Penetrative Downdraughts in Cumuli

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Abstract

Observations of cumuli have shown that the adiabatic model is quite inadequate, and that dry air must mix with the condensing upcurrent. Theories of the interaction between cumuli and their environment are discussed. The hypothesis that dry air enters these clouds chiefly from above is examined quantitatively. It is found that, as a result of evaporative cooling, such air could penetrate several kilometres into a growing cloud. This hypothesis accounts for the fine structure which is observed, and provides a natural and simple explanation for the fact that the lapse rate in cumuli is steeper than the wet adiabatic, and indeed approximates closely to that of the environment. Unlike alternative theories, it provides a means whereby the liquid water content may be automatically self-limiting, as the observations seem to require, for the motions which introduce dry air deep into the cloud depend for their energy supply on the presence of liquid water, and have velocities which are roughly proportional to its concentration.

1. Introduction

A wealth of evidence has been accumulated which shows that the properties of cumuli cannot be explained by the adiabatic expansion of air which was saturated at cloud base level. Measurements of temperature in cloud described by BUNKER et al. (1949) and by BARRETT and RIEHL (1948) indicate that the average lapse rate in cumuli is steeper than the wet adiabatic and approximates to that of the environment. Additional evidence is afforded by measurements of liquid water content (w g m^{-3}). After reviewing the observations, WARNER and SQUIRES (1958) concluded that although the adiabatic liquid water content (w_a) some times occurs over a distance of the order of a metre, the average value of w over some tens of metres ($\overline{w}$) seldom if ever equals w_a. The ratio $\overline{w}/w_a$ decreases rapidly with height and at 2 km above cloud base falls below 0.2, on the average. There is little evidence to suggest that $\overline{w}$ reaches much greater values in large clouds than in moderate ones; in all clouds, it seems rare for $\overline{w}$ to exceed 2 g m^{-3}, and it is usually less than 1 g m^{-3}. As regards distribution in the horizontal, there is a great wealth of fine structure, which becomes very striking when w is measured over a distance of about 1 metre. There is no clear systematic variation from the cloud edges towards the middle. Contrary to the impression given by ZAITSEV (1950), the low values of w appear to be distributed rather randomly, with no marked preference for the peripheral regions; the edge of a cloud is often very sharp.

These findings clearly require for their explanation a model of cumulus development which includes as a necessary concomitant of the cloud's growth a large scale mixing of dry air with the upcurrent.

2. Theories of interaction between the cloud and its environment

Since cumulus bases are typically flat, it seems impossible for dry air to enter the cloud from below, for in the present context, "dry" air means air which, if displaced adiabatically to the level of the cloud base, would arrive there unsaturated. The variations in cloud base height, which rarely exceed one or two hundred metres, are much too small to explain the observed deficiency of liquid water, $(w_a - w)$. Clearly, however, all air which originates from above the cloud base level will be "dry" in the sense defined. Thus there is room for three possibilities:

(a) The dry air may not have moved far from its original position, having mixed in situ with the saturated, cloudy air rising from below.

(b) It may have come in from around the sides of the cloud, mixing horizontally with the updraught.

(c) It may have originated from the region just above the growing cloud top, entering the cloud as downdraughts and mixing vertically with the updraught.

Model (a) is closely related to the bubble theory of convection advanced by SCORER and LUDLAM (1953), according to which the growth of a cumulus is the result of the arrival at condensation level of a succession of buoyant bubbles, each of which is steadily eroded as it rises. If other bubbles follow the same path through the air, they will penetrate further because the ambient air has been moistened by their predecessors; by this process a cumulus builds up in steps.

Model (b) was postulated by STOMMEL (1947) to explain the observed lapse rates in cumuli, and has been used as a basis for discussion by MALKUS (1949, 1952). The latter author has had considerable success in interpreting observations by means of a postulated entrainment. However, in these cases, there has been no independent evidence relating to the amount of entrainment, and it seems possible that some other mechanism which introduced dry air into the cloud in the required amounts could explain the observations equally well. Indeed, the only direct evidence of entrainment is provided by measurements of wind fields around large cumuli and thunderstorms made during the Thunderstorm Project. BYERS and HULL (1949) have described how the proportional rate of change in area of triangles formed by three balloons were used to compute horizontal divergence. The typical rate of entrainment into these large clouds was found to be 100 % in 4 km. By contrast, STOMMEL (loc. cit.) computed that, in order to explain the observed lapse rates in small trade wind cumuli, it was necessary to postulate an entrainment rate of 100 % in a height interval of 300 to 1,000 m. Again, BEST (1951) computed the liquid water content in large cumuli assuming entrainment of air of 70 % relative humidity at a rate similar to that found by BYERS and HULL, and found values of w which far exceed those which have usually been observed in large cumuli; the corresponding values of w/w_a exceed 0.5 at all heights. It therefore seems that, if the entrainment rates measured by BYERS and HULL are typical of cumuli in general, this source of dry air by itself is inadequate to explain the observed properties of such clouds.

Model (c) has been outlined elsewhere (SQUIRES, 1958) in an attempt to explain some of the observed characteristics of cumuli. A parcel of dry air which enters the top of a growing cloud by turbulent diffusion will be cooled by the evaporation of liquid water, and may subside into the cloud. The existence of a mechanism of this kind is suggested by measurements of vertical velocity, such as those of JONES (1954a), which have frequently revealed an intimate mixture of up-and-down-draughts of the same order of magnitude. Again, the downward motions often seen around the outsides of cumuli in lapse time records are commonly attributed to evaporative cooling consequent on mixing between the cloud and its dry environment. If this explanation is correct, it seems likely that the similar mechanism of model (c) could lead to deep penetration of a cloud by dry air parcels. This model, like (a) and (b), explains qualitatively the observation that the effects of the admixture of dry air increase upwards through the cloud, as shown by the decrease of w/w_a. It gives a picture of cloud development not unlike that of model (a). In both cases, the final result is a heterogeneous mixture of air from below condensation level with air which originally occupied, roughly, the space now filled by cloud. The difference is that here, an active, penetrative role is assigned to the dry air. Thus a heterogeneous cloud with much fine structure could result even from a steady updraught.

Model (c) has certain advantages:

(i) The remarkable fact that lapse rates in cumuli do not differ much from that of the environment finds a simple explanation by means of the mechanism of penetrative downdraughts. If the cloud should become markedly warmer than the environment at the same level, an extremely unstable situation will exist at the top of the cloud; vertical interchange must take place between the cloud top and the clear air above it. Evidence of the existence of this kind of interchange is given by the observations of JAMES (1954) who measured the air temperature while flying some 150 m directly above small developing cumulus clouds, and found that there were regions about 100 m across in which the temperature was some 1.5° to 3° C colder than elsewhere; these regions were sharp-edged, and stood in strong contrast to the undisturbed temperature field normally found in the clear air above cloud base and well away from clouds. These cold regions may have resulted from the evaporation of cloud protuberances into the surrounding dry air.

Indeed, it seems likely that the protuberances at the top of growing cumuli may be taken as evidence of the existence of strong hydrostatic instability in the region of the cloud top; if so, there is little reason to doubt that downdraughts are entering the cloud mass with a vigour similar to that shown by the protuberances. But whereas cooling by evaporation tends to stabilize the cloud protuberances when they mix with the dry environment, it renders the dry air parcels more unstable as they mix with the surrounding cloud.

(ii) As mentioned in the Introduction, the measurements of liquid water content so far made indicate that, whatever the value of w_a, w usually averages less than 1 g m^{-3} over large volumes, and is unlikely to exceed 2 g m^{-3} except in small regions. This suggests that there is some form of self-limiting mechanism at work, that is, that the admixture of dry air is greater in wet clouds than in dry ones. In both the bubble model (a) and the entrainment model (b), the admixture of dry air is not directly affected by the concentration of liquid water; in model (c), it is the liquid water itself which provides the driving force for the motions which give rise to mixing. As will be seen later, the subsiding speed of dry air parcels in cloud is roughly proportional to w.

(iii) It is a common observation that turbulence as observed from an aircraft is much stronger in clouds than in the clear air between them; according to BUNKER et al. (1949) by at least one order of magnitude. Now it is clear that the turbulence associated with clouds derives its energy from density differences, so that the turbulent motions must in the first place be vertical; cloud turbulence is likely to be far from isotropic, vertical mixing predominating strongly over horizontal mixing.

(iv) As mentioned in the Introduction there is no clear evidence for a gradient of properties from the cloud centre out towards the edges, as must be expected under model (b). The horizontal increase of w on entering a cloud is often very rapid, and there is no counterpart here to the vertical gradient of w which WARNER (1955) found to occur regularly in the uppermost 300 – 500 m of cumuli. This contrast reinforces the argument of (iii) above, indicating that vertical mixing plays a dominant role.

The hypothesis is therefore advanced that the mechanism of model (c) constitutes an important aspect of the interaction between cumuli and their environment. It remains now to investigate this model in a quantitative manner, that is, to determine whether the air in the dry layers which the growing cloud penetrates is itself capable of penetrating the cloud mass in a manner which is unorganized on the scale of the cloud. This penetration may be envisaged as occurring by the agency of more or less continuous downdraughts, for which there is some evidence (JONES), 1954 (b) and WARNER, 1955), or in the form of individual parcels. The latter view has been chosen in the interests of theoretical simplicity.

A penetrating dry parcel will obviously be subject to attrition and may at times be torn asunder. Thus a large parcel may enter the cloud, penetrate some distance, and then disintegrate, giving rise to two or more smaller ones which thereafter follow a more or less independent course. As our purpose is merely to investigate whether the proposed mechanism is capable of introducing dry air into the interior of a cloud in a time interval comparable with the observed lifetime of cumuli, the subsiding parcels will be supposed to remain integral.

3. A physical model of the downdraught theory of interaction

The following postulates describe the physical model which is used for the purpose of computation:

(a) A parcel of dry air becomes immersed in the cloud top as a result of the erratic growth of cloud protuberances, or by turbulence reinforced by the increased hydrostatic instability which must tend to occur at the cloud top when the lapse rate of the environment is steeper than the wet adiabatic, as is usually the case.

(b) Mixing occurs between the parcel and its cloudy environment.

(c) Cloud droplets diffusing into the parcel by turbulence immediately evaporate and, as a result, the parcel is cooled. Cloud droplets immersed in unsaturated air, even at a relative humidity of 99 %, will evaporate to a diameter of a micron or so in some tens of seconds, unless they contain a giant condensation nucleus; these are relatively rare. Thus they will have lost practically all their mass within a time which is short compared with the periods considered.

(d) The resistance to the motion of the parcel through the cloud arises solely from eddy interchange of matter between the parcel and its environment. The disturbance of the pressure field in the environment consequent upon the motion of the parcel is neglected. This simplification is justified to some extent by the following consideration. No hypothesis is made as to the shape of the parcels, and they may be envisaged as quasi-spherical or, more plausibly, as being elongated in a vertical direction, so that they form short-lived downdraughts. In this case, the effect of the disturbance of the pressure field could well be small compared with that of eddy interchange between the parcel and the cloud.

(e) The parcel is envisaged as containing a constant mass of air. Following the method used by PRIESTLEY (1953) it will be assumed that, in unit time, a mass of air k is removed from each unit mass of air in the parcel and is lost to the surrounding cloud; this is replaced by an equal mass of cloudy air, which mixes with the parcel. The effect of the parcel on the cloud will be neglected in this calculation; obviously however the final effect of many such parcels will be to cool and dry the cloud, especially in its upper parts.

The mixing rate k mentioned in (e) depends on the effectiveness of eddy diffusion in transferring matter from the cloud to the parcel and vice versa. For the moment it suffices to remark that k varies inversely with parcel size; the relation between them will be discussed later.

In addition to these postulates, the following simplifications and assumptions are made:

(a) In computing the density difference between the parcel and its environment, temperatures, not virtual temperatures, are used. Since the parcel itself is close to saturation in the critical stages of its history, the neglect of the effect of humidity mixing ratio on the air density is not very serious. Moreover, this simplification results in an underestimate of the negative buoyancy of the parcel, and so reduces the computed downward speeds; it is therefore conservative.

(b) The computation begins at a point where the parcel is immersed in the cloud; it is at the same temperature as the cloud, and is at rest.

(c) The liquid water content of the cloud is taken as being constant throughout its depth.

(d) The cloud air is assumed to have uniform and constant upward velocity.

(e) The lapse rate in the cloud is taken as constant.

(f) The mixing ratio of water vapour in the cloud air is found assuming the air to be saturated and at a temperature determined by (e) above.

All these postulates seem reasonable for the purpose in hand except perhaps (c). This is quite contrary to the adiabatic model, which would indicate a steady decrease downwards to zero at the cloud base; however, it represents the observations at least as well as any other simple assumption.

WARNER (1955) shows typical curves of the variation with height of the peak value of liquid water content (w_p) found during an aircraft traverse of a cloud. These curves indicate a rapid increase from zero at cloud base through the first 500 to 1000 m, and a rather more rapid decrease at cloud top. Between these two regions, the height variation of w_p is different in different clouds, and no systematic average trend is apparent. The behaviour with height of the mean value of w, $(\overline{w})$, as distinct from the peak value (w_p), depends on the shape of the w-traces obtained during the traverses. WARNER (private communication) states that in about half the clouds measured, $\overline{w}$ behaved similarly to w_p; in the remainder the height variation of $\overline{w}$ was even more erratic than that of w_p. Measurements by DAY and MURGATROYD (1953), WEICKMANN and AUFM KAMPE (1953) and SQUIRES (1958) give indications which are in agreement with these conclusions.

As so many variables are involved in the specification of the problem, and the solution of the equations is laborious, the calculations have been carried out for only one set of conditions, which have been chosen to represent a fairly well developed trade wind cumulus.

4. Basic equations for an unsaturated parcel

On the basis of the postulates described in the last section, and neglecting some terms which are small because the density of the parcel is little different from that of the cloud, the equation of motion of the parcel is:

$$\dot{v} = \left(\frac{\Delta}{\overline{\Theta}_c} - w_c\right) g - k\ (v + V) \qquad (1)$$

where v is the downward velocity of the parcel, V the speed of the cloud upcurrent, Δ the amount in degrees C by which the parcel is colder than the cloud at the same level, $\overline{\Theta}_c$ the mean absolute temperature of the cloud, and w_c the liquid water content of the cloud in grams per gram of air. Since the overall variation of the absolute temperature of the cloud is only a few percent, no serious error is involved in using the mean temperature $\overline{\Theta}_c$ instead of the temperature at the level concerned.

The liquid water content of the cloud (w_c) enters into this equation because the suspended liquid water increases the effective density of the cloud air.

The equation governing the temperature of the parcel is:

$$\dot{\Delta} = k\left(\frac{Lw_c}{c_p} - \Delta\right) - (\Gamma_d - \Gamma_c)\ v \qquad (2)$$

where L is the latent heat of evaporation of water, Γ_d the dry adiabatic lapse rate and Γ_c the lapse rate in the cloud.

The humidity mixing ratio of the parcel varies according to the equation;

$$\dot{x} = k(x_c(z) + w_c - x) \qquad (3)$$

where x is the mixing ratio of the parcel air and $x_c\ (z)$ that of the cloud air at the level z.

Differentiating (1) gives:

$$\ddot{v} = \frac{\dot{\Delta}}{\overline{\Theta}_c}\, g - k\dot{v}$$

and substituting for $\dot{\Delta}$ from (2) and for Δ from (1) gives a second order linear differential equation in v. Taking as boundary conditions $v=0, \Delta=0$ at $t=0$, and defining the origin of z, measured downwards, as the position of the parcel at $t=0$, the following expressions result:

$$\frac{\Delta}{w_c} = \overline{\Theta}_c\,\frac{bN}{mg}\, e^{-kt} \sin\, bt + \overline{\Theta}_c\left[1 + \frac{k}{g}\left(\frac{V}{w_c} + \frac{N}{m}\right)\right]$$
$$(1 - e^{-kt} \cos\, bt) \qquad (4)$$

$$\frac{z}{w_c} = \frac{N}{m}\, t - \left[\frac{Nk}{m^2} + \frac{g}{m} + \frac{k}{m}\left(\frac{V}{w_c} + \frac{N}{m}\right)\right]$$
$$(1 - e^{-kt} \cos bt) + \left[\frac{kg}{mb} + \frac{k^2}{mb}\left(\frac{V}{w_c} + \frac{N}{m}\right) - \right.$$
$$\left. - \frac{Nb}{m^2}\right] e^{-kt}\ \sin\ bt \qquad (5)$$

$$\text{where}\ \ b = \left[\frac{g}{\overline{\Theta}_c}\ (\Gamma_d - \Gamma_c)\right]^{1/2}$$

$$N = \frac{gkL}{\overline{\Theta}_c c_p} - gk - \frac{k^2 V}{w_c}$$

$$\text{and}\quad m = k^2 + \frac{g}{\overline{\Theta}_c}\ (\Gamma_d - \Gamma_c).$$

TABLE 1

Characteristics of the downward motion of dry air parcels through a cloud against a steady upcurrent of 1 m sec⁻¹.

Liquid water content of the cloud (g/kg)	0.3				1.0						3.0						
Mixing Rate, k: $10^{-3}\times$	1.0	2.0	4.0	6.0	1.0	2.0	2.5	3.0	4.0	6.0	0.5	1.0	2.0	3.0	4.0	5.0	6.0
Final R. H. %	98.6	100	100	100	99.1	96.7	98.5	100	100	100	92.6	92.2	91.8	95.7	99.7	100	100
"Absolute" Penetration (m)	533	439	319	260	1852	[1] 3000	[1] 3000	1513	1024	866	2740	[1] 3000	[1] 3000	[1] 3000	[1] 3000	120	120
Time Taken secs.	3600	1646	823	478	3600	2999	2466	1115	650	435	3600	1983	1004	759	598	130	125
"Relative" Penetration (m)	4133	2085	1142	738	5452	5999	5466	2628	1674	1301	6340	4983	4004	3759	3598	250	245
Final Temperature Deficit (°C)	0.11	0.15	0.26	0.37	0.32	0.40	0.44	0.48	0.62	0.83	0.86	0.81	0.93	1.46	1.70	3.14	3.42
Final Downward Velocity (m sec⁻¹)	0.16	0.27	0.50	0.77	0.53	1.04	1.27	1.45	1.65	2.61	1.10	2.02	3.19	4.68	5.18	3.77	5.1

[1] i. e. To cloud base.

Using the values of z as a function of t given by (5), x may be found numerically or graphically from the solution of (3):

$$x = x_0 e^{-kt} + k e^{-kt} \int_0^t e^{kt} (x_c\ (z) + w_c)\ dt \qquad (6)$$

where x_0 is the original mixing ratio of the parcel air. Saturation is reached when the value of x found from (6) equals the saturation mixing ratio of air at the temperature and pressure of the parcel.

When the parcel becomes saturated, it will continue to subside for a time because of its temperature deficit and downward momentum. Consequently, some of the liquid water diffusing in must continue to evaporate if the parcel air is to remain saturated. In these circumstances the modified equations corresponding to (1), (2) and (3) above must be treated as simultaneous. The equation to which they lead, however, is non-linear, so that the computations become more complex than in the case of the unsaturated parcel. As will be seen, the penetration of unsaturated parcels is sufficient to establish the quantitative adequacy of the downdraught hypothesis. Consequently, the point where saturation is first reached is taken as the terminus.

The numerical evaluations have been carried out for a cloud based at 1 km (16° C) with its top at 4 km (−2° C). Thus $\overline{\Theta}_c = 280°$K, $\Gamma_c = 6°$ C per km. The parcel entering the cloud at the moment $t = 0$ is assumed to have an initial relative humidity of 50 %. With a pressure corresponding to 4 km (ICAN) and a temperature of −2° C, this gives x_0 = 2.68 g/kg. The speed of the updraught, V, is taken as 1 m sec⁻¹. Three values of liquid water content are used, 0.3, 1.0 and 3.0 g/kg. The first two values may reasonably be taken as representing average conditions in dry and wet clouds respectively (WARNER and SQUIRES loc. cit.); the last value is approximately the mean of the adiabatic value of this cloud, which varies from zero at the base to 6.2 g/kg at the top. It has been included in order to illustrate what rapid and deep penetrations would occur in a cloud of such high liquid water content. The vigour with which dry air parcels would penetrate it may explain why such high values of w have so far been observed only in isolated regions of cumuli.

Calculations have been carried out on the basis of equations (4), (5) and (6). The results of interest are shown in Table I. The calculations have been terminated either when the parcel saturates, or when it subsides 3,000 metres and so reaches cloud base, or after the lapse of an hour, which ever occurs first. The "absolute" penetration in Table I is the final depth of the parcel below its starting point. "Relative" penetration is the depth of cloud above it at the end of the calculation, assuming that the cloud top continues to rise at the speed of the updraught after the parcel enters the cloud; since dry air parcels may enter the top of a cloud equally well at any stage of growth this figure gives an idea of how deeply they may become buried in the cloud as a result of their own subsidence, and the cloud's growth. For each value of the cloud liquid water content, only those values of k are shown which are necessary to illustrate the behaviour of parcels of varying size. As will be seen from equation (7) below, the parcel size varies inversely with k. Small parcels ($k > 6 \times 10^{-3}$) are strongly cooled and develop high downward speeds, but saturate quickly. Larger parcels ($k < 10^{-3}$) remain unsaturated for a long time, but, except in the unrealistic case $w = 3$ g/kg, they are only slightly cooled and subside relatively slowly through the updraught.

Table I indicates that parcels of a suitable size can penetrate deeply into growing cumuli in a period commensurate with the life-time of these clouds, especially when it is remembered that they can enter at all stages of growth ("Relative" penetration). The degree of cooling and the falling speed are both roughly proportional to the liquid water content of the cloud, and the values found for them are consistent with observed fluctuations in cloud, especially in the more realistic cases $w_c = 0.3$ and 1.0 g/kg.

An inspection of Table I will disclose some apparent anomalies – small changes in k sometimes give rise to large changes in parcel behaviour. These discontinuities are due to the fact that both the temperature deficit of the parcel and its downward velocity have an exponentially decreasing sinusoidal component (Equations (4) and (5)). The relative humidity of the parcel shows a similar behaviour. For certain values of the parameters, the parcel approaches saturation quite closely early in its history – one of the peaks of the relative humidity curve reaches close to 100 % – but nevertheless it remains unsaturated for a considerable time afterwards. In these circumstances, a small change in the parameter k may have the result that the peak exceeds 100 %, that is, the parcel saturates. This kind of discontinuity in the data of Table I does not necessarily imply an anomaly in the physical behaviour of the parcels. It is due to the arbitrary termination of the calculations at the point where saturation is first reached. Thus it seems likely that in a case like $w_c = 3$ g/kg, $k = 5 \times 10^{-3}$, where saturation is just reached at the first peak of the relative humidity curve, the parcel would in fact quickly dry out again and continue downwards for some distance before finally saturating.

It is impossible to give more than a very rough indication of the size of these parcels. However, it can be established that the optimal parcel size deduced from the calculations is not absurd in the physical context being discussed. Treating the eddy diffusivity, K, as constant, it is easily shown that

$$k = c\ K/R^2 \tag{7}$$

where c is a constant depending on the form assumed for the profile of properties across the parcel, and R is the parcel "radius"; in the case of a vertically elongated parcel, R would be the radius of its horizontal cross section. Thus, in any one cloud, k may be regarded as a measure of parcel size, and this remains valid even though the eddy diffusivity depends on the scale of the phenomenon, that is, the effective value of K really depends on R. Priestly (loc. cit.) has discussed the value of the constant c and has concluded that $c = 8$ is a reasonable choice. The correct value of c is not likely to depart from this by more than a factor of 2. However, there seems to be no way of arriving at an appropriate value for K, even within an order of magnitude. The eddy diffusivity which is effective in causing mixing between the parcel and the cloud is that due to eddies significantly smaller than the parcel. It will be much less than that which might be deduced, for example, from an aircraft accelerometer record, assuming the turbulence to isotropic, for this would include the effects of the motion of the parcels themselves. On the other hand, it will evidently exceed the small-scale background turbulence of the clear air outside the cloud, since the motion of the parcels, and of the upcurrent, will give rise to additional small-scale turbulence. It seems in any case unlikely that the appropriate value of K should lie outside the range 10^2 to 10^6 cm^2 sec^{-1}. Taking $k = 2.10^{-3}$ as the optimal value, the resulting value of R from (7) is found to be in the range 6 to 600 m. Thus the optimal parcel sizes deduced from equation (7) are plausible in relation to the phenomena under discussion.

Table 1 indicates that, for any one value of the effective eddy diffusivity, only a narrow range of parcel sizes are capable of achieving appreciable penetrations in a reasonable period of time. In actual clouds, the steady state assumed in deriving this result never occurs, and moreover the parcels must often be reduced in size by attrition or by disruption. A large parcel may enter the cloud early in its growth and, although incapble of subsiding far, may later act as a source of dry air for the region near cloud base. Thus no literal significance can be attached to the detailed values in Table 1. The calculations can be regarded only as showing that certain downdraughts (of a physically plausible size) are capable of achieving significant penetrations, so that the proposed mechanism for introducing dry air into a cloud is quantitatively adequate. There is, of course, nothing in the calculations themselves to indicate how often such downdraughts occur in clouds, but the physical arguments of Section 2 indicate that they are common.

5. Conclusions

Theories of the interaction between cumuli and their environment have been reviewed in the light of the existing observations. Physical arguments have been given which indicate that the region around the top of a growing cloud may be an important source of the dry air which mixes with the upcurrent. A theoretical model of the penetration of a cloud by dry downdraughts has been set up, and the calculations carried out for a cloud 3 km deep, with an upcurrent of 1 m sec^{-1}. As can be seen from Table 1, these calculations show that parcels of dry air can remain unsaturated while subsiding up to three kilometres (i.e. to cloud base) through the updraught in a time which is commensurate with the life-time of cumuli. Even after saturating, such parcels, being colder than the surrounding cloud, would continue to subside for some time as regions of relatively low liquid water content.

A large number of such parcels entering the cloud top at all stages of its growth would result in a wealth of fine structure in the distribution of liquid water content and other cloud properties, and would have the effect of reducing the liquid water content and of cooling the cloud, especially in its upper parts. Thus the hypothesis affords an explanation of the presence of fine structure and of the fact that the observed liquid water content is nearly always less than the adiabatic value. It also explains the upward decrease of the ratio $\overline{w}/w_a$, and the frequent occurrence of a cloud lapse rate steeper than the wet adiabatic.

On this view, the initial immersion of dry air parcels in the cloud is greatly accentuated when the cloud top becomes appreciably warmer than the environment. In addition, the deeper penetration which follows is achieved by motions which depend for their energy supply on the presence of liquid water, and which it appears have velocities roughly proportional to it. Thus a very wet cloud would be more susceptible to deep penetration than a dry one. The mechanism described therefore provides a type of automatic action which not only ensures that the lapse rate in the cloud shall not depart far from that of the environment, but also limits the concentration of liquid water. Both these aspects of the hypothesis accord well with the observations.

A corollary of the hypothesis is that there should be a radical difference in structure between convection in the clear air and in cloud; the presence of the liquid phase in cloud introduces an element of instability by providing a source of energy which is available to drive smaller scale motions. This situation has no analogue in the case of clear air convection.

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