METEOROLOGICAL OFFICE COLLEGE

CUMULONIMBUS CLOUDS

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CUMULONIMBUS CLOUDS

1. Introduction

All atmospheric motion is ultimately caused by solar heating but there are many ways in which this energy becomes available. One of the more direct is through air being warmed by contact with the ground. Some surfaces, for example hill slopes facing the sun, are better than others at absorbing solar radiation and this leads to local increases in the air temperature. In these regions the air can become buoyant and rise as a thermal. However, as it rises, the ambient pressure decreases and the air expands, cooling adiabatically, as the energy for expansion is drawn from within.

Water vapour is an important but variable component of the atmosphere, contributing typically one per cent of the total mass. As air cools, its capacity to retain water in the vapour form decreases, until at a certain height, referred to as the condensation level or cloud base, it becomes saturated, and cloud forms by condensation onto the small particles that are always present in the air. The condensation process releases latent heat which warms the air and maintains the buoyancy of the thermal.

At the beginning of the condensation process the individual cloud droplets within the thermal are small, typically 1 or 2 μm in diameter. They grow rapidly, initially by condensation, but as they become larger, by mutual collisions and coalescence. In deep clouds some droplets begin to freeze at temperatures of about -10°C, whilst at about -40°C (typically at a height of 7 km) almost all cloud particles are ice. The drops of water and ice crystals if present, eventually become so heavy that they fall through the buoyant updraught and grow even larger by 'sweeping up' smaller droplets. They emerge from the base of the cloud as raindrops and hailstones which then fall to the ground. Below cloud base the air is unsaturated and often above the freezing point. In consequence the hail begins to melt and the rain partially evaporates, both processes leading to a cooling of the air as the latent heats of melting and condensation are extracted. The cold air sinks and forms a downdraught which spreads out on encountering the ground to produce a gust front at its boundary with the ambient air. This creates the localised, strong, cold winds that are frequently associated with thunderstorms.

Thunder and lightning, although spectacular, are in fact of minor importance to the storm evolution, and severe storms can occur without their presence. The severity of a storm is essentially determined by the strength and duration of its up and downdraughts, for which the above descriptive model gives a qualitative explanation. However we invariably wish to quantify the effects, for unless the amount of rain and the strength of the winds can be related to the initial environmental conditions, it will not be possible to give advanced warning of such storms.

The experimentalist approach to such a problem would be to observe changes in the evolution of the storm resulting from slight changes in the initial conditions. The important aspects might then be isolated and quantified. Unfortunately, in the atmosphere this approach is not viable. The researcher has no control over the phenomena that he or she is studying, but can only observe what the atmosphere provides. Furthermore, there are several practical difficulties. First, storms are widely scattered in both space and time. It is difficult, therefore, to be in the right place at the right time to make the measurements. Secondly, the storms are both large, typically 25 kilometres in diameter, and evolve quickly. A storm may last from three to six hours, but conditions within it vary much more rapidly, typically over 10 to 15 minutes. The difficulty, therefore, is to obtain the necessary spatial resolution sufficiently often to resolve the changing conditions. Thirdly, there is the problem of physically obtaining the measurements. Some information can be obtained remotely, for example by radar, but many of the measurements can only be made from an aircraft and there are very few of these which are both equipped with the necessary instrumentation and able to withstand the stresses imposed by flying within such storms.

These difficulties, which are present to a greater or less degree in all meteorological investigations, have necessitated an alternative approach, based on the construction of mathematical models.

2. Parcel Theory

The simplest of the models is called parcel theory. This is demonstrated in Fig 1. Initially the air is unsaturated and, below the point A, the parcel conserves both θ and r. But at A the saturation point is reached and from then on ascent follows the wet bulb potential temperature isotherm. Throughout the ascent from A to B, latent heat release is sufficient to keep the parcel buoyant and therefore it accelerates. In the absence of friction (turbulent diffusion) all the thermodynamic energy released between A and B (shaded area) is translated to kinetic energy and the parcel attains its maximum vertical velocity at B. Above B the parcel is colder than its environment and slows down temporarily coming to rest at C, before descending. In the absence of friction the shaded and hatched regions have equal areas.

However friction, in the form of turbulent diffusion, and mixing with dry environmental air, slow the parcel down and in practice it rarely 'overshoots' the point B. Unfortunately it is difficult to quantify friction and mixing in this simple model as the effect of these features depends on the position of the parcel within the cloud. Parcels near the edge experience more mixing than those near the centre and the effect on parcels near the centre depends crucially on the size and rate of growth of the cloud.

To fully quantify these effects it is necessary to develop more sophisticated models. These form a rather special category in numerical modelling as they are non-hydrostatic and are driven by latent heat release. The motions that they simulate result from a complicated feedback mechanism whereby condensation processes release

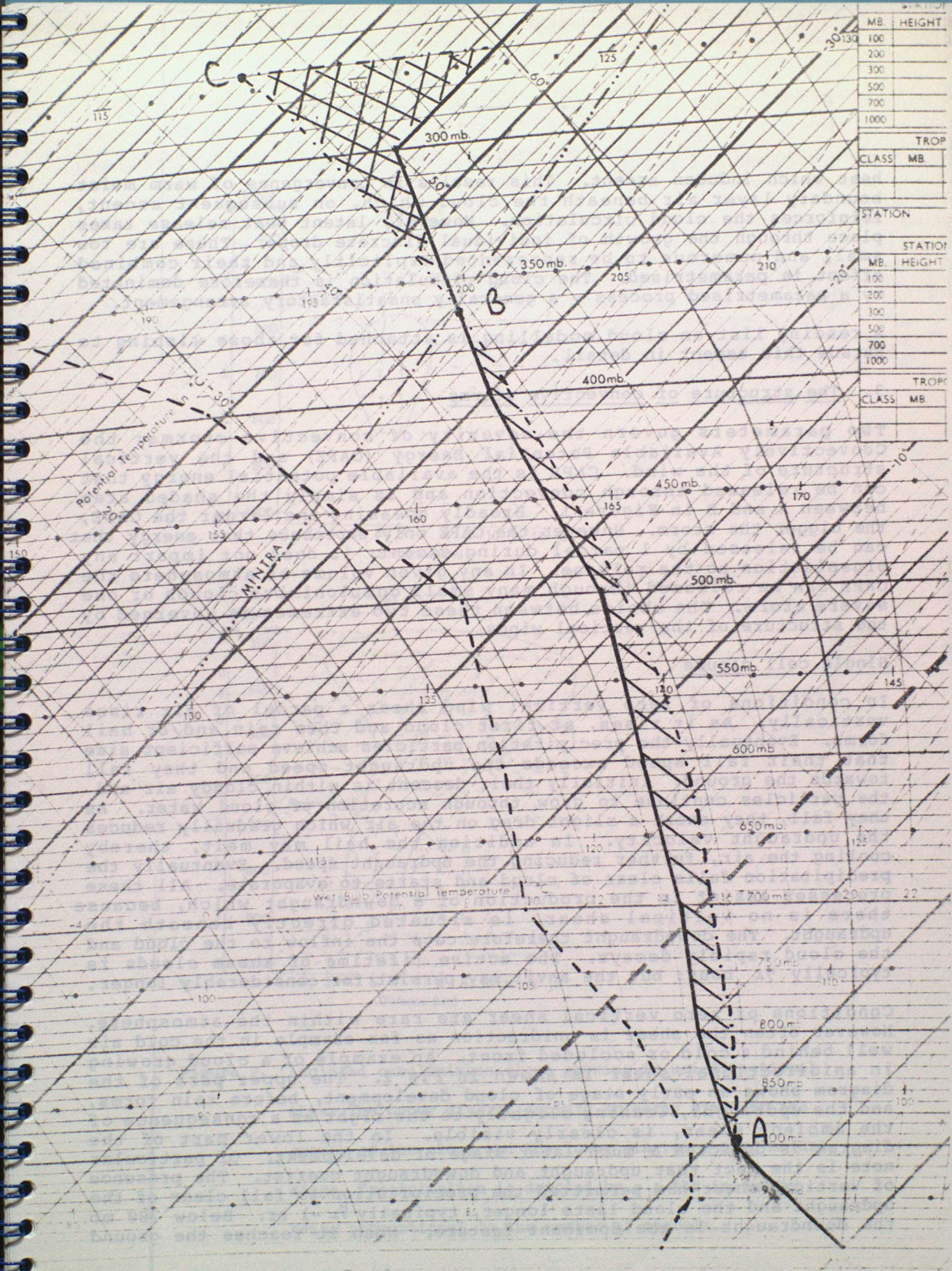


Figure 1. Parcel Theory.

heat which induces ascent. This results in convergence of warm moist boundary layer air beneath the cloud which, on subsequent ascent, reinforces the cloud circulation. However, latent heat release takes place through the growth of individual discrete drops. These are too small and numerous to be represented explicitly and their combined effect is parametrised. The cloud circulation is therefore dominated by a parametrised process - a generally unsatisfactory arrangement.

A reading list on cloud modelling is attached for those wishing to pursue this aspect in detail.

3. The structure of convective storms

Two parameters govern the severity of convective storms; the Convectively Available Potential Energy (CAPE) and the vertical structure of the wind. CAPE is the available potential energy that can be released through convection and is simply the shaded area between A and B in Figure 1. Broadly speaking the larger the CAPE, the bigger the storm. However the CAPE only expresses that energy that can be released by 1 parcel during ascent; it does not impart any organisation to the release. In any given volume of atmosphere the CAPE can be released through many small cumulonimbus clouds or one severe storm. The choice between these two extremes is governed by the structure of the vertical winds.

Single cell clouds

In conditions of zero vertical wind shear a parcel of air rises vertically. As it rises, at first cloud and then rain and/or hail forms. Eventually the precipitation particles achieve sufficient size that their fall speed exceeds the updraught speed and they fall towards the ground. Initially their descent is within cloudy air and the particles continue to grow through accretion of cloud water. As they fall, they exert a slight drag on the air which gradually reduces the updraught velocity. In addition the hail may melt, thereby cooling the air, further reducing the updraught speed. Eventually the precipitation falls clear of cloud and starts to evaporate. All these processes assist in the production of a downdraught which, because there is no vertical shear, is situated directly beneath the updraught. The downdraught therefore cuts the inflow to the cloud and the cloud rapidly decays. The active lifetime of these clouds is typically ½ hour: but the anvil may persist for considerably longer.

Conditions of zero vertical shear are rare within the atmosphere. However often the shear is unidirection as for example in the cold air well behind a cold or occluded front. An example of a cloud growing in unidirectional shear is shown in Fig 2. The upper part of the diagram shows an early stage of cloud development, before rain forms, and the updraught, leaning slightly to the right as a consequence of the ambient shear, is clearly visible. In the lower part of the diagram is depicted a much later stage of development. Of particular note is the fact that updraught and downdraught coexist. The presence of vertical shear has permitted the precipitation to fall clear of the updraught and the cloud lasts longer, typically ½ -1 hr. Below 700 mb the downdraught is the dominant feature. When it reaches the ground

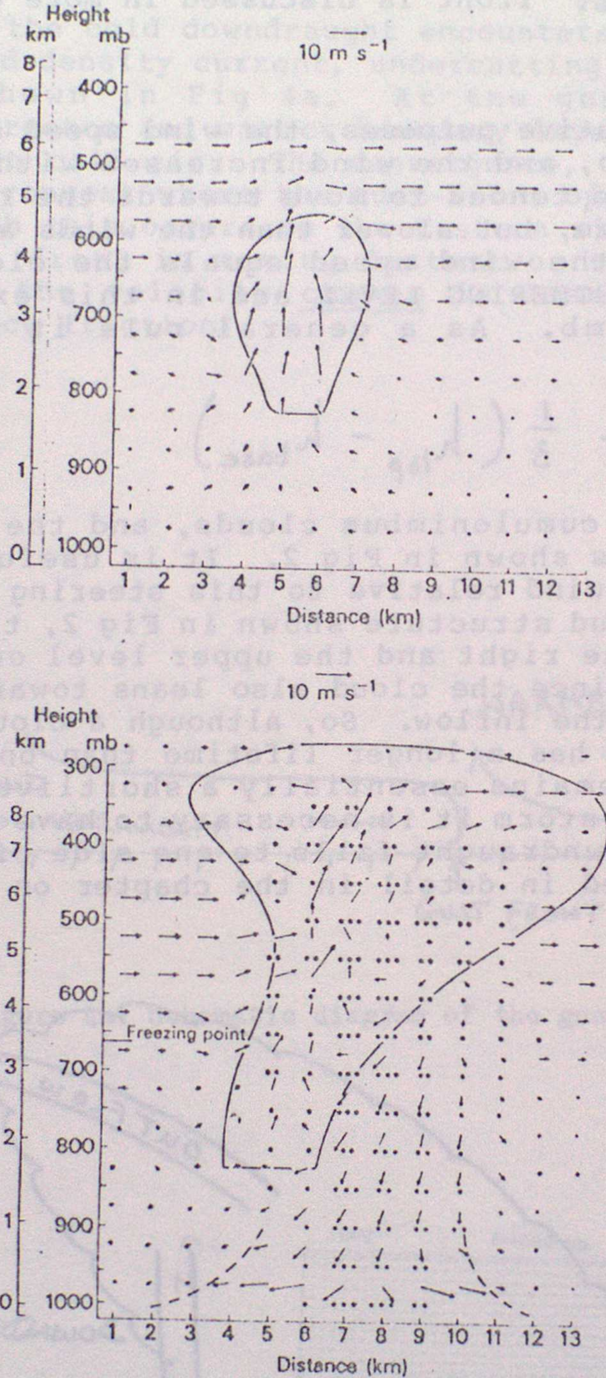


Figure 2. Isolated cumulonimbus cloud growing in uni-directional shear. The upper diagram depicts an early stage in the growth of the cloud and the lower diagram a more mature stage. Precipitation is denoted by • ; the intensity being proportional to the number of symbols.

it spreads out as a gust front, which undercuts the warmer, environmental air. The gust front is discussed in more detail below.

a) Steering Level

In Fig 2, for illustrative purposes, the wind speed at the ground was chosen to be zero, and the wind increased with height. In consequence the cloud tended to move towards the right at some speed greater than zero, but slower than the winds at high level. The level at which the wind speed equals the cloud speed is referred to as the STEERING LEVEL and in this example is at approximately 700 mb. As a general rule it will lie at approximately

$$h_{base} + \frac{1}{3}(h_{top} - h_{base})$$

for all single cell cumulonimbus clouds, and the clouds will slope "down-shear" as shown in Fig 2. It is useful to look at the structure of the wind relative to this steering level, as in Fig 3. Using the cloud structure shown in Fig 2, the low-level inflow comes from the right and the upper level outflow moves towards the right. Since the cloud also leans towards the right the downdraught cuts the inflow. So, although a cloud growing in uni-directional shear has a longer lifetime than one growing in no-shear, it still remains essentially a shortlived storm. To achieve a long-lived storm it is necessary to have 'directional shear' so that the downdraught falls to one side of the inflow. This will be discussed in detail in the chapter on multi-celled storms.

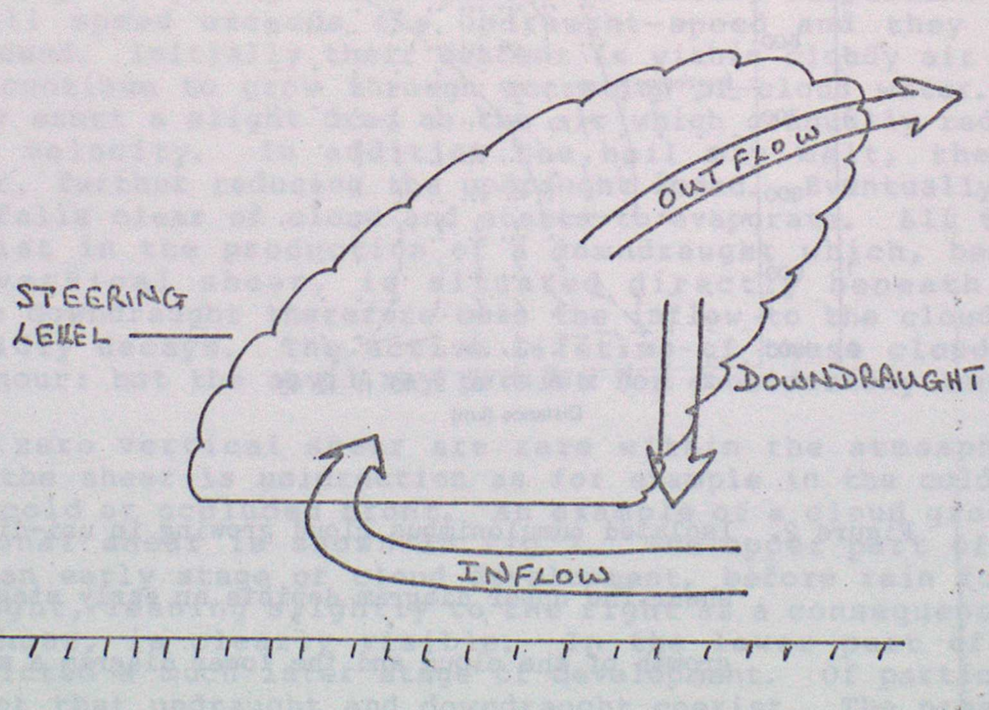


Figure 3. Schematic diagram of internal, cloud-relative motion.

b) <u>The gust front</u>

When the cold downdraught encounters the ground it spreads out as a cold density current, undercutting the warmer environmental air as shown in Fig 4a. At the gust front there is strong convergence and warmer boundary layer air is lifted over the nose of the cold air. The convergence, and hence the induced ascent, is strongest where the gust-front opposes the low level wind. In Fig 4b this occurs on the Eastern side of the downdraughts. (The gust front is moving fastest on the western side but on that side the velocity <u>change</u> is less because of the effects of surface friction).

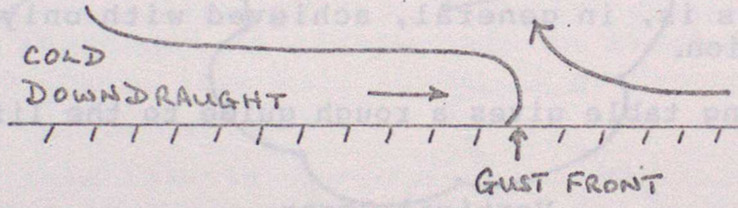


Figure 4a. Schematic diagram of the gust front.

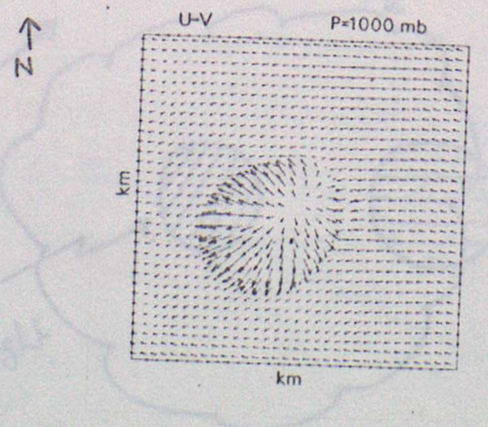


Figure 4b. Velocity vectors at 1000mb showing the downdraught spreading out beneath a cumulonimbus cloud.

4. Multicelled Storms

In the previous section it was noted that in conditions of no vertical wind shear, cells were shortlived because the downdraught developed directly beneath the updraught cutting off the supply of moist low-level air. Unidirectional shear lengthened the lifetime of the cloud by permitting updraught and downdraught to co-exist for a short time, but even in these conditions, the inflow was eventually disrupted by the downdraught.

However, if the downdraught could be displaced to one side of the inflow, much longer lived cloud might develop. This occurs if the wind changes direction with height, for example consider the hodograph shown in Fig 5a. Since the wind changes direction with height, there is no longer a 'steering level'. However the cell speed remains similar to the 700 mb wind and is always within the 'envelope' (defined by the dashed line) of the hodograph. Relative to the cell the low-level wind enters the cloud from the NNE and leaves towards the NE. In consequence air descends within the cloud on the Eastern flank, away from the inflowing air, as shown in Fig 5c. Obviously as the downdraught spreads out it will eventually cut the inflow, therefore the more the downdraught is separated from the inflow the longer elapsed time before it is cut, and hence the longer the life of the cell. However, for cloud physical reasons the maximum lifetime of a cell is ~ 1 hr and this is, in general, achieved with only moderate changes in wind direction.

In summary the following table gives a rough guide to the lifetimes of individual cells.

Time	Vertical Shear
½ hr	No shear
↓	↓
	Unidirectional shear
↓	↓
1 hr	Directional shear

Table 1

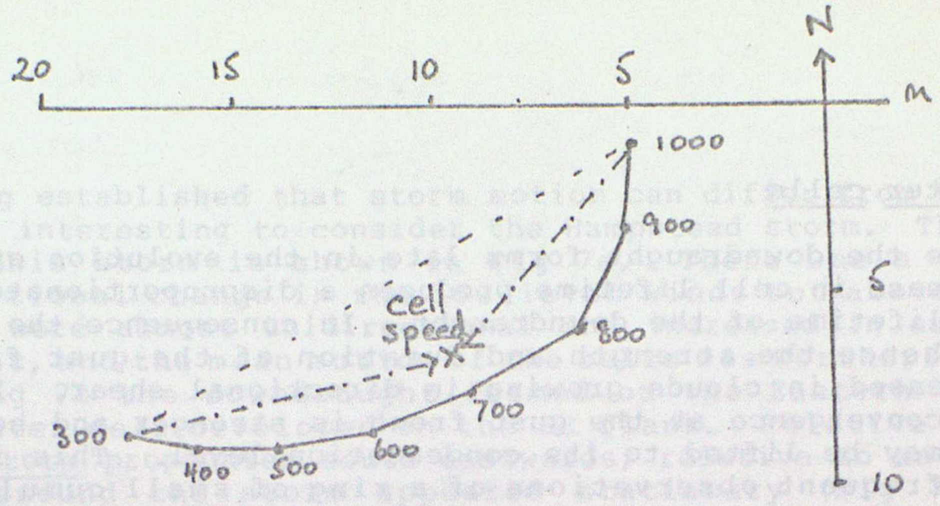


Figure 5a. Hodograph showing directional change in wind with height.

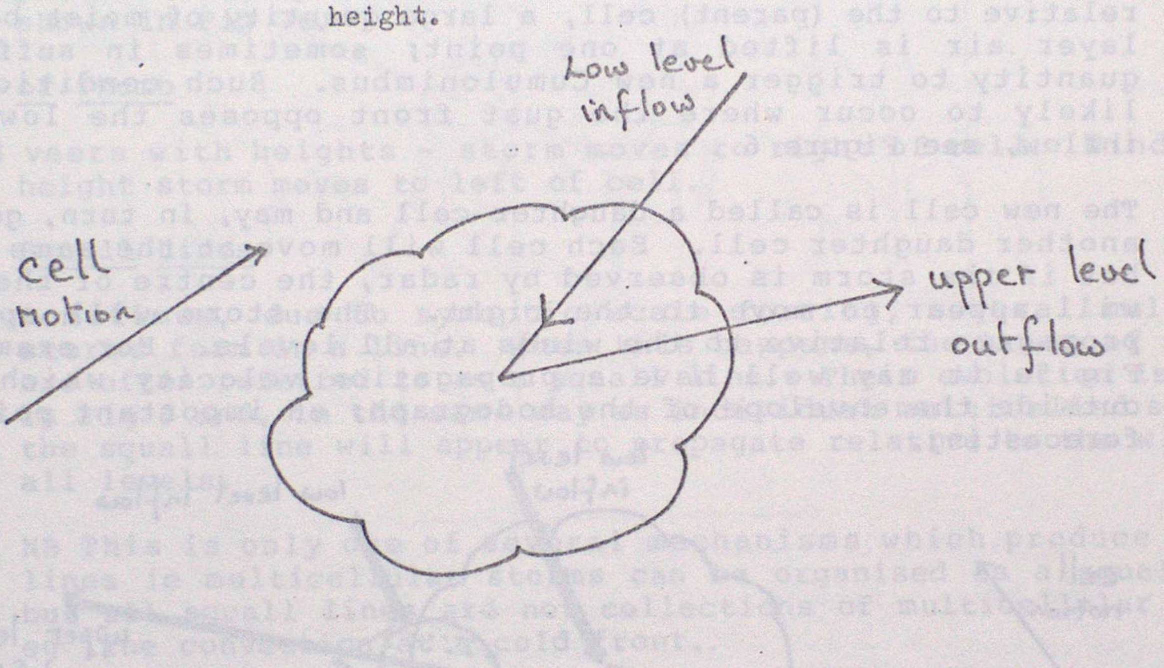


Figure 5b. Directions of cloud-relative inflow and outflow.

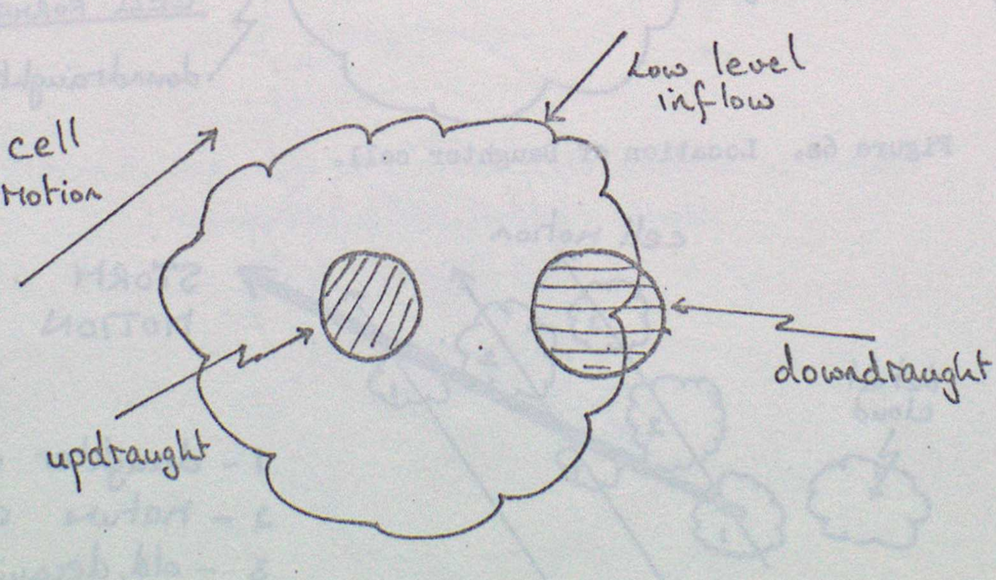


Figure 5c. Positions of updraught and downdraught.

a) Daughter cells

Since the downdraught forms late in the evolution of a cell, any increase in cell lifetime produces a disproportionate increase in the lifetime of the downdraught. In consequence the downdraught, and hence the strength and duration of the gust front is much increased in clouds growing in directional shear. In such cases the convergence at the gust front is stronger and boundary layer air may be lifted to its condensation level. This is evident in the frequent observations of a ring of small cumulus round the base of mature Cumulonimbus clouds. If the gust front is moving fast, the clouds are generally small and rapidly decay. However, if the gust front becomes quasi-stationary, ie slow moving relative to the (parent) cell, a large quantity of moist boundary layer air is lifted at one point; sometimes in sufficient quantity to trigger a new Cumulonimbus. Such conditions are likely to occur where the gust front opposes the low level inflow, see Figure 6 .

The new cell is called a daughter cell and may, in turn, generate another daughter cell. Each cell will move at the same speed, but if the storm is observed by radar, the centre of the storm will appear to move to the right. The storm will appear to propagate relative to the winds at all levels. For example in Fig 5a it may well have a propagation velocity which falls outside the envelope of the hodograph; an important point for forecasting.

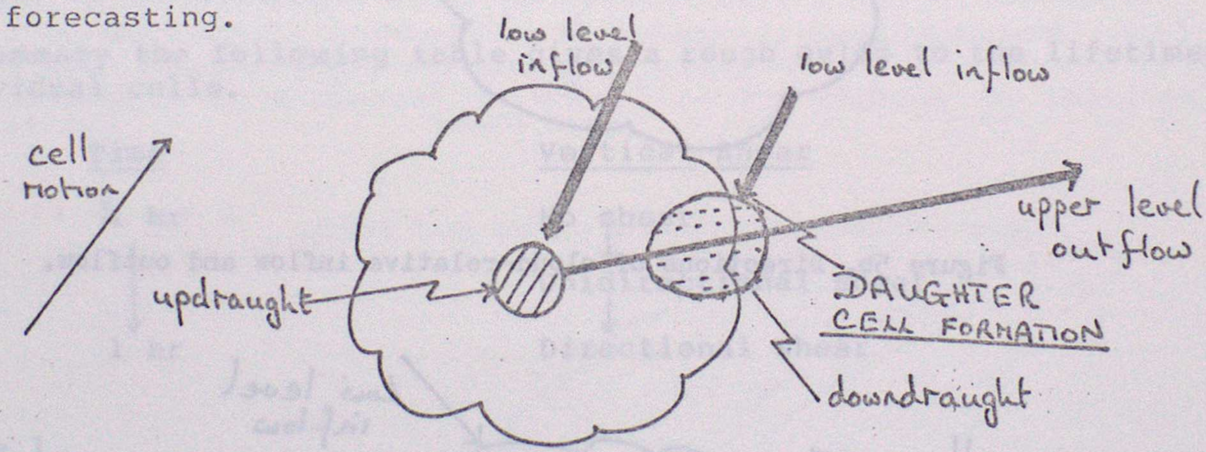


Figure 6a. Location of Daughter cell.

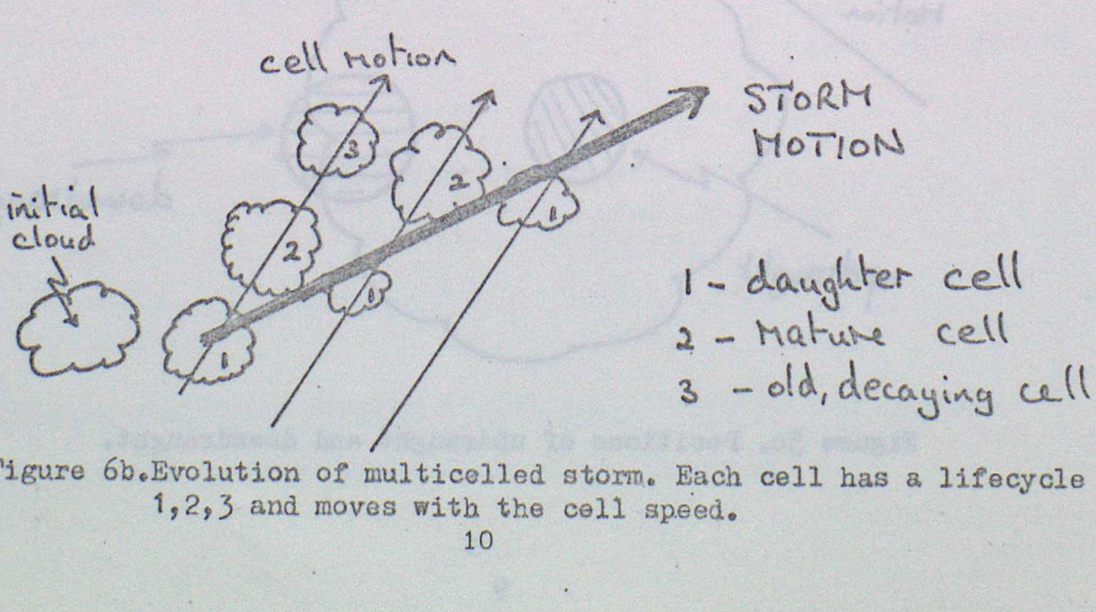


Figure 6b. Evolution of multicelled storm. Each cell has a lifecycle 1,2,3 and moves with the cell speed.

Having established that storm motion can differ from cell motion, it is interesting to consider the Hampstead storm. The hodograph for this storm is shown in Fig 7a. There was a very marked directional change in the low-level winds but above 800 mb the wind were almost unidirectional and increased in strength with height, and the mean motion of the cells was Northerly. As shown in Fig 7b the downdraught formed on the Eastern flank and a daughter cell developed on the SE flank. Relative to the cell the storm propagated south eastwards; relative to an observer on the ground the storm appeared stationary (Fig 7a). To an observer, cells formed towards the SE, deposited their rain overhead and decayed as they moved towards the North. This pattern was repeated 5 or 6 times and the accumulated rainfall is shown in Fig 7c.

Rule of Thumb

Wind veers with heights - storm moves to right of cell. Wind backs with height storm moves to left of cell.

b. Squall Lines

Sometimes, due to synoptic scale forcing, several multicell storms form in a line. When this happens, the group of storms are often referred to as a squall line. Their evolution is shown in Fig 8 and, in the same way as individual multicelled storms, the squall line will appear to propagate relative to the winds at all levels.

NB This is only one of several mechanisms which produce squall lines ie multicellular storms can be organised as a squall line but all squall lines are not collections of multicellular storms eg line convection at a cold front.

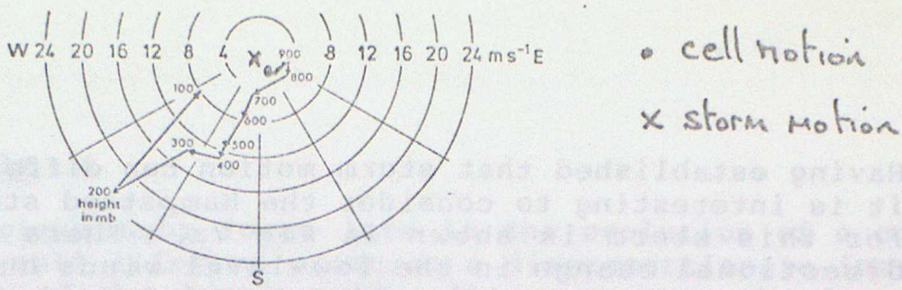


Figure 7a. Hodograph on the day of the Hampstead storm.

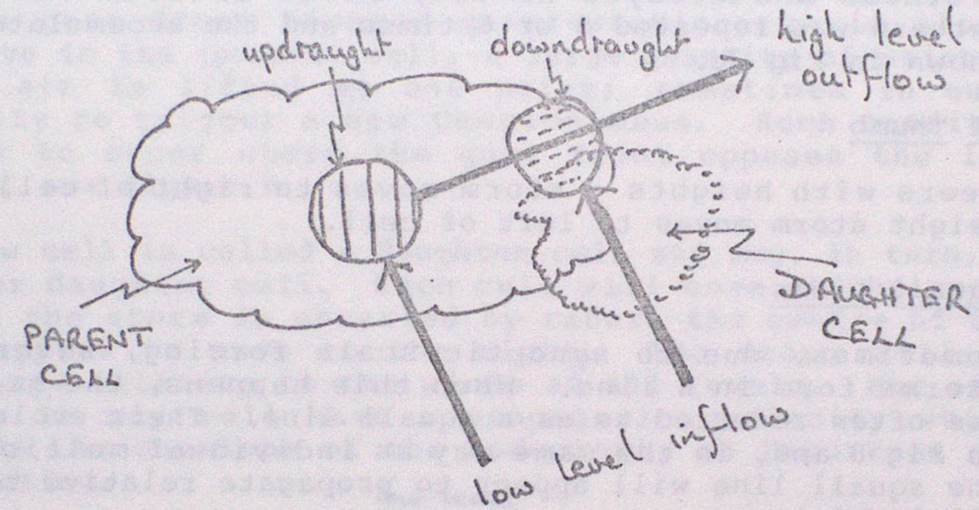


Figure 7b. Cell relative positions.

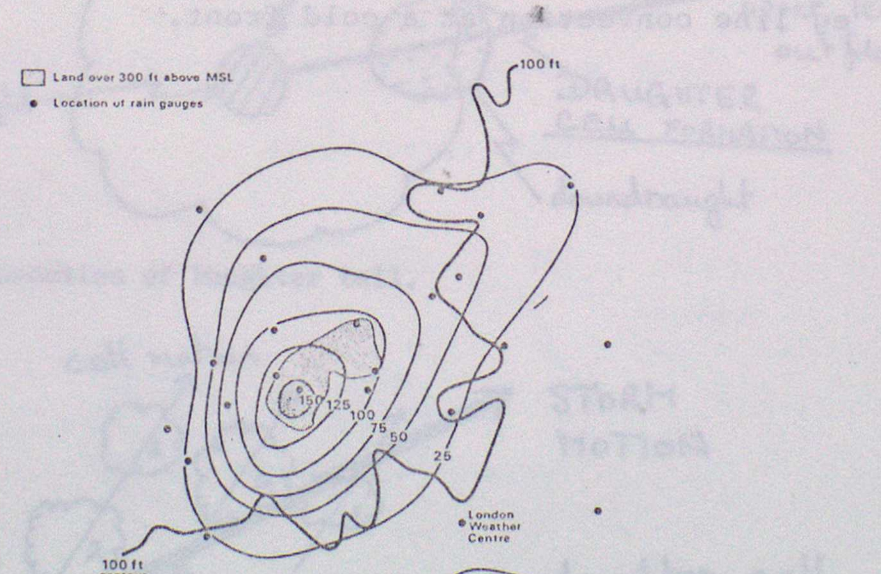


Figure 7c Isopleths of rainfall (at intervals of 25mm) for the period 09 GMT 14th to 09 GMT 15 August 1975 over north London. The map includes the 100 and 300 ft height contours.

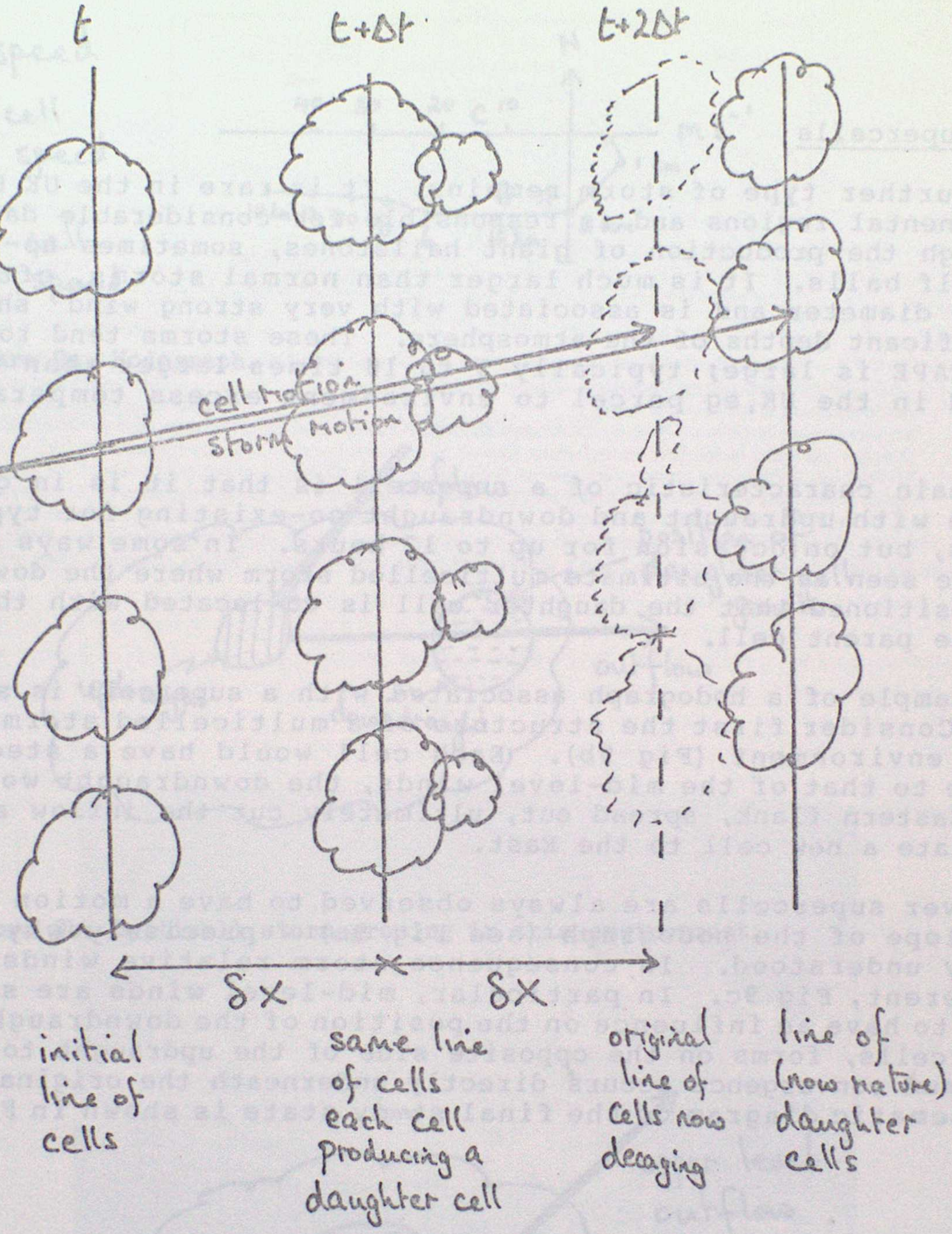


Figure 8. Evolution of a squall line composed of several multicellular cumulonimbus clouds.

5. Supercells

One further type of storm remains. It is rare in the UK but common in continental regions and is responsible for considerable damage mainly through the production of giant hailstones, sometimes up to the size of golf balls. It is much larger than normal storms, often up to 50 km in diameter and is associated with very strong wind shears over significant depths of the atmosphere. These storms tend to occur when the CAPE is large; typically 5 to 10 times larger than the values found in the UK, eg parcel to environment excess temperature up to 10°C.

The main characteristic of a supercell is that it is in quasi-steady state with updraught and downdraught co-existing for typically 1-3 hours, but on occasion for up to 12 hours. In some ways a supercell may be seen as the ultimate multicelled storm where the downdraught is so positioned that the daughter cell is co-located with the updraught of the parent cell.

An example of a hodograph associated with a supercell is shown in Fig 9a. Consider first the structure of a multicelled storm growing in this environment (Fig 9b). Each cell would have a steering level close to that of the mid-level winds, the downdraught would form on the Eastern flank, spread out, ultimately cut the inflow and possibly generate a new cell to the East.

However supercells are always observed to have a motion outside the envelope of the hodograph (see Fig 9a) - precisely why is not yet fully understood. In consequence storm relative winds are rather different, Fig 9c. In particular, mid-level winds are stronger and tend to have an influence on the position of the downdraught which, in supercells, forms on the opposite side of the updraught to the inflow. Maximum convergence occurs directly underneath the original updraught. A schematic diagram of the final steady state is shown in Fig 10.

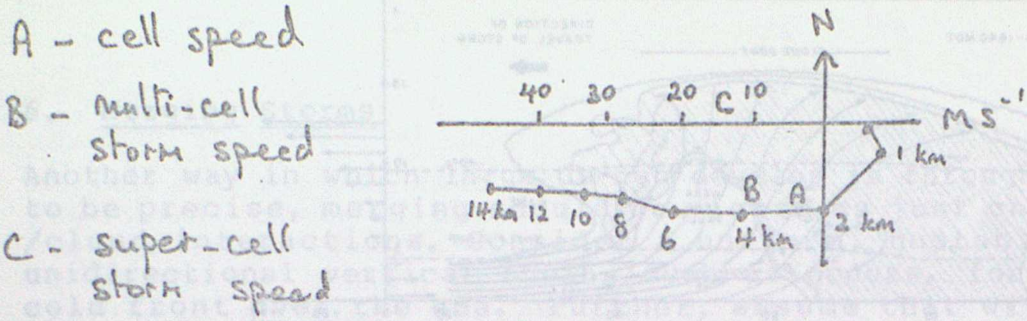


Figure 9a. Hodograph.

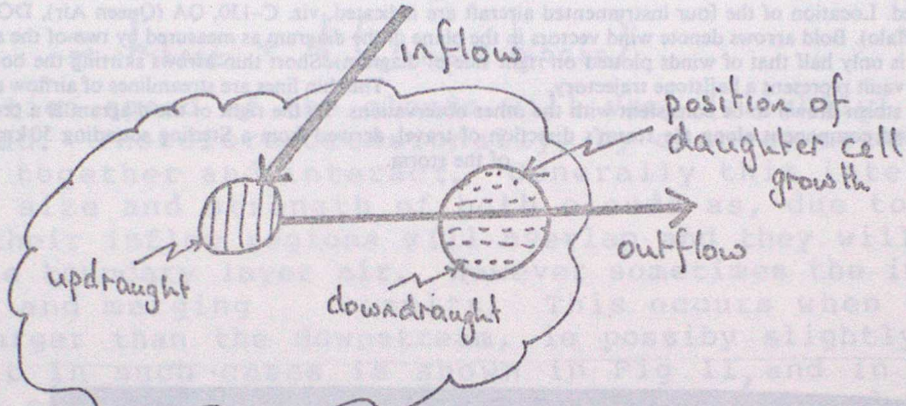


Figure 9b. Multicell storm growing in this environment.

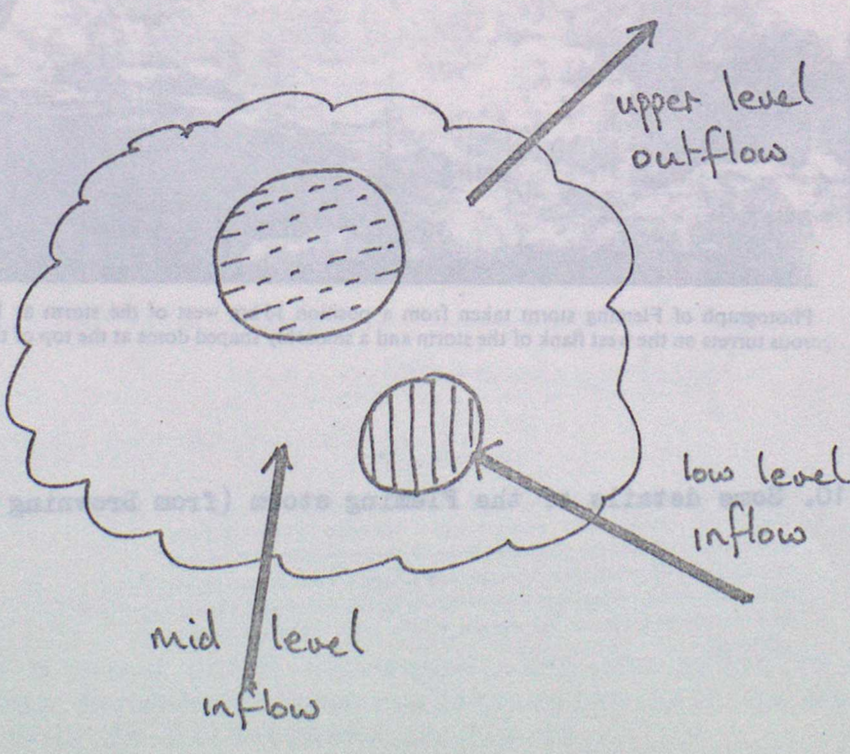
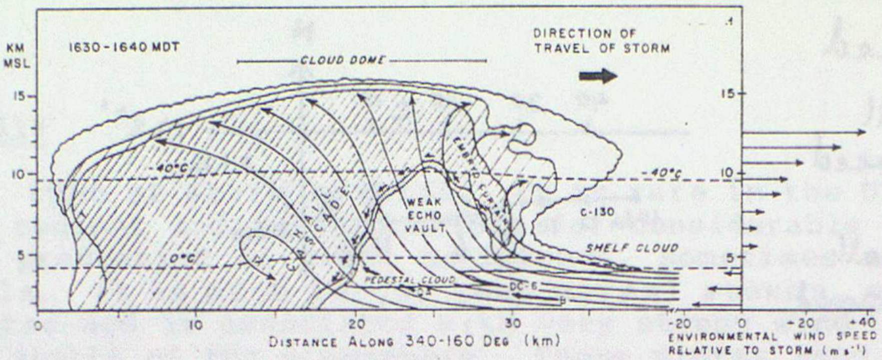
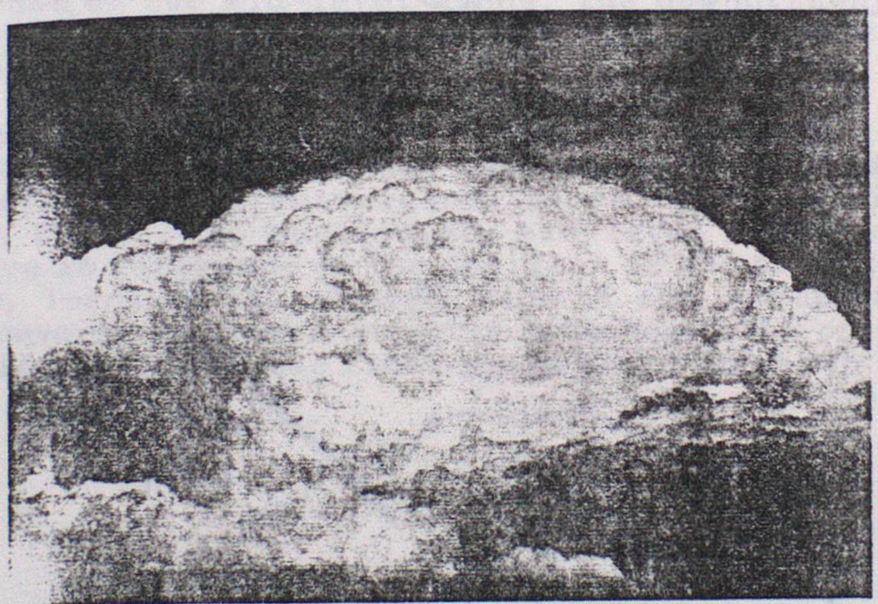


Figure 9c. Supercell.



Vertical section showing features of the visual cloud boundaries of the Fleming storm at 1630–1640 MDT superimposed on the pattern of radar echo derived. The section is oriented in the direction of travel of the storm. . Two levels of radar reflectivity are represented by different densities of hatched shading. Areas of cloud devoid of detectable echo are shown stippled. Location of the four instrumented aircraft are indicated, viz. C–130, QA (Queen Air), DC–6 and B (Buffalo). Bold arrows denote wind vectors in the plane of the diagram as measured by two of the aircraft (scale is only half that of winds plotted on right side of diagram). Short thin arrows skirting the boundary of the vault represent a hailstone trajectory. The thin lines are streamlines of airflow relative to the storm drawn to be consistent with the other observations. To the right of the diagram is a profile of the wind component along the storm's direction of travel, derived from a Sterling sounding 50 km south of the storm.



Photograph of Fleming storm taken from a position 35 km west of the storm at 1636 MDT showing vigorous turrets on the west flank of the storm and a smoothly shaped dome at the top of the storm.

Figure 10. Some details of the Fleming storm (from Browning and Foote, 1976)

6. Merging Storms

Another way in which large Cb can develop is through merging, although to be precise, merging should be viewed as just one aspect of cloud-/cloud interactions. Consider a uniform, unstable air mass having unidirectional vertical shear, such as occurs, for example, behind a cold front over the sea. Further, assume that within this air mass there develops a population of single celled clouds. At any given time clouds will exist at all stages of development ie they will have cloud tops over a range of heights, the younger the cloud the lower will be cloud top. Since the steering level

$$\approx h_{base} + \frac{1}{3}(h_{top} - h_{base})$$

the propagation speed of the clouds will vary (slightly) with the age of the cloud. Therefore occasionally, by chance, some clouds will move close together and interact. Generally this interaction will reduce the size and strength of both clouds as, due to their close proximity their inflow regions will overlap and they will compete for the unstable boundary layer air. However sometimes the interaction is beneficial and merging results. This occurs when the upstream cloud is larger than the downstream, ie possiby slightly older. The development in such cases is shown in Fig 11, and in Fig 12 is a comparison of the rainfall rate of a single cell and merger cloud growing in the same environment. Merging can enhance the rainfall rate significantly.

In consequence, even when the air mass is uniform and conditions are unsuitable for multicellular development, a few large cells can still develop. The occurrence of these is random depending on the relative position and development of neighbouring cells and so their position is not forecastable. However it is essential to remember that they can, and do, occur.

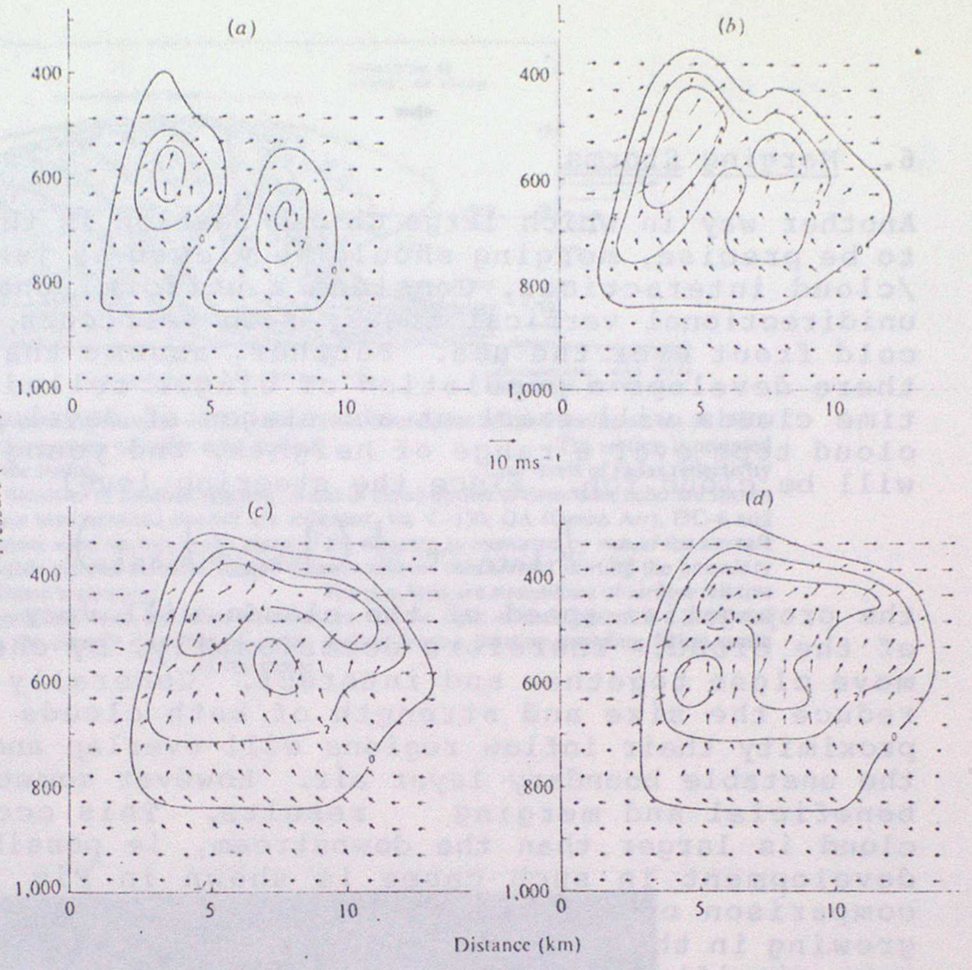


Fig. 11 Distribution of cloud content (water and ice) in g per kg within the clouds at a simulated time of: *a*, 12 min; *b*, 16 min; *c*, 20 min; *d*, 24 min. Arrows denote wind velocities in the plane of the diagram relative to the mean speed of the left hand cloud. The presence of rain (•) and hail (*) is also indicated.

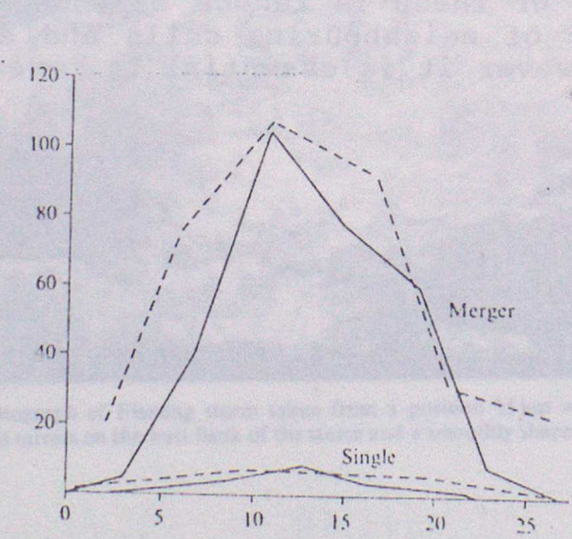


Fig. 12 Simulated and observed rainfall rates for merged and single cell clouds. The dashed curve labelled 'single' depicts observed rainfall rates that were typical of the smaller clouds and the dashed curve labelled 'merger' the rainfall rates of the largest cloud observed . The largest was chosen as the numerical simulation (solid line) shows the results of the most efficient merger achieved.

3. Precipitation

So far the storms have been discussed in terms of their dynamics. However in terms of forecasting it is equally important to know how much precipitation falls onto the ground. We have already seen that in terms of understanding the cloud the most useful frames of reference are axes moving with either cell or storm velocity. In consequence, for rainfall studies it is useful to consider the two aspects.

a) Rainfall from a given storm
b) Rainfall at the ground.

This allows us to study storms independent of their translation speed. Before the advent of radar this was difficult as the only observations were fixed relative to the ground. However radars allow the storm to be studied in their natural frame of reference, with point rainfall being readily available provided the cell speed is known.

For example, if a moveable rainguage, was kept beneath the centre of a storm, registered R mm rain then, for a storm of diameter D, moving at V m s^{-1}, having a life time L, the amount of rain falling at a given ground station would approximate to RD/LV (D/LV<1); "approximately" as the value of R would vary depending on whether the rainguage was beneath the centre of the storm in or near one of the edges. To illustrate this consider the Hampstead storm which the mean value of R was 100 mm, D ~ 10 km and L = 2 hrs, storm velocity in this case was zero and the surface rainfall was ~100 mm but if V had equalled say 6 ms^{-1} the rainfall at Hampstead would have been 25 mm, a reduction of 1/4, and the Hampstead storm would have caused little comment, being an average summer thunderstorm.

In summary, many very severe storms go unnoticed simply because they move quickly and give little rain at any given ground station.

a) Efficiency of precipitation

With the problems of surface rainfall and translation speed eliminated, it is now possible to consider the "efficiency" of a storm which may be defined as

$$\text{efficiency} = \frac{\text{rainfall at ground}}{\text{total cloud water condensed}}$$

This equation may be rewritten as

$$\text{efficiency} = \frac{\text{production of rain}}{\text{C.W.}} + \frac{\text{accretion}}{\text{C.W.}} - \frac{\text{evaporation}}{\text{C.W}}$$

where C.W = cloud water condensed, and the relative magnitude of these terms depends on the trajectory of rain as it falls to the ground. This is illustrated in Fig 14.

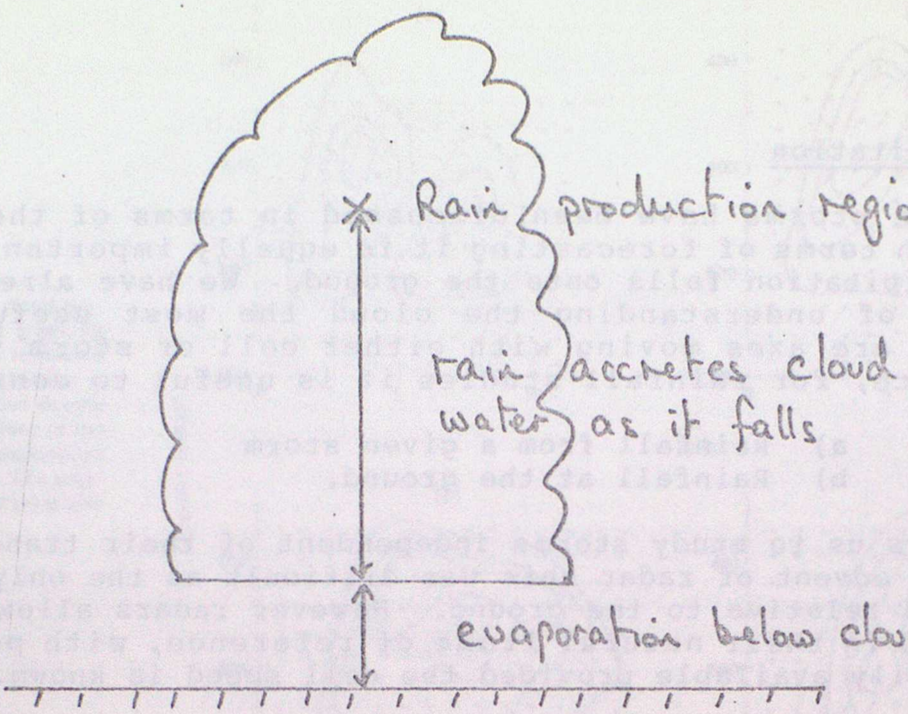
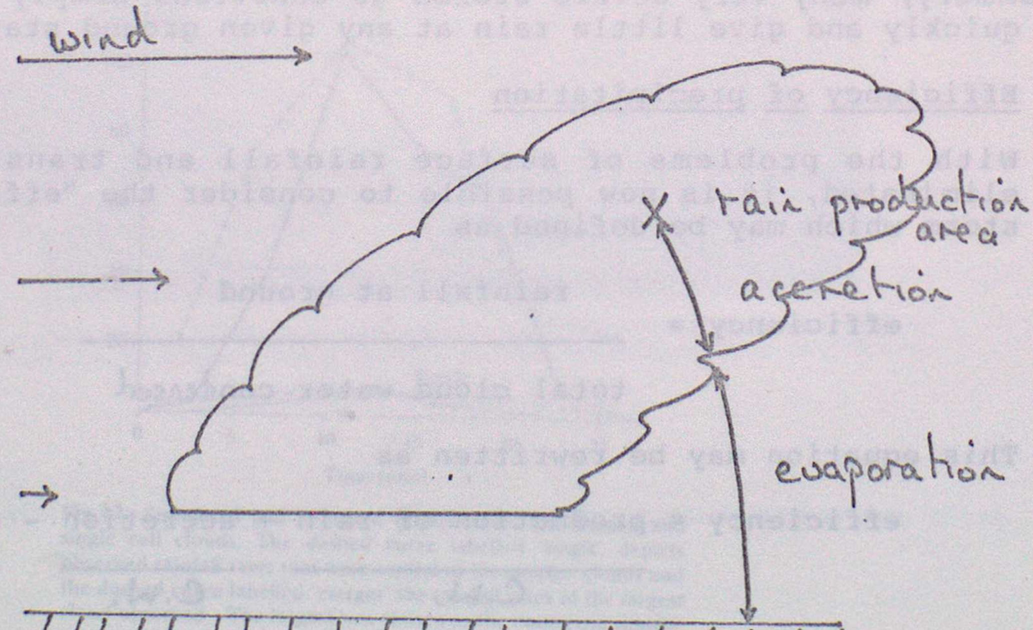


Figure 14a. Single cell cloud growing in no shear.



Single cell cloud growing in unidirectional shear.

In Fig 14a is shown the evolution of rain within a single cell growing in no shear. Precipitation is generated fairly high in the cloud, falls through cloud accreting cloud water until it reaches cloud base and then evaporates between cloud base and the ground. Compare this to the evolution within a single cell growing in unidirectional shear, Fig 14b. In the shear case the ratio of the

$$\frac{\text{accretive path length}}{\text{evaporate path length}}$$

is very much reduced and therefore the higher the shear the lower the amount of precipitation reaching the ground, (Fig 15) in spite of the fact that the lifetime of the cell slightly increases as the shear increases, since the updraught and downdraught become more widely separated. (Remember also that this is rainfall <u>relative</u> to the cell).

However, each cell of a multicellular storm has similar behaviour and therefore although each cell has fairly low efficiency, the total rainfall can be quite large as there are n cells, see dashed line Fig 15. (NB there is a lower limit of the vertical shear for multicells to form).

In both single cells and multicells efficiency remains low as each cell has a large remnant cloud which slowly dissipates after the rain has ceased. In contrast supercells achieve a steady state, they are the ultimate multicellular storm where the new cell is the old cell. As such, at the end of their life, there is only the remnant from one cell and in consequence they have high efficiency, up to 80%, although, as with multicells they have a minimum shear condition before they can develop.

Merging clouds produce high rainfall for a different reason. Here, the presence of the smaller cloud alters the path length ratio, see Fig 16. Precipitation from the upstream cloud falls through the smaller cloud and therefore the efficiency close to that of a single cell growing in no shear. See Fig 15.

The problem of quantifying Fig 15 is still the subject of active research.

References.

For furthur reading the following books are recommended

Meso-scale atmospheric circulations — B W Atkinson 1981 p 313, Academic press.

Clouds and Storms — F H Ludlam 1980 p 182, Penn. State Univ. Press.

Both have very comprehensive references.

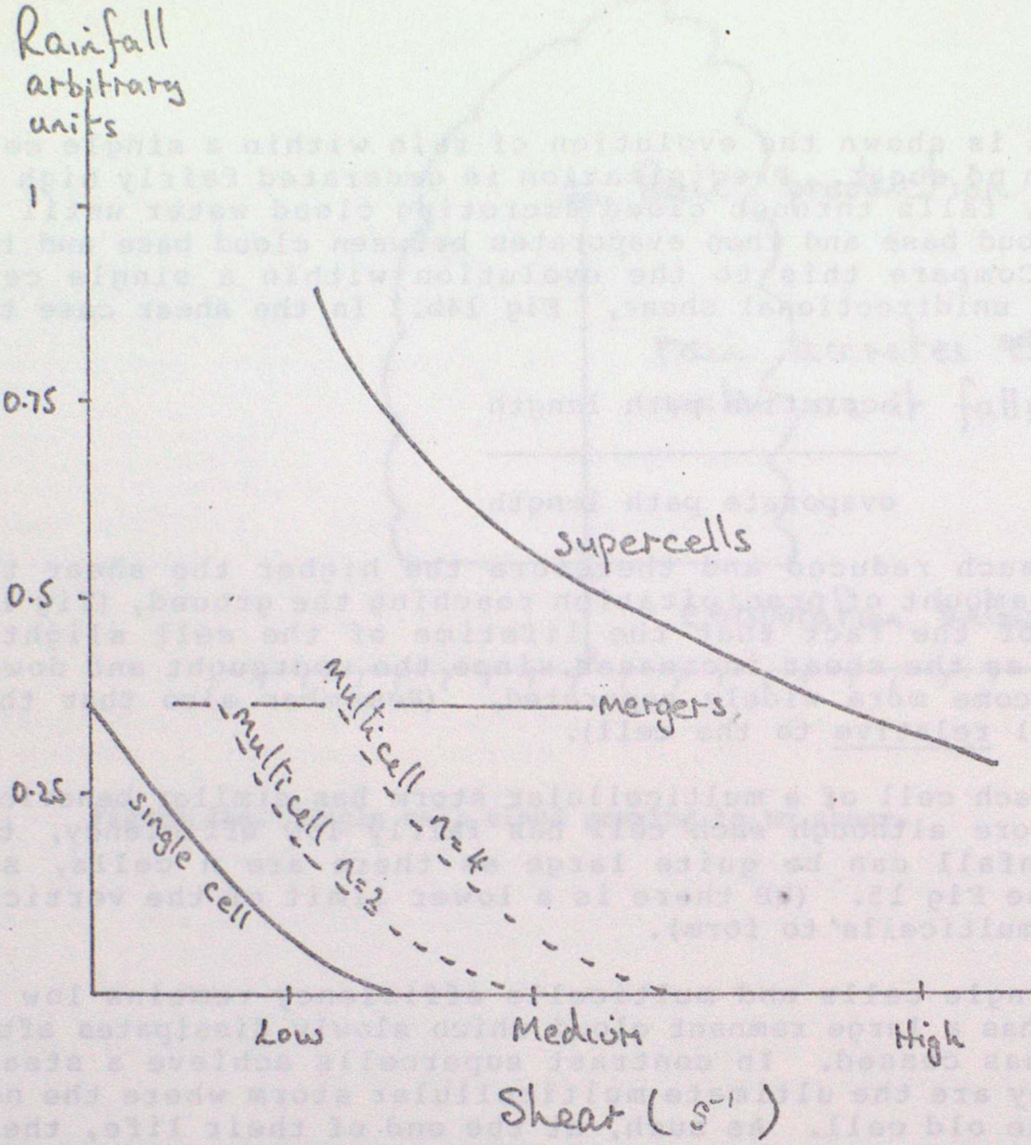


Figure 15. Sketch of rainfall efficiency of various types of cumulonimbus storm. Note, this is rainfall efficiency relative to the storm; rainfall at the ground also depends on the storm speed.

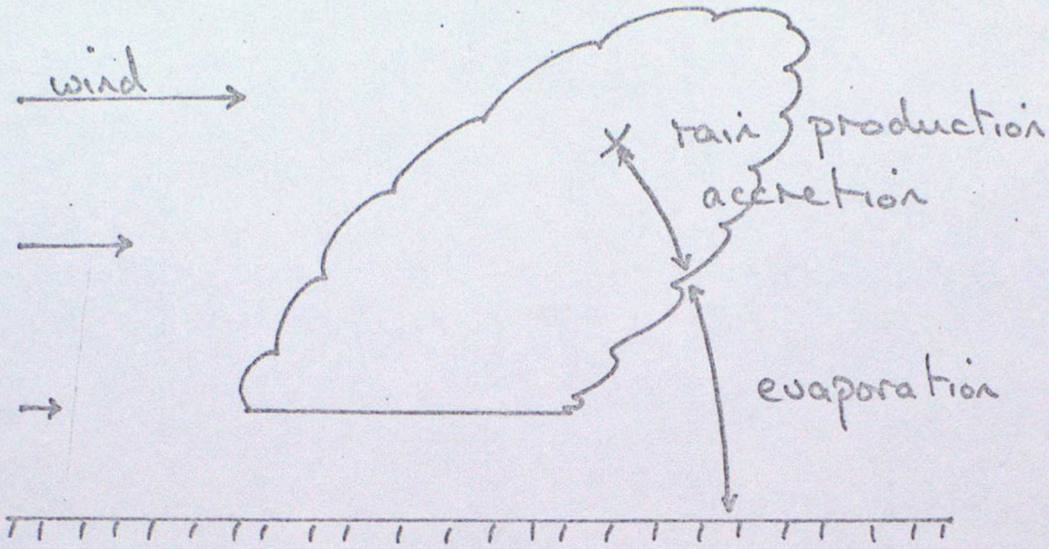


Figure 16a. Single cell cloud.

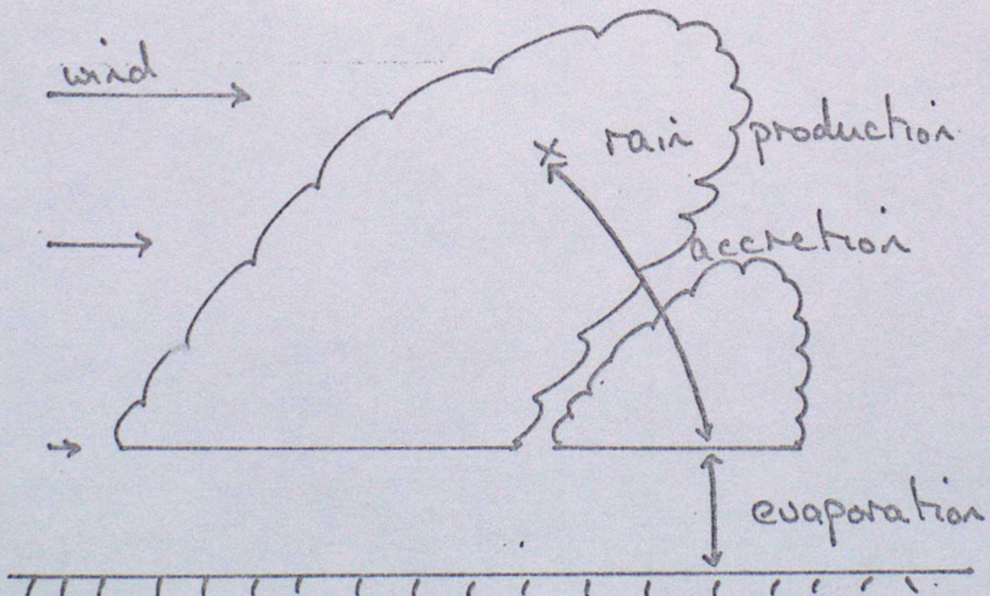


Figure 16b. Merging clouds.