



PROFESSOR TOR BERGERON

75 years on August 15, 1966

On the 15th of August this year Professor Tor Bergeron reached the age of seventy-five. This fact may come as a surprise for those who have the pleasure of meeting him frequently. On the other hand it is not surprising when considering Bergeron's fame and the fact that he has become a legend. For everyone who take a more than casual interest in synoptic meteorology or cloud physics the name of Tor Bergeron belongs to those most often seen in textbooks and scientific papers. The impact on weather forecasting all over the world of the so-called Bergen school and its members has been very strong indeed, and in this connection professor Bergeron has played a very important role.

He has had the ability not only to develop new and very fruitful ideas but also to bring them forward to the scientific world and to fight for them. Everyone who has met professor Bergeron and heard his lectures has come to admire his remarkable combination of great enthusiasm and sincerity with a profound understanding of physical processes in the atmosphere.

It is with great respect and strong feelings of gratitude that we now offer our homage to a Nestor of meteorology by bringing together a collection of papers in *Tellus* as a kind of "Festschrift".

Cumulus and cumulonimbus convection

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ABSTRACT

It is proposed that certain characteristic stratifications accompany cumulus and cumulonimbus, and on the basis of a thermal theory characteristic magnitudes are estimated for the properties of their updraughts, and the size which must be attained by cumulus before showers form and their transformation into cumulonimbus occurs.

1. Introduction

In 1933 Professor Bergeron brought forward his theory that "almost every real raindrop and all snowflakes originated around an ice crystal" formed in a cloud composed mainly of supercooled droplets (BERGERON, 1935). The strongest support for this view was seen in the transformation of cumulus into cumulonimbus, a "process almost invariably preceded by the glaciation of its upper part, occurring with the rapidity of a sudden release, and not before a certain level is reached (presumably the 'ice nuclei level')" (*loc. cit.*). The idea was strongly championed by a meteorologist so observant as Findeisen, and outside the tropics held sway for many years and perhaps still. Even if nowadays there is much evidence against the supposed crucial significance of the ice phase for the release of shower rain, the theory correctly described an important process, and without question has the strong appeal characteristic of a number of other brilliant ideas introduced into the study of weather phenomena by this distinguished analyst. I have had some experience of observing in the polar air masses of his favourite summer retreat, Jämtland, and believe that their revealing transparency must often have been his inspiration. In the following paragraphs the transformation of cumulus into cumulonimbus is discussed with the aim of representing the two forms of cloud as two different kinds of convection.

2. Kinds of atmospheric convection

Four principal kinds of convection can be distinguished in the circulations incessantly in progress in the troposphere, which can be ranked according to their scale:

1. *Small-scale convection*, including *cumulus convection*, on scales of up to a few km inside the lower troposphere, which communicates to the earth's surface from solar radiation. It is introduced partly as sensible heat and partly as the latent heat of evaporated water. This kind of convection is distinguished from the others in that throughout at least the lower part of the affected layer the potential density increases with height (i.e. there is static instability).

2. *Intermediate-scale convection*, with horizontal dimensions up to the continental, which consists of baroclinic circulations associated with irregular distributions of small-scale convection, imposed by the topographical features of the underlying surface. Familiar examples are the sea- and mountain-breeze circulations. With approach to continental scale, as in the Asiatic monsoons, it becomes impossible to distinguish this kind of convection from

3. *Large-scale convection*, on a horizontal scale of thousands of km, due to the general latitudinal temperature gradient, which outside the tropics effects a transport of heat not only vertically throughout the troposphere, but also from low to high latitudes.

4. *Cumulonimbus convection*, on horizontal scales of up to hundreds of km (sometimes organized into squall-lines, and such disturbances as easterly waves and hurricanes), which depends upon the presence of precipitation initially formed in small-scale convection. In low latitudes it is continually at work carrying heat into the upper troposphere, but elsewhere it occurs only sporadically and probably contributes little to the convective transport, which is performed mainly by the large-scale convection.

3. The collaboration of small- and large-scale convection

The various kinds of convection may be in progress simultaneously, and there is an essential collaboration of the small- and the large-scale or (in the tropics) cumulonimbus convection.

The small-scale convection generally occupies a rather shallow lowermost part of the troposphere supplied by the descending branch of a large-scale circulation. After some time the air attains a high potential temperature and enters the ascending branch, within which water vapour is condensed. Because most of it is here also precipitated, the descending branch of the circulation is cloud-free; the descent is accompanied by a radiative energy loss and a fall of potential temperature until the air again enters the region of small-scale convection.

The average magnitude of the radiative heat loss is set by the rate of absorption of solar radiation in the troposphere and at the ground, and is equivalent to a tropospheric cooling rate of between 1 and 2°C/day. Assuming that ascent in the large-scale circulations is rather rapid and nearly adiabatic, the air arrives in the upper troposphere with its potential temperature increased by the liberation of the latent heat of the original water content. If this, expressed as a mixing ratio, amounts to 10 g/kg, the increase of potential temperature is about 30°C, and since this must also be the magnitude of the decrease during the return towards the surface, the rather long descent-period (in the same latitude) of as much as 15 days or more is indicated. The mature large-scale motion systems are observed each to occupy typically about one-fifth of the middle-latitude belt in which they occur, so that air in the upper or lower troposphere moving with the observed relative speeds crosses one system within 2 to 5 days. The descent of air throughout the troposphere must therefore proceed through several systems, but because the small-scale convection is generally limited to only a small fraction of the depth of the troposphere, restoration of the potential temperature and ascent can be completed in one system.

Although usually the small-scale convection occurs in air which on the large-scale is descending, in places, particularly over land, it occurs where there is some large- or intermediate

scale ascent. The variations in the larger scale motion have a strong influence on the depth of the layer occupied by the small-scale convection (and hence upon the properties of the individual circulations), and upon the total heat, moisture and momentum transported into it. Whether the principal interest is in these transports and their effect upon the large-scale motion systems, or in the individual small-scale circulations (for example, in their capacity to produce cloud and rain), the interaction of the small- and larger-scale motions is a vital meteorological problem. It has to be faced that routine observational data will define only a rather large scale, and that the properties and effects of the small-scale circulations will have to be represented by some simple and approximately correct artifices devised on the basis of special observational and analytical studies. As already implied (LUDLAM, 1963, p. 3) it is for a start a formidable if not daunting task to obtain the special observations needed to define the properties of the small- and intermediate-scale circulations (overland, where it is most convenient to work, the latter may ruin the work if they are overlooked). The data so far available are fragmentary, consisting of measurements made at a few points or along a few lines rather than in all four dimensions, beset with instrumental difficulties inside clouds, and presented with much too little information on the intermediate and large scales. Their assembly into some kind of structure has to be speculative; in the following paragraphs an approach is outlined in terms of the bubble or thermal theory, as first proposed by SCORER and LUDLAM (1953), and later developed by SCORER (1957) and other authors.

4. The stratification associated with small-scale convection

Each kind of convection appears to be accompanied by characteristic distributions of the meteorological elements, such as velocity and temperature. The stratification of the atmosphere in the presence of small-scale convection is indicated by the radio-sonde observations made along a sloping flight-path or by tethered sounding balloons which can be moved rather slowly up and down, and rather better by aircraft flying in horizontal or spirally-ascending paths, preferably with sensitive and rapidly-

responding instruments which can determine not only the average temperature and moisture distributions, but also the fluctuations and perhaps velocity components associated with individual circulations. From the results of investigations by a number of authors (e.g. BUNKER *et al.*, 1949; JAMES, 1953; WARNER & TELFORD, 1963; GRANT, 1965) it appears that several layers can be distinguished, characterised by certain lapse-rates of potential temperature θ and mixing ratio x . With the decrease in the mean values of these parameters from the base to the top of a layer indicated by the prefix Δ_z , these layers are as follows:

1. The *surface layer*, from the surface to screen-level. $\Delta_z\theta$ is commonly several degrees Centigrade, sometimes as much as ten, especially over arid ground in intense sunshine. Δ_zx is also large, often reaching several g/kg. In this layer a considerable part of the upward heat flux may be due to radiation.

It is desirable to distinguish this layer simply because of the comparative abundance of routine observations at screen-level. It is unfortunate that screen thermometers are often not adequately ventilated; according to the hygrometric tables prepared for screen thermometers and for aspirated thermometers, screen wet-bulb temperatures in warm dry weather may ordinarily be as much as 0.8°C too high, and the error may be greater in calm conditions.

2. The *superadiabatic layer* extends from screen-level up to 100 to 300 m; $\Delta_z\theta$ is 1 to 3°C and Δ_zx is a few g/kg, the larger values occurring over the more arid terrain.

In this layer the individual circulations become organized from the fine structures causing shimmering near the ground to intermittent or transient wandering plumes or bubbles of buoyant air with vertical velocities approaching 1 m sec⁻¹, within which the temperature and mixing-ratio reach maximum values higher than those in the surroundings by about $\Delta_z\theta$ and Δ_zx , corresponding to the ascent without mixing of a proportion of the air from close to the ground.

3. The *adiabatic layer* is commonly 1 km or 2 deep, and when cumulus clouds are present reaches up to the bases of growing clouds and elsewhere to within 200 or 300 m of the cloud base level. This layer has a lapse-rate very close to the adiabatic, but is slightly yet distinctly stable; Δ_zx is usually small, perhaps

less than 1 g/kg, but $\Delta_z\theta$ varies from about -0.3 to -1.0°C, and $\Delta_z\theta_v$, where θ_v is the potential virtual temperature, is also negative and numerically only 0.1 to 0.3°C smaller.

Ascending air is encountered by sailplane pilots with vertical velocities often slightly in excess of 1 m sec⁻¹ (sufficient for soaring flight) at the base of the layer and reaching larger values, almost constant with height, of up to about 4 m sec⁻¹ higher in the layer. There is some dispute about the shape of the rising volumes. Over the most favourable sources there is evidently a wavering but rather persistent plume, but elsewhere the encountered lift is intermittent and may correspond to short-lived, perhaps travelling and inclined plumes, or to discrete volumes aptly described as thermals. According to the experience of glider pilots (e.g. WOODWARD, 1959, 1960) the air motions encountered are consistent with the structure of laboratory thermals. They report the air within updraughts to contain small-scale turbulence (SCORER & LUDLAM, 1952), in contrast to the air outside, in whose stable stratification there are probably no small-scale motions producing mixing, although there is sufficient disturbance for the pilot of a powered aircraft to experience bumpiness. Generally the terrain contains features of variable effectiveness as thermal sources, and accordingly it may be only the most buoyant or largest thermals which reach the level of the cloud base, many others reaching an equilibrium somewhere in the adiabatic layer, spreading out there and under the influence of wind shear giving it a laminated structure, most easily discernible in the moisture distribution. Moreover, during rise through the slightly stable surroundings, all thermals can be anticipated to suffer some erosion, so that the rate at which individual thermals broaden with height cannot be expected to be that shown by laboratory thermals. There is some evidence of broadening in the lower half of the adiabatic layer, followed by a shrinking (GRANT, 1965), and accordingly if a virtual origin is inferred for thermals near cloud base it is likely to be at a level well above the ground.

The heat flux associated with the convection must be a rather small fraction of the solar constant, so that the observed characteristic updraught speed of a few m sec⁻¹, itself related to the typical mean temperature excess (strictly, virtual temperature excess) of the thermals, is

consistent with this excess being only a small fraction of a degree. Therefore, regardless of the actual value of the flux, the mean lapse-rate in the rather deep adiabatic layer must be very close to the neutral dry-adiabatic.

4. *The transition layer.* Away from active clouds, some 200 to 300 m below the cloud-base level the lapse-rate of mixing ratio rather abruptly increases and the lapse-rate of temperature decreases, so that over a shallow transition layer up to the cloud-base level $\Delta_2 x$ is 1–3 g/kg and $\Delta_2 \theta$ is about -1°C . The air in this layer appears to have settled from the clear spaces between the clouds, without yet having been modified by the convection from below, which away from the cloud clusters does not effectively extend up to the cloud-base level. Horizontal flights in the transition layer accordingly recognise ascending air as more moist but cooler (by a large fraction of a degree) than its surroundings; it is above its equilibrium level, and if the clouds are shallow its deceleration is noticeable and glider pilots may be unable to reach them.

5. *The cumulus layer.* When the cumulus are more than a few hundred metres tall, some of the air within them regains buoyancy. Air enters their bases with about the potential temperature found in the adiabatic layer, but with a wet-bulb potential temperature and mixing ratio more clearly corresponding to the values found at some level in the upper part of the layer. Accordingly, in the cloud bases θ_w has a value between 1 and 2°C below that found at screen-level, and their level is about 200 m higher than the adiabatic condensation level calculated from screen observations (e.g. on the average by 650 ft in the data of PETERSEN *et al.* (1946), varying little with the cloud base height over the range 1500 to 6000 ft).

Within the layer occupied by the cumulus the lapse-rate is at first intermediate between the dry- and the wet-adiabatics, but eventually becomes less than the wet-adiabatic. The clouds are usually very variable in size and few become large; it appears that most are limited to the layer in which the lapse-rate exceeds the wet-adiabatic. Flight observations in cloud are hampered by difficulties due to wetting and evaporation at the thermometers, but clearly indicate that temperature is commonly up to 2 to 4°C lower than in the surrounding clear air,

and rarely higher by more than 1°C . The upward speeds encountered in cumulus more than 1 km deep reach several m sec^{-1} , up to about double those found within thermals in the adiabatic layer.

It is clear that the cloudy convective circulations are equally subject to strong mixing; the importance of the condensation is shown by the increase of θ with height, and the importance of mixing by the decrease of θ_w or of θ_s with height (both within and outside the clouds), where θ_s , the saturation potential temperature, is defined as the wet-bulb potential temperature corresponding to a state of saturation at the observed temperature. It seems reasonable to define the cloud layer as that between the cloud-base level and the level at which the lapse-rate becomes equal to the wet-adiabatic, where θ_s has a minimum value, and to regard a lapse-rate steeper than the wet-adiabatic as characteristic of a region extensively occupied by cumulus convection. Nevertheless, the tops of the largest clouds, comparatively few in number in a typical population, penetrate 1 km or even more above this level before a reversed buoyancy brings them to rest. They may therefore be found to be colder than their surroundings by rather more than the degree Celsius or less which can result from the evaporation of the cloud water by mixing with unsaturated air.

The convection of energy into the cloud layer must produce a proportionate rise of θ_w , but this may be more evident as a moistening than as a warming, and near the top of the cloud layer and in what we may call the *cumulus tower layer* above it, there may be a cooling.

Just as in the adiabatic layer a close approximation to uniform θ is to be anticipated by considering the heat flux carried by the convection, so in the cloud layer the distribution of θ_s and x in both the clear and cloudy air may be anticipated to be close to some state to be regarded as neutral for the process of cumulus convection. This view is strengthened by the finding that in some 20 sets of data in which soundings and cloud observations are available in some detail (including, e.g. data obtained in Jämtland (LUDLAM & SAUNDERS, 1956), special observations over England and the North Atlantic, and data obtained by Simpson and her collaborators in the trade winds), the mean lapse-rate in the cloud layer is nearly

always the same : at the top of the layer the temperature is about $2\Delta T/3$ higher than that corresponding to dry adiabatic ascent from the cloud base, where ΔT is the difference in the temperatures produced there by dry and moist (pseudo-)adiabatic ascent from the base. Thus there seems to be a mean lapse-rate, nearly halfway between the dry- and the wet-adiabatic lapse-rates, which is characteristic of the cumulus layer. It should be possible to demonstrate this and a characteristic lapse-rate of mixing ratio as a nearly neutral state, by some model of the convection in which a single rising thermal produces no significant modification. If it is allowed to reach an arbitrary level, to cool by the admixture of sufficient clear air to evaporate its liquid water content, and then subside to a new equilibrium level, it should arrive there with the vapour mixing ratio characteristic of the neutral stratification. (Tests, presuming thermals to rise with or without mixing, lead to final mixing ratios distinctly higher than shown on selected soundings, perhaps because they are insufficiently detailed or accurate. For example, the lapse-rate is greater than the mean value in the lower part of the layer and only gradually decreases with height, but such details of the temperature- and moisture-distribution are not well defined in the available data.)

5. The properties of cumulus clouds

Properties of the air inside cumulus which have a particular interest are the vertical velocity and condensed water concentration, especially in the study of shower rain formation and development. According to observations values of both are variable across the clouds and substantially less than those corresponding to the adiabatic ascent of air from the cloud base.

If the clouds consisted of thermals with a virtual origin near the ground, expanding by mixing during their ascent at the rate characteristic of laboratory thermals in neutral surroundings, then above the condensation level the evaporation accompanying the mixing with clear air of the observed properties would destroy the buoyancy after a rise of only about 1 km. Accordingly, the occurrence of tall clouds demands either a lesser degree of mixing (as explored, e.g., by SIMPSON *et al.*, 1965, using different jet- and vortex-models) or a means of

reducing its effectiveness. In the thermal theory (LUDLAM, 1958) large clouds are considered to be an assemblage of individual thermals, which intermittently appear rising as cloud towers. Those which reach highest are generally first seen emerging into clear air from the upper parts of the cloud bulk, which is regarded as the incompletely evaporated residues of previous towers. This view is based on the visible appearance and evolution of large clouds, and is supported by measurements which show the diameter D of emerging towers to increase linearly with height Z above a virtual origin between the cloud base and the ground, at a rate about equivalent to that of laboratory thermals (for which the ratio corresponding to D/Z is about 0.5; SCORER, 1957). This variation can be found in respect of individual towers for a short time before their size and rate of rise diminish ($D/Z \approx 0.5$; GLASS & CARLSON, 1963), and is expressed also as a relation between height and diameter upon emergence, over a height range of several km ($D/Z = 0.4 \pm 0.04$; SAUNDERS, 1961). It may be inferred that the towers represent thermals which expanded at the same rate within the cloud during rise above its base. It appears that some towers are able to reach high above the base because during the first part of the ascent of the thermals which they represent the mixing occurred with the incompletely evaporated residues of earlier thermals, and not with the unsaturated clear air.

According to this view the size of a cloud is likely to depend on the local rate of generation of thermals, or in effect, as is also indicated in other theories of cumulus structure, the local flux of buoyant air or the scale of the whole convective circulation. Just how this scale is determined is not known, although over land topographical features are obviously effective by tending to impose their own (intermediate) scales. This makes it difficult to predict how large cumulus clouds may grow. However, we can speculate about the properties of observed clouds with the hypothesis that they contain thermals expanding by mixing while ascending through thermal residues, which by evaporation and settling have acquired the same temperature (strictly, virtual temperature) as the surrounding clear air (presumed to have the mean lapse rate noted above), and a negligible liquid water concentration. All thermals must soon lose

buoyancy on emerging into clear air or after passing above the top of the cloud layer; however, the towers of large clouds, which have a large inertia, may rise a further km or more before the reversed buoyancy brings them to rest and they sink back and evaporate.

SAUNDERS (1961) has already shown from the Jämtland data that on this hypothesis the inferred buoyancy ($B = \overline{\Delta T}_v / T_v$, where $\overline{\Delta T}_v$ is the mean excess cloud virtual temperature) of a cloud tower is related to its observed diameter and the rate of rise (w) of its top by a Froude number ($C^2 = 2w^2/gBD$) which varies over the range 0.7 to 2.4 with a mean value of 1.5, close to the value of 1.2 found in experiments with laboratory thermals in neutral stratifications. If the experimental value is adopted then the hypothesis implies that within the cloud layer the active parts of the clouds are thermals which have characteristic average dimensions, temperatures, condensed water concentrations, buoyancies and upward speeds, which at a given level depend only upon the height and temperature of the cloud base, and the height of the virtual origin of the thermals. (In the cloud tower layer, or wherever the buoyancy may change rapidly, more elaborate considerations would be needed to estimate the rising speeds.) Over the sea and in castellanus clouds (such as the cumulus renewed by hill waves in the evening) the level of the virtual origin is probably near the cloud base level, but overland it is likely generally to be about halfway between the cloud base and the ground (because of erosion in the slightly stable adiabatic layer; over hills with a strong intermediate scale circulation the virtual origin may be lower). If the virtual origin is at the cloud base, then the average values of the temperature excess and condensed water concentration are only about $\frac{1}{4}$ of the adiabatic values; in the other case the corresponding fraction depends upon the ratio between the heights above the virtual origin of the cloud base and the level under consideration, and diminishes rather quickly from 1 at the cloud base to about $\frac{1}{2}$ a km above the cloud base, and approaches the value $\frac{1}{4}$ at higher levels. These values are not inconsistent with the available observations (LUDLAM, 1958). Inside clouds within the cloud layer the rising-speeds of the most active thermals should increase with height in warm climates (θ_w at cloud base 20°C)

at rather less than 2 m sec⁻¹ km⁻¹ over sea and rather more than this over land, rates which are about halved in cold climates (θ_w at cloud base 0°C), and which again are not inconsistent with observation (inside thermals the updraught speeds may locally be twice as great as the thermal rising speeds). However, observations of stratification and cloud properties much more detailed and more accurate than any now available would be needed to make a satisfactory test of these results, or to provide the basis of a better theory.

6. Shower formation

As BERGERON (1935) emphasised, shower formation occurs rather suddenly in cumulus whose tops reach a certain level, which varies strikingly with air mass from as little as 1 to more than 6 km above the cloud base, but which can be defined to within a few hundred meters on any one day (see, e.g., LUDLAM & SAUNDERS, 1956). Formerly it was presumed that the critical level indicated where the cloud temperatures had fallen to a degree at which for the first time some significant concentration of ice particles appeared. However, in laboratory clouds (e.g. in the slow expansions achieved by FINDEISEN & SCHULZ, 1944, which probably still are the experiments most relevant) the concentration of ice particles increases rather steadily with lowering temperature, so that it is difficult to specify some critical threshold. Moreover, the available observational evidence suggests rather strongly that commonly the first precipitation (i.e. particles of radius greater than about 10⁻² cm) consists of drops. This is quite obvious in low latitudes, where the entire cloud may always be well below the 0°C level, but it appears often to be true in temperate climates, where the large drops appear inside the upper parts of clouds at temperatures down to about -20°C (at which temperature they are all likely to become frozen). Observations both from the ground and from aircraft suggest that the ice phase develops extensively in the tops of many clouds only after the development of precipitation by coalescence, and not simply by the action of ice nuclei but by ice particle multiplication processes associated with the precipitation (e.g., DAY, 1956; LUDLAM, 1957; MURGATROYD & GARROD, 1960; BATTAN, 1963; BRAHAM, 1964). It is to be

anticipated that as cloud base temperatures and thicknesses, and therefore also condensed water concentrations, are lowered, the growth of ice particles, first by condensation and perhaps subsequently by aggregation with supercooled droplets, must eventually become the important process for the formation of any precipitation. It is evidently responsible for the production of fall-streaks from shallow middle-level droplet clouds in the temperature range from about -10°C to -30°C , and from cumulus less than 1 km thick seeded with artificial ice nuclei (see, e.g., MACCREADY *et al.*, 1964). However, except in rather arid continental or cold air masses, it seems probable that the onset of precipitation in cumulus is generally due to the coalescence of cloud droplets, and that the principal importance of the ice phase is in the development of the precipitation. Over a large range of conditions intermediate between those found over cold continents and over the warm oceans it may be extremely difficult to distinguish between the two processes; certainly a full description of their progress in individual clouds is a task beyond any facilities yet brought to bear on the problem.

Given some dynamical (or even only kinematical) model of the airflow in cumulus, it is possible to discuss the properties of the cloud droplets and the evolution of their size-distribution spectrum, and thereby to estimate the conditions for shower formation. The first discussions (e.g. BOWEN, 1950, and LUDLAM, 1951) assumed a steady updraught of constant speed. They followed the growth by coalescence of a single droplet of abnormal but plausible size which is carried up from the cloud base, to obtain a height where it reaches a size at which it can be regarded as a raindrop. This height was tentatively identified with that which a cumulus must attain in order to produce a shower. According to such simple treatments there are three important parameters. These are, first the size-distribution characteristics of the droplets developed by condensation near the cloud base (controlled by those of the aerosol particle population and the updraught speed at the cloud base; according to experiment and theory (WOODS & MASON, 1965), coalescence begins effectively only when some drops of radius greater than 20 to 30 microns occur amongst a general population of radii greater than about 7 microns); second, the concentra-

tion of condensed water, and third, the updraught speed. LUDLAM (1958) has discussed how the models may be elaborated, for example by lifting a growing droplet in a thermal whose condensed water concentration and ascent speed vary with height, and how fairly they may correspond to nature. It appeared from the Jämtland observations that the level at which incipient precipitation was seen in cumulus towers corresponded to that at which droplets already of radius about 40 microns at the cloud base attained a radius of 150 microns, if lifted in a thermal with the speed implied by the cloud tower observations and with the mean concentration of condensed water inferred from the thermal theory mentioned in section 5 above. Although these observations were made inland the air masses had all recently been over the ocean and were notably transparent, implying a small total concentration of aerosol particles but the presence of giant hygroscopic nuclei, so that a significant concentration of droplets with the rather large radius of 40 microns may well have been present near the cloud bases. As previously discussed (LUDLAM, 1958), concentrations of as little as $10^{-4}/\text{cm}^3$, below the threshold of most observational techniques, may be significant. Droplets of radius 20 microns are usually found near the bases of even small clouds in concentrations of order $1/\text{cm}^3$ in maritime air masses and $10^{-2}/\text{cm}^3$ in continental air masses. Their presence is difficult to explain, probably involving a prolonged existence in a succession of thermals, that is, depending essentially upon the mixing motions not included in the simple theoretical models. The mixing has the unfavourable effect upon the coalescence process of reducing the mean condensed water concentrations inside thermals to well below the adiabatic values, but is probably also important in the other direction by transferring a proportion of the cloud droplets from one thermal to another and hence in gradually extending their size-distribution spectrum at every level in the cloud. In this way the evolution of the whole cloud, and not only the history of one thermal, must become a factor in shower formation. Moreover, a significant proportion of droplets may travel on paths along which the concentration of cloud water is appreciably greater than the mean. As usual in meteorology, the interaction of motions on

more than one scale complicates the processes, and their four-dimensional nature makes even their description an extremely difficult task. At present, to summarise the implications of the simple thermal theory of shower formation, it can be said that on the basis of the Jämtland and other observations the variations in the thickness of cumulus in which showers form during widespread convection can reasonably be explained if the microphysics of the coalescence process is represented by a parameter r_0 , corresponding to an initial drop radius near the cloud bases which has the value of about 50 microns in maritime air masses over the ocean, 40 microns inland in air of recent maritime origin, and considerably less in continental air masses rich in Aitken but not giant condensation nuclei. The other important parameters appear to be the temperature of the cloud base and its height above the virtual origin of the thermals. Condensed water concentrations increase with the former (favouring coalescence), and with the latter, for then cloud thermals are larger and less affected by mixing (but their rising speeds are even more enhanced—hindering coalescence). Hence the thickness of cumulus needed for shower formation by the coalescence process increases from about 2–3 km in maritime air masses (r_0 about 50 microns, virtual origin of thermals close to cloud base) to 4–6 km inland (r_0 about 40 microns, virtual origin of thermals 1 km below cloud base), or even more in continental air masses. The higher values quoted are appropriate to cold or arid climates in which the cloud base temperature is low (about 0°C), but their significance is less apparent because if showers form then the process responsible is more likely to be the growth of ice particles (in such clouds a thickness of more than 4 km implies summit temperatures below -30°C).

Over land cumulus sometimes develop locally in air masses whose stratification has not been adjusted by previous widespread cloudy convection. In particular, down-wind of hot arid regions deep layers with a lapse-rate close to the dry-adiabatic may be entered by cumulus with bases in a lowermost moist layer of different origin. This happens rather frequently in summer over the middle-west plains of the U.S.A. In these situations the thermals within the cumulus may acquire exceptional buoyancies, and rising speeds of 10 to 20 rather than the

more usual several m sec⁻¹. The cloud thickness necessary for shower formation can then be anticipated to be as much as 8 km (LUDLAM, 1958), as observed by BROWNING & ATLAS (1965). The first precipitation is produced at heights where the temperature is so low that it must be frozen, whether or not it developed at first by the coalescence process.

7. The transformation of cumulus into cumulonimbus

Except in such abnormal situations, shower formation begins when the cumulus tops are in the middle troposphere. At least overland, and especially in the warmer climates in which cumulonimbus convection is prevalent, a great development in the size and vigour of the parent cloud frequently soon follows the formation of precipitation, as noted by THORKESSON (1946) in Iceland, CRADDOCK (1949) in Malaya, LUDLAM (1958) in Jämtland and England, and by HOWELL (1960) in Cuba. Within half an hour successive towers penetrate into the high troposphere, and the more intense and prolonged convection leads to their persistent glaciated residues spreading out into a mushroom or anvil cloud characteristic of the mature cumulonimbus. This has a shape quite different from that of the typical large cumulus, which is pyramidal, because the admixture of the cold clear air of the upper troposphere into the thermals is very inefficient in evaporating their condensed water and removing their buoyancy, so that they spread out and persist after reaching an equilibrium level (LUDLAM, 1958).

In the Jämtland observations it was noted that the glaciating towers of the cumulonimbus reached upward speeds of about 10 m sec⁻¹ in the high troposphere, and rose at several m sec⁻¹ through the level of shower formation which the towers of the large cumulus were barely able to reach. Since the glaciating towers possessed about the same diameter as a function of height, their behaviour implies that they acquired additional buoyancy, the cloud virtual temperature presumably increasing by the liberation of the latent heat of fusion during the growth of the frozen hydrometeors and by the reduction of the weight of condensed water during their precipitation. The maximum increment to be obtained in this way corresponds to the freezing of all the condensed water (0.6°C, if the

concentration of 2 g/kg indicated by the thermal theory is accepted), followed by its complete precipitation (giving a further increment of 0.5°C). Near the level of shower formation an increase of 0.5°C would have been sufficient to increase the buoyancy from zero to a value giving a rate of rise of 5–6 m sec⁻¹ according to the thermal theory, so that the freezing and precipitation of a substantial fraction of the condensed water could account for the observed behaviour. Evidently the ascent beyond the shower formation level could not continue far if the lapse-rate in the upper troposphere remained distinctly less than the wet-adiabatic. It is interesting that on these and some other occasions when cumulonimbus extended through the high troposphere, the soundings show that the lapse-rate, after decreasing above the cumulus layer, increased again to slightly more than the (pseudo-)wet adiabatic and then became only slightly less to near the tropopause; accordingly θ_s shows a slight maximum above the cumulus layer, where the temperature is below 0°C, which may be related to the release of the latent heat of fusion in the cumulonimbus. There may then be three successive layers with lapse-rates at first somewhat greater and then somewhat less than characteristic values, one close to the dry-adiabatic, one intermediate between the dry- and the wet-adiabatic, and one closer to the wet-adiabatic, corresponding to the three different processes of dry convection, cumulus convection, and cumulonimbus convection in the upper troposphere. In the upper troposphere little evaporation occurs during the mixing of cloud with clear air of low temperature, and the mean lapse-rate there can be expected to be little greater than that in a reference process intermediate between the pseudo-adiabatics with respect to liquid water and with respect to ice. Since in the high troposphere the latter leads to temperatures which are higher by 2 to 3°C, and the increase is still more if some of the condensate is retained and especially if some liquid is presumed to be carried above the 0°C level and frozen (see, e.g., SAUNDERS, 1957), the observed mean lapse-rate may be even closer to that shown by the pseudo-adiabatic with respect to liquid water which is generally used in the construction of aerological diagrams.

Updraught air reaches a new equilibrium and spreads out from the tops of cumulonimbus

in a layer probably not more than about 1 km thick, its top sharply defined by the upper surface of the anvil cloud and its base obscured by precipitation. If at the level of the anvil tops a value of θ_s is read from the sounding on an ordinary aerological diagram, it is found often to be within 1 or 2°C of that at the level of the cumulus bases, making it appear that in contrast to the condition in the rather shallow cumulus layer the mean lapse-rate over the entire cumulonimbus layer is nearly the wet adiabatic. For the reasons given above, however, it cannot straightway be inferred from this that in the cumulonimbus the ascent of air occurs with significantly less mixing. The available data are an unsatisfactory basis for more detailed analysis, because the observations have usually been of the height of cumulonimbus summits rather than of their anvils, because of the uncertainty of the magnitudes of the freezing and precipitation processes, and because of doubts about the representativity and accuracy of the radiosonde soundings in the high troposphere.

Massive seeding with ice nuclei of the tops of shower clouds which are barely supercooled may release sufficient heat in marginal stratifications to provoke a substantial rise in the height of the cloud tops. SIMPSON *et al.* (1965) provide strong evidence that they have accomplished such cloud modifications; their quantitative explanation is subject to the same difficulties of establishing sufficiently accurate and comprehensive observations and realistic dynamical models.

8. The influence of wind shear on cumulonimbus convection

During the period of the observations made in Jämtland the wind shear in the troposphere was very small. When it is more marked, at least in the lower troposphere, the cumulonimbus convection appears to become more organised, the intermittent rise of thermals giving place to a steadier flow, so that, for example, the top of the cloud is seen as a persistent dome above the anvil rather than as a succession of towers. The height attained by the dome, and the fall of large hailstones, shows that even in ordinary stratifications the speed of much of the updraught air reaches 20 m sec⁻¹ or more, rather than the 10 m sec⁻¹ or less which would

be anticipated from the thermal theory. The peculiarly greater intensity and duration of such cumulonimbus has been recognised by placing 'frontal' thunderstorms (i.e. those developing in zones of pronounced wind shear) in a different category from ordinary 'air mass' storms. Their organisation seems to depend upon the production of a downdraught chilled by the partial evaporation of precipitation. This draught spreads out at the ground as a cool squall behind a kind of cold front, over which the air ahead of the storm is lifted to supply the updraught (independently of the heating at the surface which is necessary to produce thermals, so that when developed the storms may persist overland into the night). Evidently in these cumulonimbus at least a proportion of the updraught air experiences significantly less mixing than if it ascended in thermals, and is more strongly accelerated (see, e.g., the discussion in LUDLAM, 1963).

In the especially unstable stratifications previously mentioned as fairly frequent during the summer in the middle-western plains of the U.S.A. and which occur occasionally elsewhere (CARLSON & LUDLAM, 1965), the acceleration is particularly great; in the upper troposphere updraught speeds of 40 m sec^{-1} and more are attained, and the cloud domes penetrate several km into the stratosphere. The associated storms become exceptionally violent, producing damaging squalls, tornadoes and giant hail. Browning proposes that they should be distinguished as SLR storms (*severe local storms* also characterised by their movement to the *right*—in the northern hemisphere—of the tropospheric winds). Their properties and the form of the air flow within them are extensively discussed in two recent references (ATLAS, 1963; BROWNING & FUMITA, 1965).

9. Conclusion

In the preceding paragraphs a summary of some of the available information on the stratifications accompanying small-scale cumulus and cumulonimbus convection has been used with a thermal theory to infer the properties likely to be characteristic of the principal updraughts in the clouds. Generally the updraughts are intermittent, and may be regarded as isolated volumes of air, buoyant during part of their ascent, which behave like

laboratory thermals, in particular expanding as they rise by a rather intense mixing with the surroundings. When convection is present in atmospheric layers about 1 km or more deep, carrying a heat flux which is only a fraction of the solar constant, the mean lapse-rate must be close to one which is neutral for a similar stirring process. This appears to be the dry-adiabatic below cloud base level, one nearly halfway between the dry- and the wet-adiabatics in the lower-tropospheric cumulus layer, and one nearer the wet-adiabatic in the upper troposphere (generally entered by clouds only after shower formation).

In the lower troposphere the evaporation of cloud water produced by the mixing motions is so important that the properties of the updraughts (such as excess temperature, speed, and degree of upward penetration) depend strongly on their scale, and hence upon ground features which impose a range of scales greater than that inherent in the depth of the layer with an unstable stratification. Since the large- (so-called synoptic-)scale circulations are related to the same solar constant, they also have a strong influence, primarily upon the depth of the affected layer and the total energy flux into it. The study of the evolution of convection on these interacting scales (as approached, e.g., by PRIESTLEY, 1959) is extremely difficult, and formidable logistical and instrumental problems must be overcome even to obtain the observations necessary for a satisfactory description.

However, in a given system of clouds, in which some are supposed to have become large where by "chance" or some intermediate-scale influence there is a local assemblage of thermals, the properties of the principal updraughts in the clouds can be estimated by regarding them as thermals rising through the incompletely evaporated residues of previous thermals. It then appears that in cumulus the mean temperature-excess and concentration of condensed water are similar small fractions of the values corresponding to adiabatic ascent from the cloud base, and are unlikely to considerably exceed 1°C and 1 g m^{-3} . The rising speeds of the thermals are dependent upon height above cloud base, and increase with the temperature and height of the cloud base above the virtual origin of the thermals. Over the sea the cloud base level is usually low

(presumably because energy is introduced at the surface more as latent than as sensible heat), and in large cumulus rising speeds of a few m sec⁻¹ are to be anticipated, rather less than the several m sec⁻¹ over warm land.

Depending upon these speeds and also on the microphysical properties of the population of cloud particles, showers form in cumulus which attain a certain thickness (commonly only 2 km over the sea, but often 4 km over land and even more in arid climates), usually by droplet coalescence. In warm maritime climates the clouds may be wholly below the 0°C level, but elsewhere their tops are likely to be supercooled. It seems that the freezing of raindrops then initiates processes of ice particle multiplication which lead to rapid and extensive glaciation of the cloud tops. In clouds with cold bases the showers and glaciation may develop by the Bergeron process of ice particle growth before the coalescence process reaches an advanced stage. In either case in shower clouds the freezing and precipitation of cloud water appears to increase significantly the buoyancy of thermals, so that with the development of the shower they may penetrate into the high troposphere with appreciably increased rising speeds, the cumulus transforming into the much larger cumulonimbus. In the upper troposphere the comparative unimportance of the evaporation of cloud water, and the release of addition latent heat as the cloud particles become predominantly solid, leads to mean lapse-rates in the presence of cumulonimbus which are rather close to the ordinarily used pseudo-wet-adiabatic with respect to liquid water.

In the presence of favourable wind shear the cumulonimbus convection becomes more or-

ganised and the ascent of at least a proportion of the updraught air occurs with significantly less mixing than during rise in thermals. In the more nearly adiabatic motion the buoyancy increases, and updraught speeds of 20 m sec⁻¹, and, exceptionally, more than 40 m sec⁻¹ are attained. These more intense cumulonimbus seem to depend upon the production of a downdraught chilled by the partial evaporation of rain, and are marked by cool squalls at the surface. They are particularly violent in some continental regions outside the tropics, but occur also within the tropics (LUDLAM, 1963). There they contribute to the more general cumulonimbus convection which is the principal if not the only agency extending the troposphere to the observed heights. Accordingly, in the regions which are the principal heat sources and where the cumulonimbus are prevalent, θ_w or θ_s (defined in terms of a pseudo-adiabatic liquid-water process) has a mean value near the tropopause rather close to that of θ_w near the surface. For example, in July inland over India and S. E. Asia θ_w has a mean value which is 24 to 27°C at 150 mb and 27 to 29°C at screen-level (in mid-afternoon). Even in the mean, θ_s (and, of course, also θ_w) has a distinctly lower minimum value near the 500 mb level because of the considerably greater mean lapse-rate in the cumulus layer; thus at 500 mb the mean θ_s varies between about 22 and 26°C. Elsewhere in low latitudes θ_w near the surface is less and the cumulonimbus convection cannot extend so high, but even over the oceans the same characteristic stratification is found (MALKUS, 1962), with the minimum in θ_s at a level (between 700 and 600 mb) consistent with the lower top of the cumulus layer to be anticipated over the sea.

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КОНВЕКЦИЯ В КУЧЕВЫХ И КУЧЕВО-ДОЖДЕВЫХ ОБЛАКАХ

Предполагается, что кучевым и кучево-дождевым облакам соответствует определенная характерная стратификация. Оцениваются характерные величины для параметров восходящих потоков, а также размеры, которых

должны достигнуть кучевые облака прежде чем образуются осадки и произойдет трансформация кучевых облаков в кучево-дождевые облака.