Mason (1988) and are capable of measuring mean and turbulent properties of pressure, winds, temperature and humidity. In addition, two Funk or Shultze-Dake type radiometers can be flown and also a total water probe (TWP) which provides mean and turbulent measurements of total water content (liquid plus vapour). This instrument is described by Price et al. (1998).

Measurements usually take the form of either profile runs or level runs. Profiles involve measurements from one or more probes which typically ascend or descend at 0.5 m s$^{-1}$, and traverse the boundary layer in 20 to 40 minutes with a vertical resolution of 2 m. Level runs involve several probes which are placed at fixed heights for periods of one to several hours. A typical arrangement for a level run investigating Sc is illustrated in Fig. 2. Typically, six turbulence probes, two radiometers and the TWP are flown, with a spacing of 30–60 m between probes, two of which are usually placed above cloud top with a radiometer, with at least two others measuring in cloud. The TWP is located below, but close to, cloud-top. A logging rate of 4Hz was used, which is sufficient to resolve turbulent fluxes adequately at these altitudes. During profile runs data were averaged over 16 points to give values every 4 seconds. Profiles made at the main site with the tethered balloon are termed Pxxxx, where 'xxxx' denotes time (GMT).
In addition to the tethered balloon, surface instrumentation was deployed which consisted of one standard Cardington turbulence probe together with total and short-wave radiation measurements, mounted at 4 m. A further facility deployed was a mobile radiosonde unit. This was used to measure profiles upwind so that data could be analysed in a Lagrangian context, allowing advective processes to be eliminated from the data and \textit{in situ} ones quantified. Typical transit times for air parcels between the sonde release site and main site ranged between 2 and 4 hours, which is often long enough to detect significant evolution of the boundary layer. It is noted that for this type of analysis wind shear within the boundary layer will affect the integrity of the air parcel, and its usefulness is thus limited to when values are small. Upwind profiles measured by radiosonde are termed $R_{xxxx}$, where ‘$xxxx$’ denotes time (GMT).

3. \textbf{ENCROACHMENT AND ENTRAINMENT}

\textit{(a) Profile data}

During the experiment, two radiosondes were released upwind of the main site, at 1120 GMT and 16 GMT. Simple consideration of the wind vectors in the upper boundary layer at the sonde and main sites showed that air sampled during the first ascent passed within 15 km of the main site at 14 GMT, and that from the second approached within 25 km at 18 GMT. During the lifetime of the first parcel, conditions were convective,
and the tethered balloon measured significant amounts of turbulence. During that of the second parcel, convection was decaying, again reflected in the turbulence measurements from the tethered balloon (discussed later). It may be expected, therefore, that the first parcel will have experienced more encroachment/entainment. Figures 3(a), 3(b) and 3(c) show temperature, potential temperature and humidity profiles respectively for the 1120 GMT sonde ascent (hereafter profile R1120) together with those from a tethered balloon descent profile (hereafter profile P1400) made at approximately 14 GMT. The data displayed for P1400 are from the uppermost probe. P1400 was timed to coincide with the arrival of air at the main site that was sampled upwind by R1120. The profiles therefore represent the start and end conditions for an air parcel traversing between the two locations (the sonde was released 79 km upwind). Although clouds were present during both profiles, each one traversed a region of clear air. It is clear from Figs. 3(a) and 3(b) that the air parcel warmed significantly during the 2.7 hour period. As discussed in the previous section, the inversion at 500 m on R1120 was already breaking down, with some clouds penetrating from there up to 910 m. This is evident from the surface potential temperature which was already in excess of both that at 500 m and the then inversion base at 910 m. Note that the small magnitude and limited stability of the inversion at 910 m in R1120 has allowed a relatively modest increase of boundary-layer temperature (~2.5 degC, see below) to encroach and significantly raise the inversion. A feature of both profiles is a relatively stable upper boundary layer. The potential-temperature lapse rate (dθ/dz) for R1120 was about 1.4 degC km⁻¹ and that for P1400, 2.5 degC km⁻¹. Apart from subsidence effects it is possible that this stability may have been caused by entrainment of more stable air down from the inversion. The increase in stability observed between the two profiles is consistent with this possibility. Also note that conceptually, above the inversion base, P1400 and R1120 should be the same. The most likely explanation for the difference in height of the stable layer is that it was caused by a local variation. This is supported by observations with the tethered balloon which sometimes show variations of inversion-base height of the order of 100 m (see also Boers et al. (1984)). Variations in potential temperature there are much smaller however, as indicated by comparison of profiles P1100 and R1120, which were made at approximately the same time and 79 km apart (also see later discussion). This result suggests the height variations are caused by adiabatic distortions from impinging thermals or gravity waves. A further possibility, that the air had experienced convergence, is discussed later. Figure 3(c) reveals several points of note. The hydroloape (i.e. the humidity lapse rate) at 500 m on profile R1120 is associated with the small inversion there. As stated in section 2, heavy rain had fallen during the night (typical of recent weather at that location) so that there was a plentiful supply of surface moisture. This is a likely explanation of the moist conditions below 500 m. Profile P1400 shows that once the inversion at 500 m had eroded, humidity at higher levels increased, but it is also apparent that the transport of drier air downwards dried the layer below 500 m. Note on P1400 the significant gradient in specific humidity above 800 m. This is a clear indication that that region was not well mixed with the air below, and is consistent with the temperature profiles which show increased stability from that level upwards. It appears, therefore, that in the clear-air region sampled, vertical overturning throughout the depth of the boundary layer was inhibited.

Table 1 presents mean quantities from all of the profiles made during the day, averaged into lower boundary-layer and inversion-base regions. These data quantify the profile evolutions illustrated in Figs. 3 and 4. The boundary-layer height region was 50–500 m, and the inversion-base data were averaged over approximately 20–30 m,
Figure 3. Vertical profiles as measured by the remote sonde launch at 1120 GMT (profile R1120, dashed line) and the tethered balloon at 14 GMT (P1400, solid line) for (a) temperature, (b) potential temperature, and (c) specific humidity on 3 September 1998.
## TABLE 1. AVERAGED PROFILE DATA FOR LOWER BOUNDARY-LAYER AND INVERSION-BASE REGIONS

<table>
<thead>
<tr>
<th>Profile/time (GMT)</th>
<th>Lower boundary layer</th>
<th>Inversion base</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Temperature (°C)</td>
<td>Potential temperature (°C)</td>
</tr>
<tr>
<td>P1100</td>
<td>13.88</td>
<td>16.7</td>
</tr>
<tr>
<td>R1120</td>
<td>14.0</td>
<td>17.3</td>
</tr>
<tr>
<td>P1400</td>
<td>16.42</td>
<td>19.3</td>
</tr>
<tr>
<td>P1600</td>
<td>17.30</td>
<td>20.1</td>
</tr>
<tr>
<td>R1600</td>
<td>16.8</td>
<td>19.5</td>
</tr>
<tr>
<td>P1800</td>
<td>16.60</td>
<td>19.4</td>
</tr>
</tbody>
</table>

The lifting dew-point is explained in the text. Lower boundary-layer data are averaged between 50 and 500 m. That at the inversion base is over approximately 20–30 m, except specific humidity which is averaged over the first 300 m below the inversion. See text for explanation of column one.

except the specific humidity which was averaged over the 300 m immediately below the inversion to reflect its value in the upper region of the boundary layer. The definition used to locate the inversion base was based on temperature lapse rate in a similar manner to the World Meteorological Organization (WMO) definition of the tropopause (Meteorological Office 1991). The criteria used were: the lowest level for which the lapse rate falls to –2 degC km⁻¹ or less and for which the average lapse rate between that level and all higher levels within 100 m does not exceed –2 degC km⁻¹. The value of 100 m was chosen by judicious inspection of the profiles. It is clear from the Table that the potential temperature, θ, at the inversion increased by 1.8 degC between R1120 and P1400. In addition, the pressure there decreased by about 24 mb. However, it is not possible to attribute this pressure difference to encroachment/entrainment, due to orographic and convergence effects. An estimate of the mass transferred can be made, however, by interpolating the inversion-base temperature from P1400 onto R1120, and is given by the pressure difference between those two levels (see Fig. 5, presented later). This was found to be 15 mb or 153 kg of air per m². This amount of transfer occurred over approximately 2.7 hours, giving a transfer rate of 0.015 kg s⁻¹ m⁻². This pressure difference also represents the vertical depth of the air transferred in the absence of convergence effects (the distance between the height at θ = 18.3 °C on R1120 and θ = 20.1 °C on P1400), however, does include convergence and other effects). That distance was 142 m, which translates into an encroachment/entrainment velocity of 1.5 cm s⁻¹. Note that these calculations assume the pressure difference between adjacent potential-temperature levels measured was representative and not unduly affected by gravity waves (unlikely, due to the unstable layer above the inversion). The actual increase in height of the inversion between the two profiles was 260 m, which means that about 118 m of ascent was due to other processes. Although the most likely explanation of this difference is local adiabatic variation in the inversion-base height, larger-scale convergence could also explain the result. Convergence was expected in this case since the air mass in question was flowing out of the high-pressure ridge into a cyclone and would therefore have experienced increasing vorticity. If so an 118 m increase in height equates to a vertical velocity of 1.2 cm s⁻¹, though clearly this figure is inconclusive.

An error estimate to the above calculation of encroachment/entrainment can be made with reference to the sensor accuracy and how representative each measurement is likely to be. Figures presented in Table 1 are mostly averages from two or more probes. Profile P1100 made the most extensive sample of the boundary-layer inversion. Five probes traversed there over a period of 13 minutes, representing a horizontal length-scale of
about 5 km. The standard error in the measured values of inversion potential temperature was 0.24 degC. Assuming this figure is similar for all profiles, then the probable error in the potential-temperature change calculated between R1120 and P1400 is 0.4 degC. The error in the above encroachment/entrainment calculations, including a weighting for the error in pressure measurement, is 43%.

A second remote sonde, R1600, was launched from a different remote location some 60 km from the main site, with the same aim of assessing boundary-layer development between that profile and profile P1800. During this operation the wind had backed more than expected and the air sampled by R1600 came within 25 km of the main site. Figure 4 is similar to Fig. 3(b) and shows potential temperature against height for R1600 and P1800. R1600 was launched at 16 GMT and at this time extensive and deep (i.e. up to the boundary-layer top) convection was observed there (note the super-adiabatic layer at the surface). The increase in potential temperature in the upper region of the boundary layer seen in P1800 is thus consistent with continued heating. However, P1800 was measured when it was clear that evening cooling was underway. Note the significant surface inversion for P1800, and that cooling appears to have reached about 500 m above ground level. The increase in potential temperature recorded at the inversion between the two profiles (given in Table 1) amounted to 0.7 degC, with a similar error to that above of 0.4 degC. This leaves a probable minimum change of 0.3 degC. Given that little turbulence was observed at the main site after about 1645 GMT it seems likely that the actual encroachment/entrainment was closer to the minimum figure. An increase of 0.7 degC equates to a transfer of 61 kg m\(^{-2}\), and one of 0.3 degC to 26 kg m\(^{-2}\). Because the precise time at which convection ceased is not known it is not possible to calculate an encroachment/entrainment rate.

(b) Turbulence data

Table 2 shows turbulence data taken during LR1, a level run conducted between P1100 and P1400. Most of the run was conducted in 8 oktas of Sc, though the cloud broke during the last few minutes. During this time the air sampled by R1120 was approaching the site. It is not known to what extent the cloud cover developed
whilst in transit, but by the time it reached the site cloud cover there was decreasing to approximately 6 oktas of Sc. All probes, apart from the surface, were in cloud. Unfortunately, a probe located in the inversion above cloud failed at the start of the run. As expected for this type of situation, it can be seen that mean winds decreased towards the surface whilst momentum flux was negative and became more so downwards, with maximum stress at the surface. Temperature and water fluxes measured in cloud by the Cardington probe are usually difficult to interpret, as discussed by Price (1999). This is because, in cloud, the humidity flux, $wq$, becomes the saturated vapour flux, except when measured by the TWP, and the sensible-heat flux, $\omega T$, is influenced by liquid-water changes. Entrainment will normally cause $wq$ measured by the turbulence probes in cloud to be the same sign as $\omega T$, since the saturated value of $q'$ will always follow $T'$ where here the primes denote a deviation from the mean. The sign of $\omega T$ will depend on the change of equivalent potential temperature across the inversion base, and whether mixing of inversion air into cloud heats it, or cools it by evaporation (Lilly 1968; Deardorff 1980; MacVean and Mason 1990), making these measurements more difficult to interpret. They are therefore not presented here. Note that turbulent kinetic energy (TKE) for the run was quite high, constituting a significant source of energy for driving entrainment. The value for $wq$ at 1089 m was measured by the TWP and, considering the TKE there, is quite small. One reason for this was the weak hydrolapse which is typical for weaker inversions. Note the higher water flux at the surface, however, where the moderate values of TKE are accompanied by a strong gradient in specific humidity. The average net radiation measured at the surface during this run was 328 W m$^{-2}$ (presented later) and the energy fluxes from $\omega T$ and $wq$ were 83 W m$^{-2}$ and 135 W m$^{-2}$ respectively. The relatively high value of the latent-heat flux compared with the sensible-heat flux is typical of situations where CuSc is present. The remainder of the net radiation of 110 W m$^{-2}$ can therefore be attributed to the soil heat flux.

Data in Table 2 were used to calculate the heat budget during LR1, using radiative balance, sensible-heat flux and an estimation of the latent-heat flux calculated from the average liquid water. The entrainment heat flux was not considered since it is difficult to measure in cloudy situations (see above). During the run the average increase in potential temperature throughout the boundary layer deduced from the probes was 0.55 degC, and that given from the calculated heat budget was 0.44 degC±0.18 degC. The measured fluxes therefore accurately represent the boundary-layer evolution during LR1. However, if those fluxes are used to estimate the temperature rise between profiles R1120 and P1400 the prediction is 0.75 degC±0.18 degC, compared with the measured value of about 2 degC. The fluxes therefore do not represent the boundary-layer evolution for the air parcel measured by R1120 and P1400. The most likely explanation

<table>
<thead>
<tr>
<th>Height of probe (m)</th>
<th>Wind (m s$^{-1}$)</th>
<th>Momentum flux</th>
<th>Temperature (°C)</th>
<th>Sensible-heat flux</th>
<th>Specific humidity (g kg$^{-1}$)</th>
<th>Humidity flux</th>
<th>Turbulent kinetic energy (m$^2$s$^{-2}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1153</td>
<td>8.50</td>
<td>-0.103</td>
<td>7.94</td>
<td>-</td>
<td>7.45</td>
<td>-</td>
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<tr>
<td>1089</td>
<td>8.39</td>
<td>-0.137</td>
<td>8.06</td>
<td>-</td>
<td>7.55</td>
<td>0.013</td>
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</tr>
<tr>
<td>1032</td>
<td>8.23</td>
<td>-0.149</td>
<td>8.47</td>
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<td>7.75</td>
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<tr>
<td>976</td>
<td>8.05</td>
<td>&quot;</td>
<td>8.93</td>
<td>-</td>
<td>7.78</td>
<td>-</td>
<td>*</td>
</tr>
<tr>
<td>739</td>
<td>6.82</td>
<td>&quot;</td>
<td>10.65</td>
<td>-</td>
<td>8.32</td>
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<td>*</td>
</tr>
<tr>
<td>4</td>
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<td>18.52</td>
<td>0.069</td>
<td>9.57</td>
<td>0.044</td>
<td>1.33</td>
</tr>
</tbody>
</table>

An * denotes insufficient data quality.
for this is that the air sampled by LR1, which was down-wind of that sampled by R1120 and P1400, was more cloudy and thus experienced a much reduced heat flux. The result therefore illustrates the sensitivity of boundary-layer temperature to cloudiness, and cautions the use of localized measurements when attempting to predict the bulk properties of an evolving boundary layer.

Table 3 shows similar turbulence data but for run LR2 (1635–1750 GMT). These data represent clear-air measurements for the period shortly after which convection ceased. The figures therefore show a marked contrast to those in Table 2. Turbulence values are in most cases not significantly different from zero. Note, however, that the heat flux measured by the balloon-borne probes (but not at the surface) are negative, which is consistent with cold air from the surface being mechanically forced upwards by surface roughness-driven turbulence (see Fig. 4).

(c) Effect of enroachment/entrainment on boundary-layer structure

It is clear from the preceding sections that enroachment/entrainment were significant processes during the middle part of the day. Measurement of stability (dθ/dz) in the upper part of the boundary layer showed that it had increased from 1–2 K km\(^{-1}\) during R1120 to 2–3 K km\(^{-1}\) for P1400. Since results suggest the air mass experienced convergence, which would decrease stability, the increase must be due to entrainment transporting potential temperature downwards and increasing values above those created by enroachment. In this subsection an attempt is made to recreate this effect in a model, with the emphasis on simplicity. A large number of entrainment models and parametrizations have been developed to date, including Lilly (1968), Deardorff (1974, 1979), Tennekes and Driedonks (1981) and Lock and MacVean (1999). A summary of many of these theories is presented by Stull (1988), and Fernando (1991). The approach here is similar (though simplified) to those discussed by Fernando (1991) and Breidenthal and Baker (1985), which both considered the effects of thermal vortices impacting and penetrating the inversion structure. It is consistent with the established view of the entrainment flux being proportional to the surface heat flux for a convective boundary layer.

Figure 5 shows a schematic diagram of an idealized boundary layer and inversion potential-temperature structure, shown by the solid line. Heating of the boundary layer will move the line segment aa' to the dashed line bb'. In addition, thermals reaching the inversion at 'b' will overshoot to say 'c', where they will mix to varying degrees with the environmental air. Note that the equilibrium position of this air will be above point 'b', and it will act to cool the air between 'b' and 'c'. The TKE associated with the overshoot, however, is likely to recoil and mix some of this air down beneath point 'b', creating the profile cdb'. An important question raised here is how permanent are
these changes to the profile? It is clear, for example, that considering the impact of a single thermal plume on an otherwise undisturbed inversion, air on line cd above bc will be surrounded by warmer air and will thus sink until along line bc. Similarly, air mixed along cd below bc will be surrounded by cooler air and it will therefore rise until it also lies along bc. At some point therefore (once the thermal has dissipated), the original profile will re-establish itself, and the net effect will have been to transfer some boundary-layer air into the inversion, but no inversion air into the boundary layer. Once the density of thermal plumes increases however, the horizontally averaged profile will appear more like cd′, and the buoyancy forces described above will be much reduced. Significantly less re-adjustment back to the original profile will occur and the transfer of air across the inversion base will be mostly permanent. Encroachment of inversion air into the boundary layer can be defined as transfer caused by heating of the boundary layer such that it becomes warmer than a portion in the inversion. The latter air then becomes negatively buoyant and sinks into the boundary layer. In contrast, entrainment is transfer caused by dynamic forcing of inversion air down into the boundary layer (e.g. caused by overshooting thermals), where it mixes with the environmental air. The resulting air normally retains some of the inversion’s buoyancy and is thus likely to remain near the top of the boundary layer. Using these definitions the encroachment length-scale can be quantified by $z_b - z_a$, and the entrainment mixing-length-scale by $z_c - z_b$ where $z$ is height and the subscripts refer to the locations marked in Fig. 5. Clearly the proportion of inversion air in the entrainment zone is maximum at ‘c’ and decreases downwards. Note, however, that the inversion base is not necessarily at ‘c’, since the stability criteria used to define it in section 3(a) may place it anywhere along
the line dc, allowing some boundary-layer air to transfer into the inversion, as described above. The encroachment length-scale is given by:

\[ l_{\text{enc}} = (\theta_b - \theta_i) \left( \frac{d\theta}{dz} \right)^{-1} \]  

(1)

where \( \theta_b \) and \( \theta_i \) are the (lower) boundary-layer and original inversion-base potential temperatures respectively. Given the initial profile and heating rate, \( \theta_b \) can be estimated. In this study it is taken from Table 1. Estimation of the entrainment mixing-length-scale can be made from consideration of the momentum of the rising plume and the inversion stability. Neglecting dissipation, the potential energy at the top of the overshoot will be equal to the kinetic energy at point 'b' in Fig. 5. Assuming a linear temperature profile the potential energy at point 'c' is therefore:

\[ \text{PE} = \frac{1}{2} g (\rho_i - \rho_c) V l_{\text{ent}} \]

where \( g \) is the gravitational constant, \( \rho_i \) is the density of the overshooting thermal, \( \rho_c \) is the density at point 'c' (see Fig. 5) and \( V \) is the volume of air under consideration. Using \( (\rho_i - \rho_c) = l_{\text{ent}} d\rho/dz \), substituting for \( d\rho/dz \) by differentiating the equation for an ideal gas and including the adiabatic lapse rate, we obtain (per unit mass):

\[ \text{PE} = -\frac{1}{2} \frac{g}{T_b} l_{\text{ent}}^2 \left( \frac{dT}{dz} + \Gamma \right) \]

where \( \Gamma \) is the dry adiabatic lapse rate and \( T_b \) the temperature at point 'b'. Since

\[ \frac{d\theta}{dz} = \frac{\theta}{T} \left( \frac{dT}{dz} + \Gamma \right) \]

then

\[ \text{PE} = -\frac{1}{2} \frac{g}{\theta_b l_{\text{ent}}^2} \frac{d\theta}{dz}. \]  

(2)

An estimation of the kinetic energy can be made with the commonly used scale velocity \( W_\ast \) (Arya 1988), defined as:

\[ W_\ast = \left( \frac{g}{\theta \rho c_p} h \right)^{1/3} \]

where \( H_0 \) is the surface heat flux, \( c_p \) is the specific heat at constant pressure, and \( h \) the height of the boundary layer. The kinetic energy per unit mass is then given by:

\[ \text{KE} = \frac{1}{2} \left( \frac{g}{\theta \rho c_p} h \right)^{2/3}. \]  

(3)

Finally, equating Eqs. (2) and (3) gives an expression for the entrainment mixing-length-scale:

\[ l_{\text{ent}}^2 = \theta_b g^{-1/3} \left( \frac{h}{T_0 \rho c_p} \right)^{2/3} \left( \frac{d\theta}{dz} \right)^{-1} \]

(4)

where \( T_0 \) is the surface/screen temperature. The height of point 'c' in Fig. 5 is therefore given by \( z_i + l_{\text{enc}} + l_{\text{ent}} \), where \( z_i \) is the height of the inversion. The question arises as to the nature of the curve cd in Fig. 5. In this study it is taken to be exponential type,
and scaled with \( l_{\text{ent}} \). This is an arbitrary choice, but produces a curve of similar shape to those often seen in profiles of potential temperature. The increase in \( \theta \) at any level below point ‘c’ due to mixing down of inversion air (denoted by \( \theta' \)) is then given by:

\[
\theta'_z = (\theta_c - \theta_b) b^{-\Delta z/l_{\text{ent}}}
\]

(5)

where the value of \( b \) is chosen empirically and \( \Delta z \) is measured downwards from point ‘c’. Given an initial potential-temperature profile, the effects of boundary-layer heating can be added by first including the encroachment, then the entrainment perturbation given in Eq. (5):

\[
\begin{align}
\text{If } \theta_z < \theta_b & \quad \text{then } \theta_z = \theta_b \\
\theta_z = \theta_z + \theta'_z.
\end{align}
\]

(6a)

Thus, (6a) produces the line bb’ in Fig. 5 and (6b) adds the curve cd. This algorithm was applied to R1120, and heating rates were obtained from comparison with P1400. The situation for this data was complicated by the presence of cloud, which will affect the value of \( H_o \) in Eq. (4) by supplying an additional source of energy to the sensible heat. The latent-heat energy released can normally be estimated from the liquid-water flux at cloud top. However, the error in this calculation is estimated as 0.015 g kg\(^{-1}\) m s\(^{-1}\), and the result was not significantly different from zero. An alternative method can be used, using the total water flux presented in Table 2, which consists of both liquid and vapour fluxes. The liquid-water flux can be approximated by the ratio of the standard deviations of the liquid, \( \sigma_l \), and total water, \( \sigma_t \), time series multiplied by the total water flux, so that the latent-energy flux is given by:

\[
E_l = \frac{\sigma_l}{\sigma_t} \rho L w' q'_t
\]

where \( L \) is the latent heat of vaporization, \( q_t \) is the total water content and the overbar denotes an average value. The ratio of standard deviations \( \sigma_l \) to \( \sigma_t \) was 0.63 giving a latent-heat flux of 21 W m\(^{-2}\), which was added to the sensible-heat flux of 83 W m\(^{-2}\) to give \( H_o \). Results are plotted in Fig. 6, which shows profiles of potential temperature against height, and is similar to Fig. 3(b). The dashed line is the result of the above algorithm being applied to R1120, with a value of 1.5 for \( b \) in Eq. (5). An additional operation has been applied to the data to compensate for the discrepancy in height of the inversion between R1120 and P1400 caused by effects other than encroachment/entrainment (as discussed above). The height-scale of the calculated profile has thus been expanded by 8% (otherwise the profile would merge with R1120 at \( z_i + l_{\text{enc}} + l_{\text{ent}} \)). It can be seen that the profile shows reasonably close correspondence with P1400, and that variations between the two are mostly below 0.2 degC, except in the super-adiabatic layer below 150 m. Table 4 presents the entrainment statistics presented in section 3(a) with those derived from the above model. The maximum potential temperature for thermal overshoot predicted by the model was 20.2 degC, very close to the observed value on P1400 for the inversion base. However, due to an unstable region immediately above there, the definition for inversion-base height given in section 3(a) placed it at \( \theta = 20.4 \) degC, 0.3 degC higher than observed with P1400. The overestimate of the transfer predicted is therefore mostly due to the shape of R1120, rather than the model. It can be noted, however, that the value still lies within the 43% error quoted for the observed transfer. The ratio \( l_{\text{ent}}/(l_{\text{ent}} + l_{\text{enc}}) \) gives the fraction of transferred air due to entrainment. Again
the model estimate is larger, and both figures are a little larger than the generally accepted value of 0.1–0.2 (e.g. Boers et al. 1984; Stull 1988). This is most likely due to the unusually weak stability of the inversion, allowing a large value of $l_{\text{ent}}$.

Clearly the scope of the model is limited; it does not consider entrainment mechanisms such as wind shear or breaking gravity waves. In addition, the factor $b$ in Eq. (5) may vary from case to case (this factor may be better described as a function of stability). It does, however, provide support for the idea that increasing stability in the upper part of the boundary layer was caused by entrainment, and that the observed amount of air transferred was similar to what would be expected in a system driven mainly by surface-driven convection. It is noted in the next section that the increasing stability may have had a significant effect on the cloud evolution.

4. CLOUD EVOLUTION

In this section the development of the cloud layer is quantified. The bulk of measurements made in cloud were during LR1, the figures from which are presented in Table 2. During this 1.25 h run five turbulence probes performed measurements over a height range of 400 m. At the start of the run the probes had recently entered a rapidly developing band of Sc. Figure 7 shows a contour plot of relative humidity over this range, for LR1. It can be seen that cloud was at least 400 m thick during the first half

---

**Figure 6.** Similar to Fig. 3(b). Solid line is profile R1120; dotted is P1400, averaged from several probes; dashed is profile predicted by the model as described in the text.

**TABLE 4. OBSERVED AND MODEL-DERIVED ENTRAINMENT PARAMETERS**

<table>
<thead>
<tr>
<th>$\theta_b$ (°C)</th>
<th>$l_{\text{enc}}$ (m)</th>
<th>$l_{\text{ent}}$ (m)</th>
<th>$l_{\text{ent}}/(l_{\text{ent}} + l_{\text{enc}})$</th>
<th>Transfer (kg m$^{-2}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Observed</td>
<td>20.1</td>
<td>100</td>
<td>40</td>
<td>0.29</td>
</tr>
<tr>
<td>Derived</td>
<td>20.4</td>
<td>100</td>
<td>62</td>
<td>0.38</td>
</tr>
</tbody>
</table>

$\theta_b$ = inversion-base potential temperature, $l_{\text{enc}}$ = encroachment length-scale, $l_{\text{ent}}$ = entrainment length-scale and 'Transfer' is the air mass transferred into the boundary layer.
of the run (cloud top was measured at about 1200 m at the start of the run). At \( t = 12.8 \) (where \( t \) is time in hours (GMT)) a sharp rise in cloud base to 1000 m is evident, it then rises more slowly until \( t = 13.17 \) when all of the probes exit cloud. Note that at \( t = 12.93 \) and \( t = 13.16 \) cloud base decreases sharply for a short time. At these locations the probes have detected small Cu clouds which were penetrating into the main Sc sheet. Before \( t = 12.8 \) the cloud was therefore thicker and more uniform, and after this the probes detected a different regime where cloud was more variable.

Figure 8(a) shows the liquid-water content measured at about 1090 m during the run. The error in these values is 0.2–0.3 g kg\(^{-1}\). Comparison with Fig. 7 shows that in the region of thick cloud before \( t = 12.8 \), peak liquid-water concentrations reach 0.8–1.0 g kg\(^{-1}\). Note that the larger values occur in the ‘middle’ of this part of the cloud field, with smaller values on either side. This suggests the main area of uplift, and newly formed cloud, is at the centre of the cloud, and that air towards the edges is older, has mixed more with the environmental air, and experienced partial evaporation. Visible in this region is an apparent oscillation of liquid water (\( t = 12.4–12.7 \)), which is likely to represent the structure of the convective plumes (it does not correlate with changes in probe height). The scale of these variations is about 900–1000 m, just slightly less than the boundary-layer height. After \( t = 12.8 \) (the negative spike there is anomalous, caused by a sensor response lag when the turbulence probe temporarily exited cloud) liquid-water content was mostly insignificant. This is expected in the much thinner cloud, where the TWP was located near cloud base. The exception to this was between \( t = \sim 12.9–13.0 \) where the Cu penetration increased values significantly. Note that the second Cu penetration at \( t = \sim 13.16 \) did not make any impact on the upper layer. Either
the main plume had not yet reached there, or the cloud was too small to produce an impact. Figure 8(b) shows net radiation measured at 4 m during LR1. From $t = 12.5$ onwards there was a general increase in values consistent with a general decrease in cloud cover. Note at $t = 12.45$ the minimum value of radiation is coincident with the maximum measured liquid-water content, and that minima are recorded during the presence of the two Cu clouds discussed above. Maxima in radiation occurred when the cloud almost broke at $t = 12.8$, and after $t = 13.2$, when the cloud had moved away. Note that the correlation does not extend to smaller time-scales, however, probably because the probes represent a spot measurement and the radiation is area averaged.

Table 1 shows data for the lifting dew-point, ITd. This is defined as the dew-point that lower boundary-layer air would have if transported directly to the inversion base. Thus, in a convective situation, comparison with the in situ inversion-base temperature gives an indication to whether cloud is likely or how much heating/cooling is required to disperse or form cloud. Not surprisingly the inversion-base temperature for cloudy profiles is less than ITd, but it can be seen from Table 1 that was also the case for the clear-air profiles. The lack of cloud there was therefore due to the presence of drier air and not higher temperature. Comparison of upper boundary-layer specific humidity for R1120 and P1400, between which significant encroachment/entrainment was observed, shows that the air became drier which (since the air at lower levels was moister) must have been due to encroachment/entrainment of drier air from above. Thus the absence of cloud in the clear profiles observed appears to have been due to encroachment/entrainment (this result was also found for a case described by Price (1999)).
A further mechanism by which entrainment is likely to influence the development of cloud is by modifying the stability profile in the upper boundary layer, as shown in section 3(c). Increasing potential temperature there will mean that fewer of the thermal plumes leaving the surface (assuming they have a range of potential temperatures) will reach a given level, say the dew-point level, and therefore cause a decrease in cloudiness. For example, the lifting condensation level for P1400, taking a lower boundary-layer specific humidity of 8.49 g kg\(^{-1}\), was at 1070 m. It can be seen by reference to Fig. 6 that the potential temperature at that level was about 0.7 degC higher than the lower boundary-layer value of 19.3 degC (Table 1). Thus, for thermal plumes rising from the lower boundary layer to reach that height, and form cloud, they must be 0.7 degC warmer than if the profile were adiabatic, and therefore fewer of them.

A further observation evident in Table 1 is that inversion-base temperature decreased during the day, including the period between R1120 and P1400 when the boundary layer was warming. In a typical boundary layer where the inversion is more stable and \(dT/dz\) is positive, inversion-base temperature normally increases in line with that of the boundary layer. However, in this case, inversion \(dT/dz\) was slightly negative, and therefore inversion-base temperature decreased as encroachment progressed. This has implications for cloud evolution since the persistence of cold temperatures at the inversion base will have acted to perpetuate the presence of cloud. Thus in this case warming of the boundary layer did not cause the observed decrease in cloud cover during the afternoon, contrary to more normal anticyclonic conditions. The observed decrease was thus most likely due to a combination of increasing upper boundary-layer stability and weakening solar forcing. Decreasing cloud amount clearly requires evaporation of existing cloud and could be achieved by mixing with adjacent clear air. The fraction of clear air \(\chi\) required to evaporate a cloud completely is given by:

\[
q\chi + q_t(1 - \chi) = q_s
\]

where \(q\), \(q_t\) and \(q_s\) are the specific humidities of adjacent clear air, total water content and saturation values in cloud respectively. As \(\chi\) approaches 1, evaporation of the cloud becomes less likely. Data from Table 2 were used to estimate \(\chi\) at 1089 and 1153 m altitude (using the value of \(q_t\) at 1089 m). \(q_t\) was taken from the final part of the data series when the probes had moved into clear air immediately adjacent to cloud, and \(q_t\) was averaged between \(t = 12.4\) and 12.7 when the cloud was thickest (see Fig. 7). These values give \(\chi\) ranging between 0.49 and 0.65, i.e. the cloud must mix with between one and two times its volume before it will evaporate. Given the near clear conditions in the evening (1 okta), the time-scale for such mixing appeared to be up to a few hours.

5. UK Meteorological Office Cloud Scheme

In this section a brief appraisal of the current UKMO operational cloud scheme is presented, using data from LR1 and LR2. Details of this scheme are presented by Smith (1990), but here that part which calculates the cloud fraction is discussed. The estimation of cloud fraction is based on the grid box mean liquid-water content \(Q_c\), and its standard deviation \(\sigma\), given by Smith (1990) as:

\[
Q_c = a_1(q_w - q_{sat})
\]

\[
\sigma = a_1 \left( q_{sat}^2 - 2a_1 q_w \bar{T}_1' + a_1^2 \bar{T}_1'^2 \right)
\]

where

\[
a_1 = \frac{1}{(1 + L\alpha_i/c_p)}, \quad \alpha_i = \frac{\partial q_{sat}}{\partial T},
\]
TABLE 5. Cloud fraction calculated from measured data, observed (by eye), and calculated from the model parametrization of Smith (1990), for level runs LR1 and LR2

<table>
<thead>
<tr>
<th>Run</th>
<th>Data</th>
<th>Observed</th>
<th>Model</th>
<th>L</th>
</tr>
</thead>
<tbody>
<tr>
<td>LR1</td>
<td>0.75</td>
<td>0.81</td>
<td>0.51</td>
<td>37.8</td>
</tr>
<tr>
<td>LR2</td>
<td>0</td>
<td>&lt;0.125</td>
<td>0</td>
<td>30.1</td>
</tr>
</tbody>
</table>

$L$ is the scale length of each run (km).

$T_l$ is the liquid-frozen water temperature, $q_w$ is the total water content, and $q_{sat}$ the saturated vapour value of specific humidity. The cloud fraction $C$ is then given by:

$$
C = \begin{cases} 
0, & Q_c \leq -\sqrt{6} \sigma \\
\frac{1}{2} \left( 1 + \frac{Q_c}{\sigma \sqrt{6}} \right)^2, & -\sqrt{6} \sigma < Q_c \leq 0 \\
1 - \frac{1}{2} \left( 1 - \frac{Q_c}{\sigma \sqrt{6}} \right)^2, & 0 < Q_c \leq \sqrt{6} \sigma \\
1, & Q_c \geq \sqrt{6} \sigma 
\end{cases}
$$

(9)

The model-predicted cloud fraction was calculated using (8a), and (9). $\sigma$ was estimated directly from the time series of $Q_c$. Results are shown in Table 5 for comparison with observed values. The value derived from the data time series is the ratio of the time spent in cloud to the duration of the run, and the observed value was that estimated by eye over a larger portion of the sky.

It can be seen from Table 5 that the observed and data values of cloud fraction are very similar for LR1, but that the model has significantly underestimated cloud cover by about a third. This result is consistent with the general behaviour of the scheme which tends to underestimate cloud fraction (Lorenc et al. 1996). For the non cloudy case during LR2, however, the model has predicted the cloud fraction successfully. Note the length-scale for each run (i.e. the linear dimension of air sampled during the run) was similar and is approximately equal to two grid box lengths of the UKMO mesoscale model.

6. Summary and conclusions

This case-study represents a moderately complicated situation where a deck of Sc formed from the spreading out of Cu during the morning from clear conditions, and subsequently decayed back to similarly clear conditions by evening. This type of situation is relatively common, though in many cases the Sc cloud is more persistent. A characteristic of this day was the relatively weak capping inversion which allowed boundary-layer convection to penetrate it significantly, causing substantial entrainment. It is tempting to suppose that, because of this, the calculated entrainment rate of 0.015 kg m$^{-2}$s$^{-1}$ or 1.5 cm s$^{-1}$ represents a higher bound to convectively driven entrainment for late summer mid-latitudes. Results show that this entrainment increased the stability in the upper part of the boundary layer, and that this was likely to have reduced the cloud fraction. A simple model, which uses a mixing length calculated from the estimated overshoot of thermal plumes into the inversion, showed reasonably close agreement with the observed profile, supporting the case that the increased stability
was due to entrainment. The model estimate of mass transfer was also similar to that observed.

It was shown in section 4 that inversion structure can have a significant impact on the evolution of inversion-base temperatures, and thus cloud. In this case $dT/dz$ in the inversion was slightly negative which meant that as encroachment proceeded the inversion-base temperature decreased slightly, enabling cloud to persist despite a positive heat balance. It was discussed in section 1 that identifying whether the boundary-layer sensible-heat balance is positive is a prerequisite to inferring cloud warming and evaporation. However, it can be seen from this case that $dT/dz_1 > 0$ is also required for this to occur. Warming of the boundary layer could not therefore evaporate the cloud, as in the more common case where $dT/dz_1$ is positive. Another explanation is therefore required to explain the Sc dissipation. Observations show that Sc formed from the spreading out of Cu often does so in relatively inhomogeneous conditions. This is because the penetrating Cu usually form initially under a lower inversion where the humidity is rather higher than above. When these clouds penetrate up to the higher level they therefore introduce higher humidity there and create (or enhance) an inhomogeneous humidity field. The fate of the Sc formed depends on the distribution of this field. Clearly cloud air will mix with the local drier air, and if it is sufficiently dry will evaporate all of the cloud. The data for this case show that this could occur if the cloud mixed with between one and two times its own volume of upper boundary-layer air, and this is the most likely explanation for the cloud dissipation observed in this case. The lower values of liquid water present in the Sc during the latter part of LR1 suggest that part of the cloud was dissipating. This process is in addition to entrainment of inversion air directly into cloud top, which will also act to dissipate the cloud. The results do suggest that entrainment played a role in cloud dissipation by mixing slightly drier air down into the boundary layer. Clearly if the upper boundary-layer air had been significantly more humid the cloud would have been more persistent. Cloud dissipation may start once the thermal plume supplying the moisture decays. The turbulent energy available for mixing is therefore likely to decrease from that point, but it is noted that an additional source of energy could be cloud-top radiative cooling, which will encourage vertical mixing in particular (it can be noted from scrutiny of a tephigram, for example, that cooled cloud-top air, in the absence of lower-level inversions, is likely to sink to the surface).

When considering the resulting cloudiness caused by CuSc, it is clear that the precise morphology of the system is of significant importance. Given a fixed mass flux in a Cu updraught, for example, the depth of the detrained Sc will govern the resulting cloud area created, and may form either a large area of thin cloud, or a smaller area which is thicker. In addition, detrained Sc will be optically thicker than that formed in situ. Current fine-scale models are capable of reproducing this type of structure, though it is clear that the simulation must be very accurate to produce the correct resultant cloud field. The accuracy of such simulations does not appear to have been widely verified so that experimental data which can reveal the morphology of individual cloud cells are likely to be of significant benefit.

The requirement for detailed information means that forecasting this type of cloud is not straightforward. A forecaster or operational model, for example, will normally have very limited observational data available, such as a single radiosonde ascent for a particular area which cannot represent horizontal variability properly. An estimation can be made, however, for this particular case, if it is assumed that the humidity of updraughts (and therefore cloud) penetrating to the upper boundary layer is equal to that in the lower boundary layer, and that its distribution is given by the cloud fraction of
the Cu clouds. The fate of Sc can then be estimated as described above using Eq. (7). Estimating the time-scale for evaporation of the cloud is more difficult. Some schemes (e.g. Tiedtke 1993) estimate an evaporation rate from the saturation deficit of the environmental air and a fixed diffusion coefficient. In practice the diffusion coefficient will be a function of the turbulent energy so that a time-scale for cloud evaporation might take the form of:

\[ t = \frac{a l \chi}{U' (1 - \chi)} \]

where \( a \) is an empirical constant, \( l \) is the mixing-length scale, \( U' \) the turbulent velocity and \( \chi \) is the fraction of clear air the cloud must mix with to evaporate, given in Eq. (7). Note that as \( \chi \) tends to 1 the evaporation time becomes infinite.

Data were used to test the UKMO model cloud parametrization. It was found that during cloudy conditions the parametrization underestimated the cloud by about 2 oktas, which is often the case with this scheme. The scheme correctly estimated clear conditions during the evening. The current model is purely diagnostic, and as such it does not represent cloud dissipation explicitly, as described above. Once layer cloud is present, it will remain until removed by a subsequent model iteration or assimilation (in this case, for example, it could potentially overestimate cloud fraction during the evening). As such it is likely to benefit from the addition of a cloud-dissipation scheme. A more extensive assessment of model performance will be conducted using tethered-balloon data in future work.

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