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The influence of entrainment on the evolution of cloud droplet spectra: II. Field experiments at Great Dun Fell

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SUMMARY

Measurements have been made of the liquid water content and droplet-size distributions within clouds enveloping the summit of Great Dun Fell, Cumbria. These covered extensive periods and a variety of meteorological situations.

It was found that in highly stable atmospheric conditions cap clouds contain narrow drop-size distributions and near adiabatic liquid water content. In less stable conditions mixing with the cloud environment was found to have a substantial effect on the microphysical properties of the cloud.

In the cases studied the admixture of undersaturated air was found to produce a substantially sub-adiabatic liquid water content and the spectral shape experienced broadening in a way more consistent with the extreme inhomogeneous model of mixing described in Paper I* than with the classical treatment. In particular, it was found that a small number of large drops experience a greatly enhanced growth rate.

In a particular study, it was observed that the interaction of the cap cloud with a pre-existing strato-cumulus deck produced a broad drop-size distribution and an enhanced liquid water content.

1. INTRODUCTION

In this paper we present findings from the analysis of meteorological and microphysical data obtained in field experiments conducted in December 1977 and February 1979 from the UMIST field research station on the summit of Great Dun Fell (GDF) in Cumbria. These were performed in collaboration with scientists from the Meteorological Office, but the comprehensive droplet measurements made by the latter are not discussed here.

The primary objective of this analysis is to examine the changing relationships between the meteorology, the drop-size spectra and liquid water content over continuous extensive periods (1630–1700h on 12 December 1977, 1138–1824h on 14 December, 1930h on 15 December to 0021h on 16 December, 1000–1115h on 16 December 1977 and 1100–1300h on 28 February 1979). All times are GMT. A particular goal is to establish whether, in some circumstances, there exists evidence for the influence, on the spectrum and on fluctuations in the liquid water content, of mixing-in of environmental undersaturated air and if so, to learn more about the true nature of the mixing process by assessing this evidence in the light of both the classical and extreme inhomogeneous descriptions, presented in paper I. It is probably useful to mention here that in the absence of mixing the droplet spectra measured within clouds enveloping the research station at GDF are narrow and generally close to that displayed in Fig. 1 (curve A), the liquid water content being close to adiabatic.

Measurements of cloud droplet spectra were made using an electrostatic disdrometer (Keily probe) mounted in a wind tunnel, as described in the paper by Corbin *et al.* (1978), in which it was shown that this device functions reliably in ground-based usage and that its

* *Quarterly Journal*, July 1980, pp. 581–598 (Baker, Corbin and Latham).

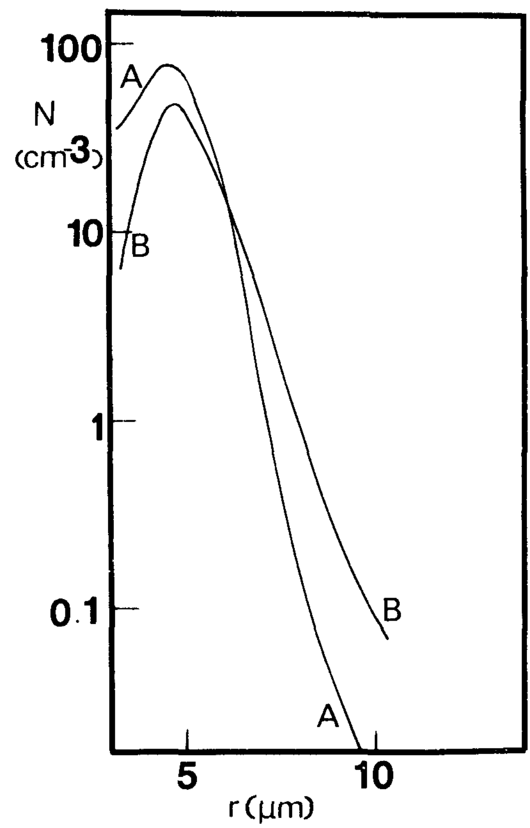

Figure 1. Measured size distributions at A, 2200h ($L = 0{\cdot}08\,\mathrm{g\,m^{-3}}$) and B, 1930h ($0{\cdot}003\,\mathrm{g\,m^{-3}}$) on 15 December 1977. (For the purposes of comparison the drop concentrations for B have been multiplied by 10.)

absolute collection efficiency can be determined for a prescribed set of operating conditions. In these field experiments a ventilation speed of $20\,\mathrm{m\,s^{-1}}$ and an orifice diameter of $440\,\mu\mathrm{m}$ were employed. The disdrometer system was located 50 m from the buildings on the summit of Great Dun Fell, in such a position that there were no obstructions upwind.

For the greater part of the experimental period the electrostatic disdrometer was employed simultaneously with the PMS ASSP device operated by the scientists from the Meteorological Office. It was reassuring to find, from subsequent analysis of data, that the major features of the droplet spectra revealed by both devices were essentially identical at all times.

The electronic pulses produced by droplets entering the Keily probe were amplified, linearized with respect to droplet radii and then recorded on magnetic tape. These tapes were replayed in Manchester, in conjunction with a PDP8e computer, to produce size histograms and tabulated values of liquid water content L, dispersion σ (defined as the ratio of the standard deviation to the mean size), droplet concentration N, and visibility V. The minimum sensible period of integration was estimated to be 2 seconds, and since the wind-speed during the course of the experiments was typically around $10\,\mathrm{m\,s^{-1}}$ information was obtained on fluctuations in the water properties on ali scales in excess of about 20 m. In some situations, when conditions were steady and we wished to obtain good statistics on the concentrations of the larger droplets, the Keily probe pulses were integrated over several minutes. It should be recollected, when examining the spectra presented in the following sections, that the Keily probe cannot detect droplets of radius less than 4 μm.

Standard measurements of temperature, dew-point and wind speed and direction were made by the Civil Aviation Authority from their meteorological station situated on the summit of Great Dun Fell at 0900h, 1200h, and 1500h on each day. At other times measurements of temperature and wind speed and direction were made as appropriate using our own equipment.

2. Methods employed in the meteorological analysis

The details of the analysis are dependent on the conditions during the particular period in question. Specific points are made at the appropriate places in the ensuing discussion. The general approach is described here.

Trajectories of the air to GDF were constructed from synoptic analyses available for the period of the experiment. As a test for accuracy, trajectories to neighbouring areas were sometimes constructed also – this assisted the computation of the arrival times of air of different properties. Vertical cross-sections, across Britain, of temperature, dew-point, wind speed and direction, wet bulb potential temperature, etc., were constructed as appropriate. These were used, together with the trajectories, to estimate the vertical profiles of temperature, dew point, wind speed and wind direction close to the mountain. Further evidence was obtained from reports from nearby land stations, particularly Leeming, Carlisle, Ringway, Ronaldsway, Newcastle and Squires Gate, depending on the air trajectories. These stations together with VHRR and IR satellite photographs obtained from the University of Dundee were used, again in conjunction with the ascent data, to estimate the general cover, type, height and depth of cloud cover in the area of the mountain.

The effect of the mountain on the meteorological parameters was estimated using a simple computer model in which the ascent data obtained from various stations (close to sea level) were extrapolated to the mountain summit. It was assumed, following Collier (1975), that the vertical motion field in the planetary boundary layer was dominated by orographic upslope motion. This gives vertical velocities of typically a few metres per second with which only free convection due to atmospheric instability is comparable. (This was taken account of in case studies I and V.) At this stage only qualitative account has been taken of the topography of the GDF area. To perform this lifting calculation, a wind-profile above the mountain was inferred from the measured speed at the summit. We assumed a simple logarithmic vertical wind-profile, and a roughness length Z_0, of 0·02 m (calculated from simultaneous measurements made at 2 m and 10 m above the mountain top during one of the experiments). These calculations are probably subject to some error when conditions deviate significantly from neutral stability. However, it is found that the calculated wind profile merges with the observed free air profile some distance above the mountain top at about the expected altitude (Mason and Sykes 1978) or reaches a sharp discontinuity of the shear at an inversion level. Air parcels are lifted dry adiabatically to their condensation level and moist adiabatically thereafter, in ascending sections of 100 m above the ground and piled on top of each other at the mountain top. It was assumed that once the air had risen to its condensation level the water which became available on subsequent lifting remained in the air as cloud water. The vertical stretching (or shrinkage) was estimated from the ratio of the wind speeds of the air parcel above lowland terrain to those above the mountain top, and also from the pressure change experienced by the parcel as a result of its ascent by using continuity of mass. Since the windspeed at a given height above the hill top was higher than the windspeed at the same height above the lowland, the vertical displacement of the air decreased with increasing starting heights. The condensation level, mountain top liquid water content and temperature were calculated taking account of mixing between the parcels.

Values of the Richardson number

$$\mathrm{Ri} = \frac{g}{\theta}\left(\frac{d\theta}{dz}\right) \Big/ \left(\frac{du}{dz}\right)^2$$

where g = acceleration due to gravity
θ = potential temperature
z = height
u = wind speed

were estimated for the lower layers of the atmosphere as relevant (mostly in the turbulent boundary layer below 1 km), and above the summit, as a measure of the atmospheric stability. Values of the mass exchange coefficient K were estimated in each case so that the time constants for mixing of the various parameters during ascent could be roughly estimated. The expressions for stability-dependent exchange coefficients developed by Pruitt, Morgan and Lawrence (1973) were used, in which

$$K = k_0/\phi$$

where $\phi = 0{\cdot}855(1+34\mathrm{Ri})^{0{\cdot}4} \qquad 0 \leqslant \mathrm{Ri} \leqslant 0{\cdot}3$
$\phi = 0{\cdot}855(1-22\mathrm{Ri})^{-0{\cdot}4} \qquad -3{\cdot}5 < \mathrm{Ri} < 0$
k_0 is the adiabatic value of the exchange coefficient, given by

$$k_0 = kzu^*$$

where k is Von Kármáns constant (about 0·4), z is the height above ground and u^* is the frictional velocity. u^* was calculated for the unmodified wind profile using a standard geostrophic departure method for the boundary layer.

In these rough calculations the upper expression for ϕ was used for values of Ri up to +2. Taking account of the local trajectories of the air to the mountain summit these coefficients were used to make a rough estimate of the degree of mixing taking place.

The purpose of the model was to enable the long time-scale changes in the observed structure of the cloud to be interpreted in terms of synoptic scale variations in the atmospheric stratification and moisture content.

Throughout this study the effects of radiation on the droplet spectrum were considered to be negligible. The highest radiative cooling rate likely to be encountered anywhere in the clouds studied was estimated to be near $10^\circ\mathrm{C\,h^{-1}}$, which is equivalent to an updraught speed of around $20\,\mathrm{cm\,s^{-1}}$. The updraughts and hence adiabatic cooling rates were always much higher than these figures.

3. Specific case studies

Examination of the information obtained over the four days covered by these field experiments revealed that there were four periods, of extent ranging from one-half to several hours, during which good data were obtained in conditions where environmental air was being entrained into the cloud formed over the research station. Detailed meteorological analyses were performed for the four cases and these, together with the cloud-water characteristics, determined by means of the Keily probe, are described below.

Case I: 1630–1700h on 12 December 1977

A westerly airflow covered northern England. A weak ridge extending northwards from a belt of high pressure stretching through the Bay of Biscay was approaching from the west, followed by a weak warm front. During the run the axis of the ridge was just off the east coast of Ireland and the surface warm front was between 160 and 240 km to the west of Ireland.

During the 48 hours prior to arriving at GDF, the air had been involved in the circulation of a deep depression over the eastern Atlantic. Thus, although the air was advected to GDF from higher latitudes, it had spent a considerable time over the ocean and had not been sufficiently far north to pass over any land or sea ice. Accordingly, the temperatures were relatively high although the air was somewhat unstable in the lower layers (more so than at any time during the experiments conducted) due to the passage over progressively warmer sea as it was advected S.E. to GDF.

Cloud cover over the north of England was generally well broken during the run with between $\frac{1}{8}$ and $\frac{3}{8}$ cumulus cover with its base at around 540 m (this was dispersing due to slight radiative cooling at the surface) and $\frac{3}{8}$ to $\frac{6}{8}$ stratocumulus at about 1 km, also dispersing. Several stations had reported light rain showers during the afternoon.

It was decided that the temperature/dew-point soundings obtained at Long Kesh and Aughton at 1100h on 12 December were representative of the GDF area during the run. The ascent showed that the air was conditionally unstable from the surface to 1·2–1·5 km, above which it was more stable, and eventually drier, probably due to slight subsidence associated with the ridge. The wind profile for the GDF region was deduced from the data obtained from several sonde stations at 0500 and 1700h on 12 December together with geostrophic wind data obtained from the synoptic charts. The observed windspeed 2 m above ground on the summit was $10\,\mathrm{m\,s^{-1}}$. The analysis was then performed as described.

The lifting of the air to the summit produced a deep cloud, extending to at least 400 m above the mountain top. The lapse rate within the cloud was in excess of the moist adiabatic. The observed mountain-top temperature was 3·5°C (calculated 3°C) and was substantially in excess of the free air temperature at the same level (2°C). This was true throughout the depth of the cap cloud (except, perhaps, at the top where the environment becomes more stable).

Shortly after the run it was observed that the cloud base on the mountain side was at 400 m above m.s.l., and that apart from the cap cloud the sky was clear. It was considered unlikely that any droplets from the overlying stratocumulus cover, present early in this period, were mixed down to the observation point at any time during the run.

The drop-size distribution observed in the cap cloud at about 1630h 12 December is shown in Fig. 2. The period of integration was 15 min. The cloud base was at around 400 m (approximately 2·5 km upwind of the summit).

As discussed, the atmosphere was conditionally unstable (Ri for dry adiabatic motion slightly in excess of zero) and some cumulus had been present in the area. It is clear that in such circumstances, once cloud has been formed by forced ascent, air will be mixed into the cloud from outside its boundaries in the free air. Further, large-scale motions are to be expected with free convection contributing substantially to the speed of ascent up the mountain side. It may also be expected that the entrainment at the cloud edge, well away from the mountain, may be somewhat similar to that for a cumulus cloud. Evidence for entrainment is provided by the fluctuations observed in the liquid water content measure-

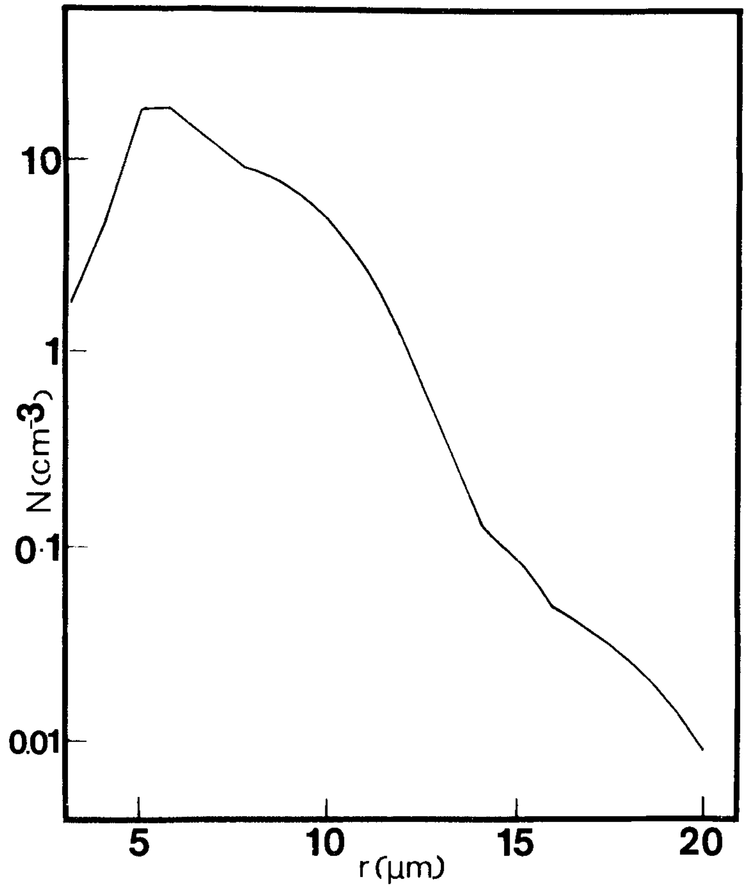

Figure 2. Measured size distributions at 1630h on 12 December 1977 (L = 0·22 g m^{-3}).

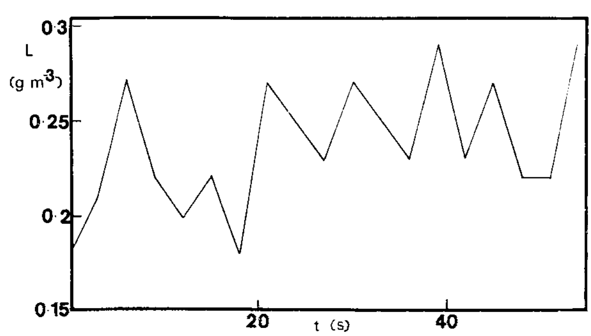

Figure 3. Example of liquid water content (L) record obtained around 1630h on 12 December 1977.

ments, L, during the experiment. An example of these records is presented in Fig. 3. We find an 18% r.m.s. deviation in L during the run, compared with 6% expected from random fluctuations due to the finite size of the drop count (which was integrated over successive three second time periods to obtain this data). The period of the fluctuations was typically around 10–15 s, corresponding to a scale length of 100–150 m, substantially in excess of the length of ground-induced turbulence at this height. The mean liquid water content during the experiment was $0{\cdot}22\,\mathrm{g\,m^{-3}}$. It is difficult to compare this with a figure to be expected in the absence of entrainment of outside air as it is dependent upon the relative importance of free convection and forced ascent up the mountain side, which affects the contribution of air from different levels. In the absence of entrainment, the estimated liquid water content is about twice the observed value.

A prediction of the crude model of inhomogeneous mixing presented in Paper I is that despite fluctuations in the liquid water content between adjacent volumes in the cloud, the drop-size distribution will be independent of the changing values of liquid water content through these regions. Field evidence in support of this prediction has been reported by Knollenberg (1976), Corbin *et al.* (1977) and Rodi (1978), but, in each case, no supporting meteorological evidence was provided, which would have permitted a more detailed examination. In the present case this prediction was examined by constructing and comparing a pair of droplet spectra based on larger and smaller values of droplet concentration N (which is closely correlated with L) measured during this period of fluctuating water content (Fig. 3). The spectrum for higher L was obtained by accepting three-second histograms for which $N/\bar{N} > 1+2/\bar{N}^{\frac{1}{2}}$ and for lower L, $N/\bar{N} < 1-2/\bar{N}^{\frac{1}{2}}$, where $\bar{N}$ is the mean droplet concentration over the entire period of measurement. It was found that a total of 40% of the histograms lay outside these ranges compared to 4·5% expected purely on the basis of sampling statistics. This indicates that significant fluctuations exist in drop concentrations within the cloud. The histograms in the 'largest' and 'smallest' categories were respectively summed and normalized to the same total sampling volume for comparison. The total

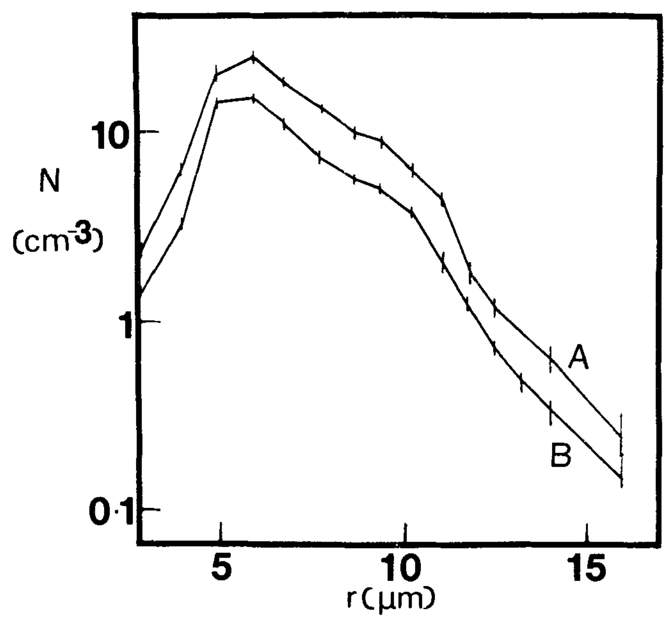

Figure 4. Size distributions for regions of high (A) and low (B) liquid water content, measured around 1630h on 12 December 1977.

period of integration for each spectrum was 300 s. The results of this analysis are shown in Fig. 4 with individual error bars due to sampling statistics plotted for each size category. There is no significant difference between the spectral shapes. These findings are in sharp contradiction to the classical description of entrainment, in which the smaller droplets would be preferentially evaporated, and the peak radius reduced. However, they are consistent with the predictions of the crude model of inhomogeneous mixing, presented in Paper I. In this connection it may be worth noting that the measured broad spectra presented in Figs 3 and 4 are distinctly non-classical, but are similar to the predicted ones displayed in Fig. 1 of Paper I, which agree well with those measured by Warner (1969) in cumulus clouds. Precise comparison between theory and measurement is not possible, because we have no information, in our studies, on the CCN activity spectrum.

Case II. 1138–1824h on 14 December 1977

During this period an anticyclone, of central pressure 1032 mb, was centred over north France and remained stationary. A weak cold front lying from north Scotland to Northern Ireland at 1200h moved slowly south-east to reach central Scotland by 1400h and its southern limit, Newcastle, by 1700h. During this period it was returning north, as a warm front, in the west due to a shallow wave depression which ran north-east then east along the front off the north-west coast of the British Isles.

The air arriving at GDF at 1800h on 14 December was situated to the west of Iceland at 1200h on 10 December. This was advected south-east to the Bay of Biscay by 1800h on 11 December where it became involved in the circulation of an anticyclone which was located off North Africa at 1200h on 11 December but which, by 1200h on 12 December, was extending a ridge into the Bay of Biscay. Over the next 24 hours the original centre declined in favour of a new centre over northern France which advected the air north then north-east to GDF. By contrast, the air arriving at GDF at 1200h on 14 December had remained on the west flank of the anticyclone, and at 1200h on 13 December was at 43°N 17°W; 12 hours earlier it had been to the west of North Africa. The gradual slight drying of the air and the presence of more broken cloud over north England during the run may be partly attributed to the slow change between these two air masses. The data presented here was taken mostly under the influence of the early moist air over north-west England. At 1200h on 14 December the cloud cover consisted of $\frac{1}{8}$ stratus with base 600 to 900 m and broken layers of stratocumulus at about 900 m. There was about $\frac{4}{8}$ total cover. Also, there was a general layer of altocumulus and altostratus at 3 km, probably associated with the cold front. The surface wind was between 220° and 250° and the speed 5–7 m s^{-1}. At 1800h the situation was little changed. Cloud cover was well broken with $\frac{1}{8}$ stratus between 200 m and 400 m and some stratocumulus at around 1 km generally around $\frac{1}{8}$ to $\frac{3}{8}$ in total, less than at 1200h. Some altostratus and altocumulus was again evident in patches, but was less extensive than at 1200h (as the cold front continued to weaken). The surface wind was generally 200° to 240°, speed 4–5 m s^{-1}. Some stations in south-west Scotland and extreme north-west England reported intermittent slight rain during the afternoon. The rain reported at GDF around 1200h was probably falling from the altocumulus and of frontal origin.

Examination of the radiosonde ascent data taken over the British Isles shows a subsidence inversion associated with the anticyclone over northern France. This lowered considerably and, intensified during the period up to 1100h on 14 December and more slowly thereafter. It was estimated from the ascents that the inversion height was around 900 m at GDF during the experiment and that below the inversion the Long Kesh ascent for 1100h on 14 December, together with the wind data obtained at 1400h were representative of conditions at GDF.

This most representative tephigram has a shallow well-mixed layer with a dry adiabatic lapse rate, extending 300 m above the surface. Above this is a layer with a near moist adiabatic lapse rate (but still unsaturated) in which the bulk Richardson number was estimated at 0·5. This was capped by a fairly weak inversion. It was considered likely that in advection to Great Dun Fell and the early stages of lifting the profile would be modified, the two lowest layers becoming well mixed. Evidence for this, together with the persistence of a well-mixed state throughout the lifting is provided by the good agreement between the observed cloud base (500 m) on the mountain side at 1400h on 14 December, and the calculated cloud base (approximately 500 m) based on this assumption.

The liquid water content and temperature calculated for the mountain top under these conditions were 0·4 g m^{-3} and 3·7°C respectively. The observed mean liquid water content during the run rose from 0·04 g m^{-3} at 1300h to 0·09 g m^{-3} at 1421h, while the temperature rose from 4·5°C at 1200h to 5°C by 1500h. The drop-size distribution observed at 1300h is shown in Fig. 5. (The subsequent increase in liquid water content is accompanied by an increase in the drop count but there is no significant change in the shape of the drop-size distribution.)

The differences between the observed and calculated liquid water contents in the cloud strongly suggest that dry air is being mixed in from outside the cloud. Further evidence for

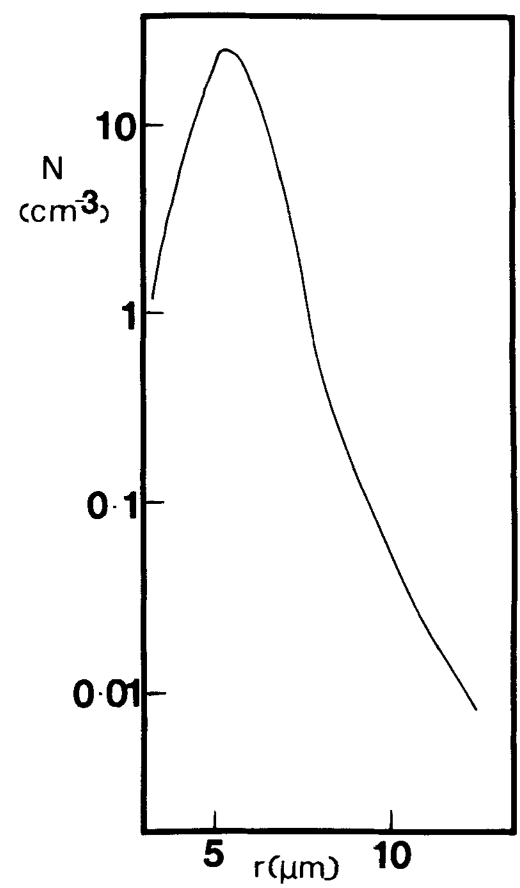

Figure 5. Size distribution measured at 1300h on 14 December 1977. L = 0·09 g m^{-3}.

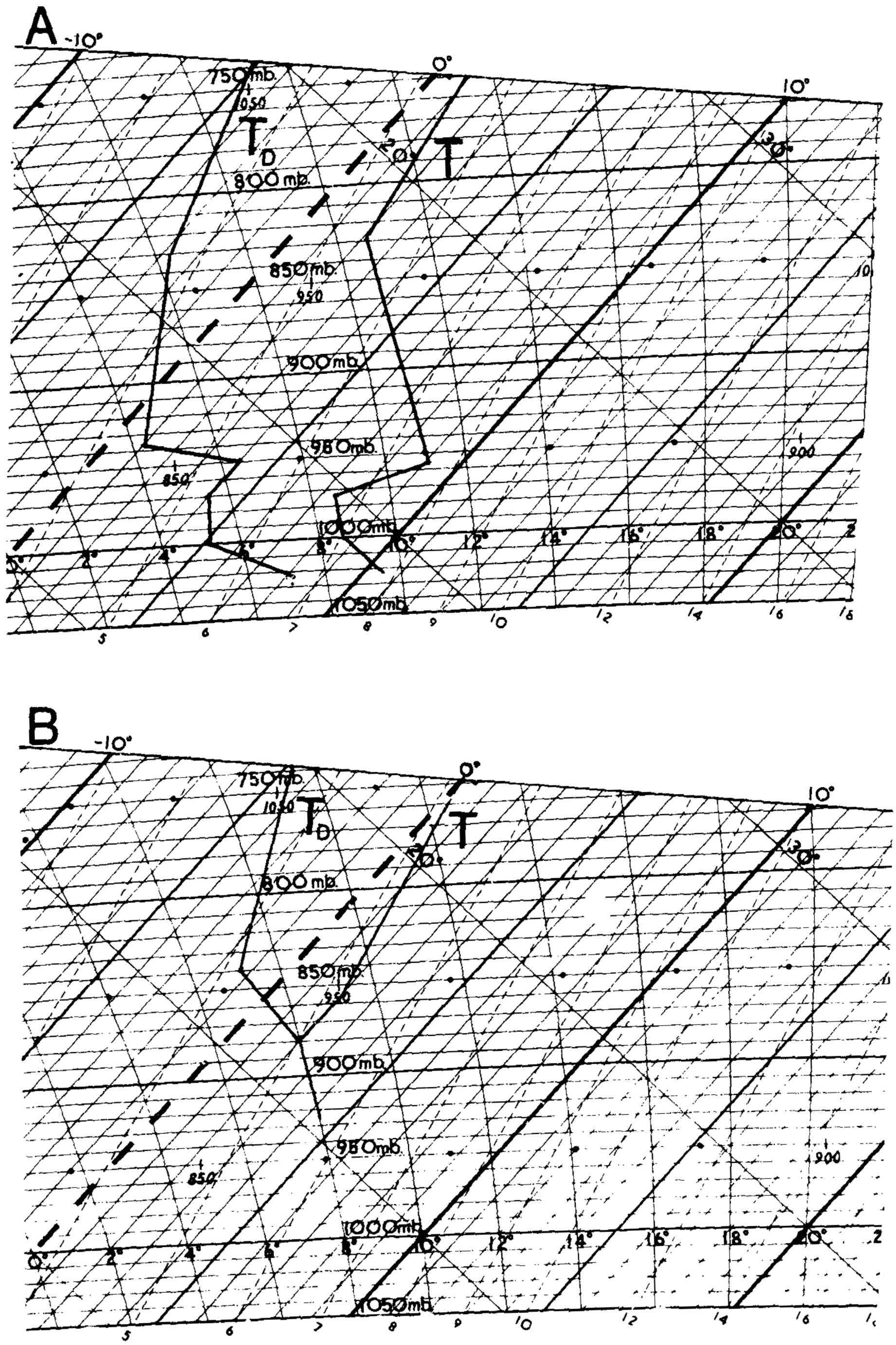

Figure 6. Tephigram data estimated for the Great Dun Fell area for the early afternoon of 14 December 1977. A, before lifting to the mountain summit; B, after lifting.

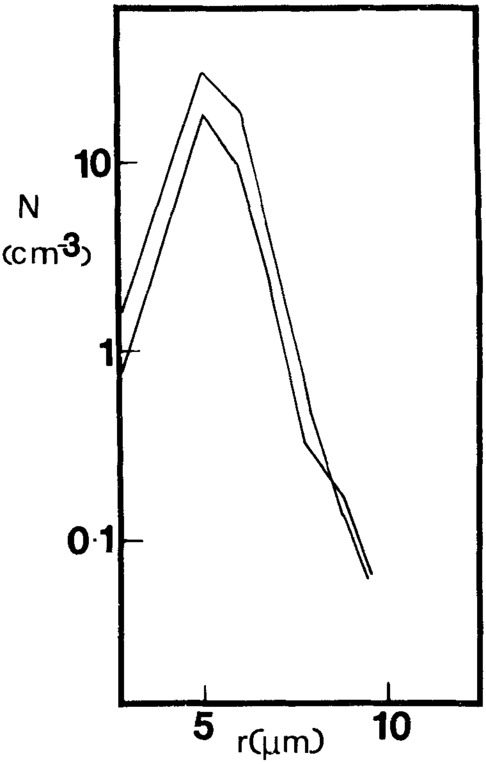

Figure 7. Size distribution for regions of high and low liquid water content, measured around 1300h on 14 December 1977.

this exists in the comparison displayed in Fig. 6 between the initial temperature profile and calculated values above the mountain top. This indicates that the inversion zone has been substantially destabilized during the lifting process. It was not possible on this simple model to calculate accurate values of the Richardson number in this layer after lifting. It is clear, however, that Ri was reduced sufficiently for some mixing to occur with the well-mixed cap cloud below. If, for example, the last 50 m of the air was mixed into the initially 300 m deep cap cloud then we estimate that the mountain top temperature would be increased to about 4°C and the liquid water content decreased to about 0·15 g m^{-3}. Further entrainment could increase the temperature and decrease the liquid water content. Hence entrainment was capable of generating values close to those observed.

Further evidence for the mixing-in of dry air comes from the observation of fractional r.m.s. fluctuations in the liquid water content of about 18 %, compared with 5 % expected purely on the basis of sampling statistics. Droplet spectra for regions of high and low water contents were constructed in the manner described earlier (Case I) and are displayed in Fig. 7. The two spectra are seen to have more or less identical shapes for droplet radii r up to about 8 μm, but for higher values a convergence is apparent, which was found to be statistically significant. It was argued in paper I that the extent to which spectral modification produced by entrainment approaches the classical or extreme inhomogeneous descriptions depends upon the ratio τ_T/τ_r of the time constants for turbulent mixing and droplet evaporation (in the former case $\tau_T/\tau_r \to 0$, and in the latter $\tau_T/\tau_r \to \infty$). τ_r is, of course, strongly dependent upon droplet size and the convergence displayed in Fig. 7 is explicable in terms of the dimensional arguments presented in Paper I if $\tau_T/\tau_r \to 1$ for $r \sim 8\,\mu$m. This condition can be expressed as:

$$\frac{2S(X^2/\varepsilon)^{\frac{1}{3}}}{(r+\alpha)^2-\alpha^2} \to 1$$

where S is the undersaturation in the entrained air, X is the eddy size, ε the rate of kinetic energy dissipation via turbulent mixing and α the length associated with the condensation coefficient. Taking $S = 4\%$, $\varepsilon = 100\,\text{cm}^2\,\text{s}^{-3}$, $\alpha = 5\,\mu$m and $r = 8\,\mu$m this condition is met when $X \sim 8$ m; if $r = 10\,\mu$m then $X \sim 13$ m. These crude calculations appear consistent with the observation that the scales of the fluctuations in L within the cloud lay within the range 10–20 m.

The observation that the droplet spectra at the research station were narrow (Figs. 5 and 7) despite the fact that the water content was substantially sub-adiabatic suggests that the entrainment may have occurred predominantly near the summit. Support for this idea is provided by Fig. 6 which shows that the inversion layer was destabilised as the air was lifted to the mountain top. This process occurred gradually as the air was lifted up the mountain side but mostly after the cloud had formed in the well-mixed layer below the inversion at about 500 m above m.s.l. Initially the amount of entrainment from the inversion layer would be small and the size of the entrained eddies also small. As this layer was progressively destabilized, however, both the entrainment rate and mean eddy size would rise. It is clear, however, that at all times the scale of the entrained eddies will be much smaller than in Case I. This argument is consistent with the observation of convergence in Case II but not in Case I.

The finding that although the liquid water content was approximately doubled later in the period of investigation while the spectral shape and peak size remained constant is (in a similar manner to Case I) consistent with the extreme inhomogeneous model. It should be noted that there is no conflict between this conclusion and the calculation on convergence just presented – the condition $\tau_T/\tau_r \to 1$ will not be achieved in the narrow spectra found at GDF if the values of X or S are only marginally greater than those assumed in this specific calculation. In such a situation detectable convergence will not occur and, to a first approximation, the observations will conform with the predictions of the crude model of inhomogeneous mixing.

Case III and Case IV: covering the period 1930h on 15 December to 1230h on 16 December 1977

During the period of the runs an anticyclone situated over the North Sea at 1300h on 15 December moved slowly south-east into North Germany and intensified.

Early in this period the air trajectories to Great Dun Fell were slightly to the east of the Pennines where lowland stations were reporting between $\frac{1}{8}$ and $\frac{3}{8}$ cover of stratocumulus with base at 600 m. After 2200h the wind had veered sufficiently for the air trajectories to be somewhat to the west of the Pennines, where they remained. At 1800h stations in this area reported a stratocumulus cover of similar extent with a base at 450 m.

During the period up to midnight the cloud cover over north-west England became more complete; most stations reporting $\frac{8}{8}$ stratocumulus cover at between 300 m and 400 m. A nearly total cover of stratocumulus ($\frac{7}{8}$ to $\frac{8}{8}$) then persisted at most stations for the remainder of the period. By 0600h on 16 December cloud base was generally somewhat higher (around 600 m) before falling again to around 400–500 m by 1200h. The extent of cloud cover and the cloud-type were confirmed by examination of satellite photographs.

The Aughton temperature ascents were considered to be representative of Great Dun Fell throughout the period, and the ascents for 1100h on 15 December, 2300h on 15 December and 1100h on 16 December were studied in some detail.

The period of the runs was characterized by a gradual cooling of the air between the 1000 mb and 950 mb pressure levels, together with an increase in relative humidity and a decrease in stability from the surface up to 950 mb (between 700 m and 800 m above m.s.l.). During the same period the inversion associated with the developing anticyclone intensified substantially and lowered from 980 m at 1100h on 15 December to 900 m by 2300h on 15 December, and then to about 600 m by 1100h on 16 December over lowland regions in the vicinity of GDF.

Air trajectories were constructed at various times throughout the period. The Aughton wind-profiles were used to perform the lifting calculations described previously.

Case III: 1930h to 2245h on 15 December

At the start of the run a thin and tenuous cap cloud was observed through which the moon could be clearly seen. This cloud was observed to be highly inhomogeneous with some completely clear patches, lasting for several seconds. By 2100h the cap cloud was thicker and much more homogeneous, and the moon was totally obscured.

Figure 1 shows the drop-size distribution and mean liquid water content obtained for the early, highly inhomogeneous cap cloud (this showed variations of liquid water content of a factor of 50 over a period of 60 s) compared with a thicker much more homogeneous cap cloud in which the r.m.s. fluctuations of water content were about 6%. It is apparent that relatively more large droplets are present in the earlier spectrum.

The lifting, deduced from the tephigram for Aughton at 1100h on 15 December, resulted in a shallow layer, at the mountain top, that was less than 100 m deep and slightly supersaturated. Above this was a layer of undersaturated highly stable air ($Ri \approx 8$) up to the main inversion layer. At this time, however, the summit of the cap cloud was sufficiently close to the ground that the windshear was large enough for some entrainment of dry air to occur. A gradual transition to the air on the 2300h tephigram resulted in a cap cloud of progressively greater vertical depth – finally reaching the main inversion layer – and a gradual decrease in stability. This was characterized by a decrease in the bulk Richardson number in the sub-inversion layer before lifting from 1·3 to 0·9; and in the cap cloud a gradual decrease to a final value around 0·3. It is to be expected that as the cap cloud deepened, and thus its top moved away from the surface, a substantial decrease in the amount of entrainment of dry air would occur. Thus from 2100h to around 2245h the spectrum was narrower than at first, and remained almost completely unchanged in shape as the cloud gradually thickened. The observed mean liquid water content at 1930h ($0{\cdot}02\,\mathrm{g\,m^{-3}}$) was substantially below the calculated value ($0{\cdot}1\,\mathrm{g\,m^{-3}}$), which took no account of mixing. By 2157h the observed value had risen to $0{\cdot}08\,\mathrm{g\,m^{-3}}$, and the cloud had become highly homogeneous. At this stage the fractional r.m.s. deviations in liquid water content were 6%, compared with an expected value from statistical fluctuations of 5%.

It appears, therefore, that the entrainment of undersaturated air when the cloud was thin was responsible for the broadening of the spectra revealed in Fig. 1. We cannot explain the observed spectral shape changes on the 'classical' picture of mixing as a homogeneous process, but comparison with Fig. 1 of Paper I shows that the spectrum measured at 1930h closely resembles that predicted on the inhomogeneous model for a similar liquid water content. The drop concentration ($140\,\mathrm{cm^{-3}}$) suggests, as would be expected from the trajectories, a rather more continental aerosol distribution than on 12 December and so closer to that used to calculate the curves in Fig. 1 of Paper I. There is no evidence, either from air trajectories or drop counts, that any significant change in the CCN activity spectrum occurred between 1930h and 2200h.

Case IV: 2245h on 15 December to 1230h on 16 December

Values of all the microphysical parameters measured remained almost unchanged until around 2245h. At this time intermittent slight rain was observed at GDF which, in the period up to midnight, gradually became somewhat heavier and more continuous. No precipitation was reported from nearby lowland stations. This could not have originated from higher cloud as the air aloft was dry and largely cloud-free (this is confirmed by the satellite photographs). At the same time the liquid water content began to rise and the drop-size distribution became progressively and appreciably broader, as shown in Fig. 8. The last two effects continued into the morning of 16 December, by which time the rain had stopped.

An explanation for the development of this rainfall can be offered, which depends upon the observed development of an extensive stratocumulus cover, with base below the mountain summit. Taking the known altitude of the base of the subsidence inversion to coincide with the summit of the stratocumulus it may be estimated that a droplet of ini-

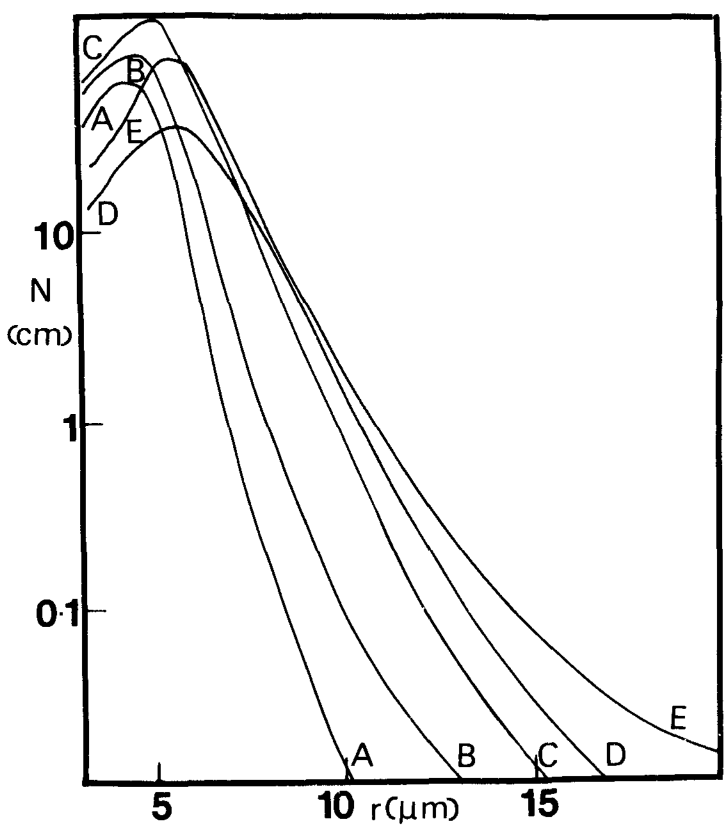

Figure 8. Size distributions measured at various times on 15 (A and B) and 16 (C, D and E) December 1977. A, 2245h, $L = 0{\cdot}07\,\mathrm{g\,m^{-3}}$; B, 2340h, $L = 0{\cdot}12\,\mathrm{g\,m^{-3}}$; C, 0020h, $L = 0{\cdot}14\,\mathrm{g\,m^{-3}}$; D, 1030h, $L = 0{\cdot}37\,\mathrm{g\,m^{-3}}$; E, 1100h, $L = 0{\cdot}31\,\mathrm{g\,m^{-3}}$.

tially around 20 μm radius may grow by coalescence as it falls through the stratocumulus and then grow further to a size of 1 mm as it falls through a few hundred metres of cap cloud. This growth would be facilitated by the lifting of the stratocumulus close to the mountain. The smaller droplets from the stratocumulus above, away from the mountain, would evaporate before reaching the ground.

The cessation of precipitation by 1000h on 16 December may be accounted for in terms of the lowering of the subsidence inversion which occurred overnight, and the consequent reduction in cloud depth above the mountain.

It has already been mentioned that the air below the subsidence inversion became progressively more unstable during the experiments. The tephigram for Aughton at 1100h has a bulk Ri number of 0·2 in the boundary layer below the inversion (0–600 m). The continued broadening of the drop-size distribution after 2245h, observed in the cap cloud, is probably to be accounted for by the progressively greater degree of entrainment of the long-lived stratocumulus cloud as the eddy exchange coefficients around the mountain increase as the Ri number decreases. The large droplets would be produced in the stratocumulus away from the mountain due to processes such as radiative cooling, coalescence and mixing. Further evidence supporting this explanation for the observed spectral broadening is provided by the measurements of liquid water content. These increase, as the spectra broaden beyond 2245h, reaching a maximum of 0·37 g m^{-3} at 1034h on 16 December. During this period the calculated adiabatic values rise from 0·1 to only 0·2 g m^{-3}. This calculation, based on the tephigrams, takes no account of the presence of any pre-existing cloud. We conclude, therefore, that those super-adiabatic water contents result from the downward mixing, to the point of observation, of droplets from stratocumulus. The spectral shapes are fundamentally different from those produced by the mixing-in of dry air with the largest changes in spectral width occurring at the largest drop size. During the period of spectral broadening, fluctuations developed in the liquid water content, which increased with time, but the maximum fractional r.m.s. value reached was around 10%, toward the end of the run. This can be explained in terms of the increased turbulence due to decreased stability. These fluctuations are considerably less than those experienced in other runs, when dry air was being entrained into the clouds.

Throughout the experiment the predicted and observed temperatures on the summit were in good agreement showing that the observed rise in liquid water content above the adiabatic values could not be accounted for by errors in the input temperature soundings or in the lifting model. Also, base altitude away from the mountain agreed well with mixing condensation levels deduced from the tephigrams.

Case V: 1100h to 1300h on 28 February 1979

A cold front had crossed northern England during the early hours of 28 February. For the period of the measurements it was lying just off the East Coast, moving east at about 20 kt. The precipitation area was well clear of the mountain throughout the run although some broken altocumulus was present at 2–3 km above m.s.l. Dry air, produced by subsidence behind the front, severely restricted the vertical growth of cumulus cloud in the GDF area. There was, however, a gradual cooling of the air at and above the mountain top level due to cold advection as the front moved away. It was estimated that the free air temperature at the mountain top level fell from about −3·5°C to −4°C between 1100h and 1300h. The air at this level was conditionally unstable throughout. However, due to solar heating, the average temperature rose from 3·5°C to 4·5°C at lowland stations over northern England and the low cloud cover increased from $\frac{2}{8}$ to $\frac{3}{8}$ of small cumulus with bases rising from 500 m above m.s.l. at 1100h almost linearly to 750 m above m.s.l. at 1400h. Taking account of the Aughton ascent for 1200h (which was broadly representative of the Great Dun Fell area on this occasion) the combination of cooling aloft and surface heating resulted in an estimated increase in the lapse rate close to the mountain top level from $8°C\,km^{-1}$ at 1100h to $9{\cdot}5\,°C\,km^{-1}$ at 1300h. The geostrophic wind was light ($5\text{–}7\,m\,s^{-1}$) during this time.

Using a combination of direct observations and inferences drawn from the hourly meteorological data and the rates of change of liquid water content observed at the mountain top it was estimated that the cloud base on the mountain side remained roughly constant at about 425 m until about 1130h or 1145h and then lifted, approximately linearly with time, to reach the mountain summit (850 m) at 1300h. The uncertainties in this information do not have a crucial effect on the arguments that follow.

The maxima and minima spectra and average liquid water contents for the periods 1132–1139h, 1149–1158h and 1230–1240h are shown in Figs. 9, 10, 11. The cloud tended to be patchy, especially early in the period, with regions of thick cloud lasting for periods of several minutes and holes for periods of around one minute. These holes became progressively less frequent during the period 1058h to 1140h and were totally absent later. The spectra discussed are all for the thick cloud regions. These observations, when considered with the thermal stratification and the light winds suggest that the cloud near the mountain summit was largely produced by free convection triggered by weak orographic lift.

Comparison of the maximum and minimum count curves for each graph in turn shows that early in the run the two curves converge at both large and small drop radii. The liquid water content is around 0·5 to 0·6 of the adiabatic value (calculated from cloud base). These are consistent with the picture of the cloud having been modified substantially by the mixing-in of dry air and, as in Case II, the effect of mixing on the cloud properties is not completely explicable in terms of either the classical or extreme inhomogeneous description. Later, as shown in Fig. 10, the maximum and minimum count curves became much more similar in shape, as was also found in Case I. This occurred as the environmental temperature lapse

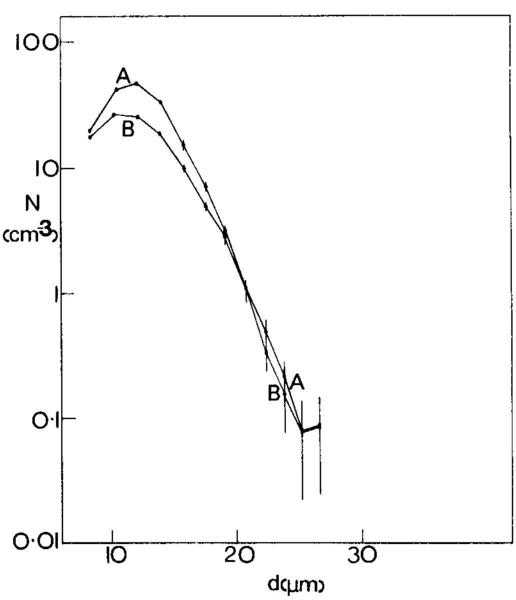

Figure 9. Size distributions for regions of high (A) and low (B) liquid water content, measured over the period 1132–1140h on 28 February 1979. $L = 0{\cdot}18\,g\,m^{-3}$.

rate increased. In addition, the fraction of the adiabatic water content in the cloud fell progressively (both before and after the cloud base started to rise).

Further evidence for the increased scale of entrained eddies is supplied by the increasing separation of the curves A and B (Figs. 9–11) with time, suggesting that the droplet evaporation is becoming much more inhomogeneous in space. This effect may possibly be partly

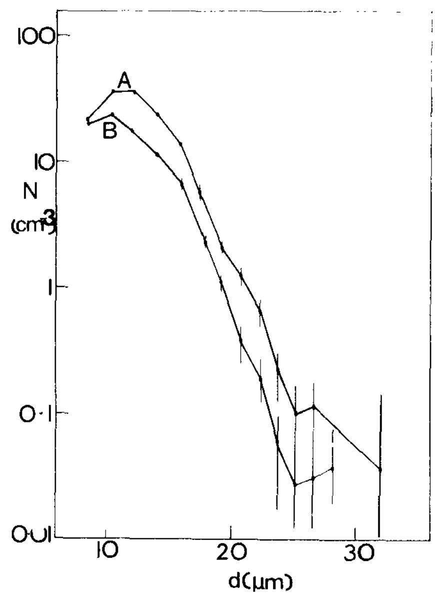

Figure 10. Size distributions for regions of high (A) and low (B) liquid water content, measured over the period 1149–1158h on 28 February 1979. $L = 0{\cdot}12\,\mathrm{g\,m^{-3}}$.

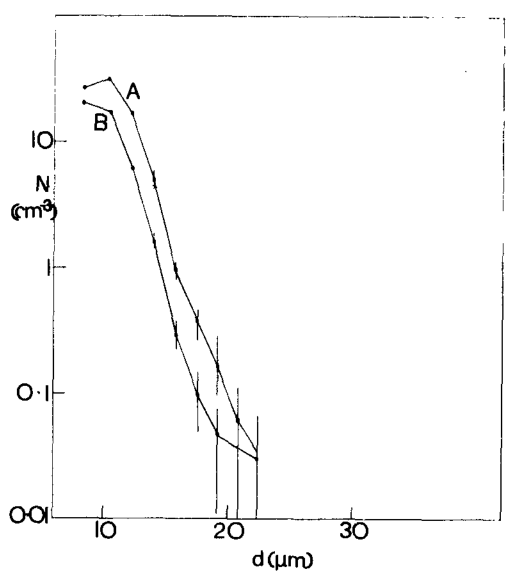

Figure 11. Size distributions for regions of high (A) and low (B) liquid water content, measured over the period 1230–1240h on 28 February 1979. $L = 0{\cdot}05\,\mathrm{g\,m^{-3}}$.

attributable to a decrease in the relative humidity at low levels as the ground is warmed. It is unlikely, however, that this is pre-eminent, as the relative humidity of the environmental air remains close to 80% at 850 m above m.s.l.

Figure 12 shows that the dispersion of the drop-size distribution initially remains approximately unchanged as the liquid water content falls, consistent with a progression towards a spectral shape predicted by the model described in Paper I and later, as the cloud base rises rapidly, the dispersion falls with the liquid water content to around 0·15 near cloud base. This behaviour is similar to that observed by Warner in cumulus clouds and is again consistent with the model described in Paper I.

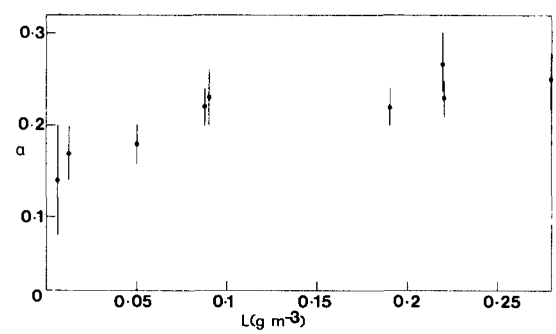

Figure 12. Measured relationship between dispersion α and liquid water content L for the entire period covered by Case V.

The enhancement of growth rates of large drops

A prediction made by the model described in Paper I is that large drops in the spectrum will experience an enhanced growth rate compared with that expected classically for a given spectral shape and drop concentration. In order to test this idea comparisons were made between the growth of droplets of intermediate and large size in the period covered by Figs. 10 and 11. Within this period the peak diameter (about 10 μm) is seen to increase by about 1·5 μm. On the classical picture of droplet growth (assuming a characteristic length of 10 μm associated with the condensation coefficient) droplets of original diameter 20 μm would be expected to increase by about 1 μm over this period. However, the observed increase in diameter of these larger droplets is seen to be about 5 μm. Thus we conclude that the growth rate of the largest droplets is enhanced by a factor of about five, which fits reasonably well with values predicted by the model presented in Paper I. The estimated uncertainty in this enhancement factor is 50%. Further support for this crude model is provided by the finding that earlier in the experimental period, when the mixing was more homogeneous, the enhancement factor was much lower.

4. Discussion

The field experiments described in this paper show that the cap clouds enveloping the

UMIST research station on the summit of Great Dun Fell may be profoundly affected by the entrainment of air from outside the cloud.

When the atmosphere is stably stratified cap clouds can develop which possess negligible structure in their water properties and within which the water contents approach the adiabatic value, entrainment being insignificant. The droplet spectra are always narrow, in these circumstances. However, in conditionally unstable air isolated cap clouds can form within which the temperature is substantially warmer than the environment at the same level. Thus, these tend to be convective, and rather like cumulus. In these situations substantial entrainment can occur yielding appreciably sub-adiabatic water contents and (as reported herein) large-scale inhomogeneities in the water properties and broad droplet spectra.

In the three clear-cut cases considered (I, II and III) where significant entrainment of droplet-free air occurred the spectral changes were inexplicable in terms of the classical (homogeneous) description of mixing but, to a first order, were consistent with the extreme inhomogeneous model described in Paper I. In one situation (Case II) a more detailed examination provided support for the idea, advanced in Paper I, that the influence of turbulent entrainment on cloud droplet evolution is determined, in part at least, by the relative values of the time constants for droplet evaporation and turbulent mixing. Additional evidence supporting this contention is presented in Case V, where a close correlation was found between the degree of inhomogeneity in the water properties of the cloud and the extent to which the microphysical properties are best described by the extreme inhomogeneous mixing model. Further, a significant enhancement of the growth rate of the largest droplets in the condensation spectrum was found, confirming a prediction made by the model described in Paper 1.

Clear evidence was found (Case IV) for the significant modification of the cap cloud by the entrainment of an overlying pre-existing stratocumulus deck and the growth of raindrops due to the seeding of the cap cloud by larger droplets from the stratocumulus. Although the meteorological context is very different it is interesting to note that the mechanism of spectral broadening by the entrainment into a growing cloud of environmental air containing droplets from, another cloud was advanced by Mason and Jonas (1974) in order to explain droplet evolution in cumulus.

Acknowledgments

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