Thunderstorms consist of one or more cells. A cell is characterized as a region where vertical motion is sufficiently strong to initiate deep convection. Each cell has a distinct life cycle consisting of three stages, i.e., a towering cumulus stage, a mature stage, and a dissipating stage.

A. *The towering cumulus stage is dominated by updrafts.*

1. Cell growth is associated with the formation of precipitation particles as moist air rises.
2. Water loading from precipitation initiates a downdraft during the latter part of the towering cumulus stage.

B. *Substantial updrafts and downdrafts coexist in the same cell during the mature stage.*

1. Precipitation usually (but not always!) reaches the surface during the mature stage.
2. Outflow associated with the downdraft spreads at the surface.

C. *The dissipating stage of a cell is dominated by downdrafts.*

1. Typically, cold air from the downdraft undercuts the updraft.
2. The source of warm moist air is cut off by the spreading cold pool and the cell dissipates.

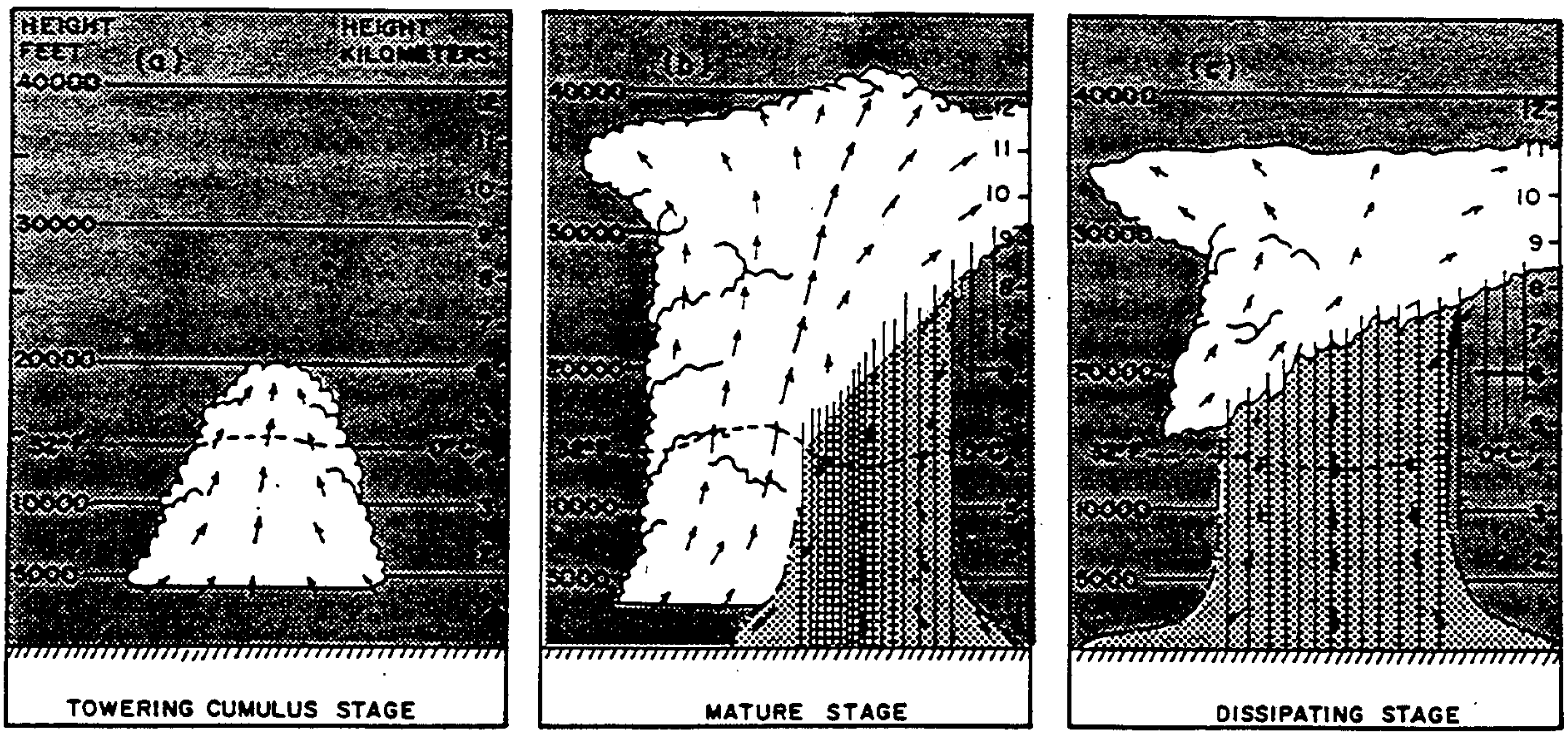
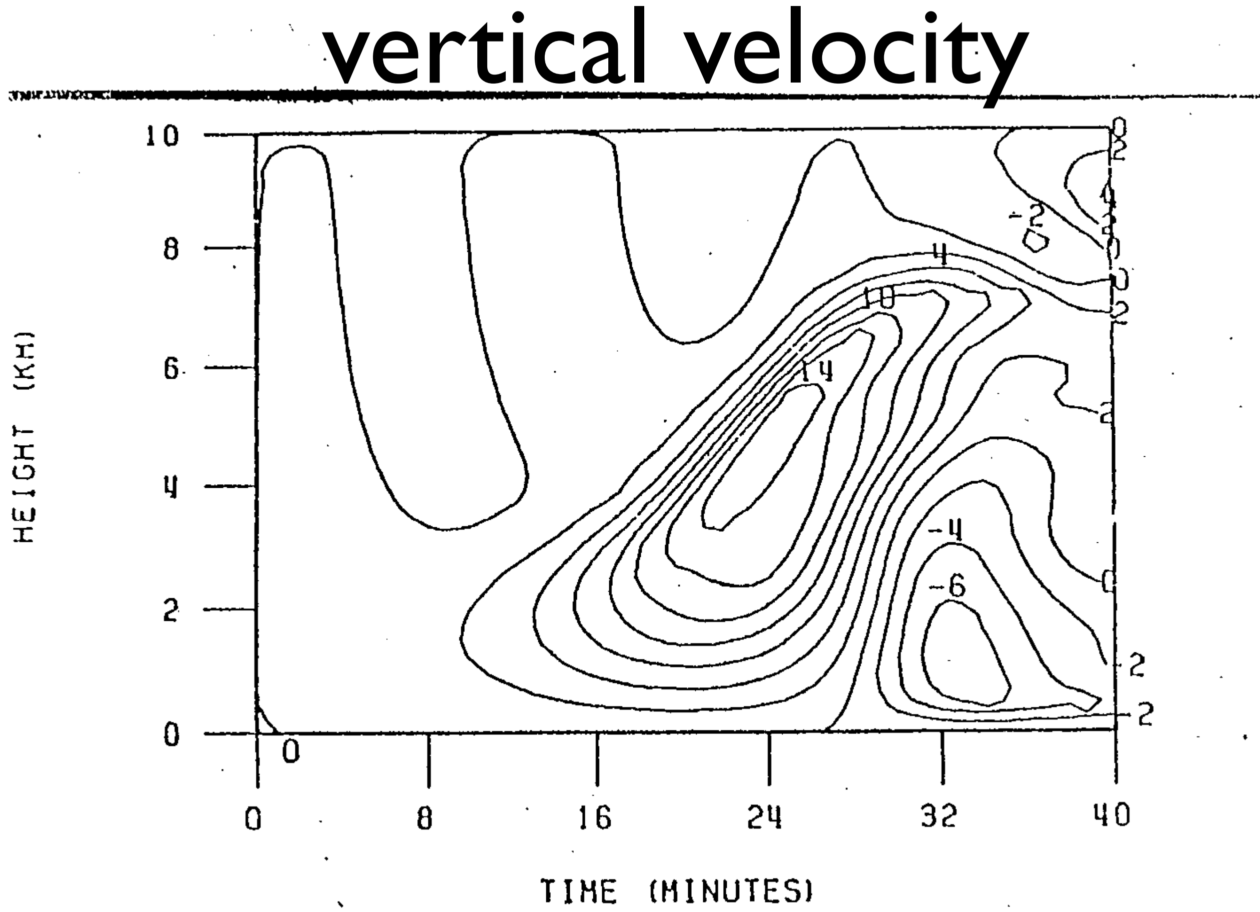
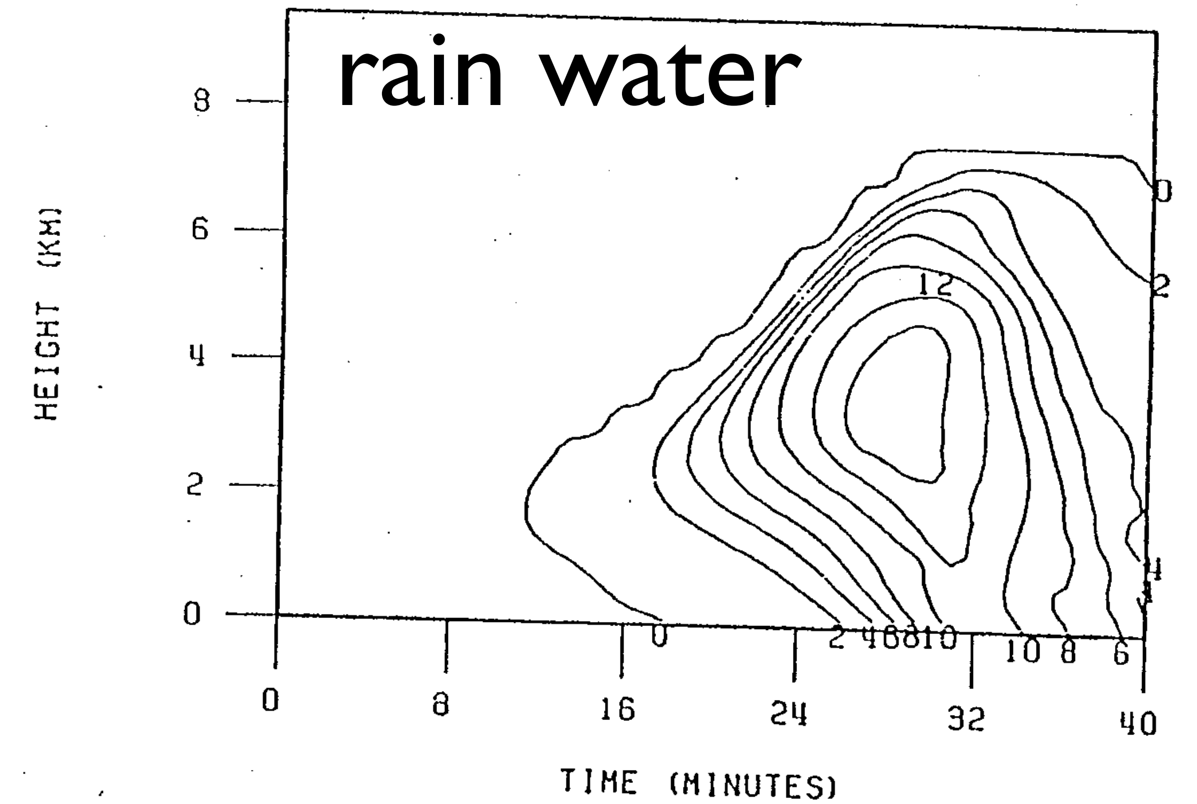
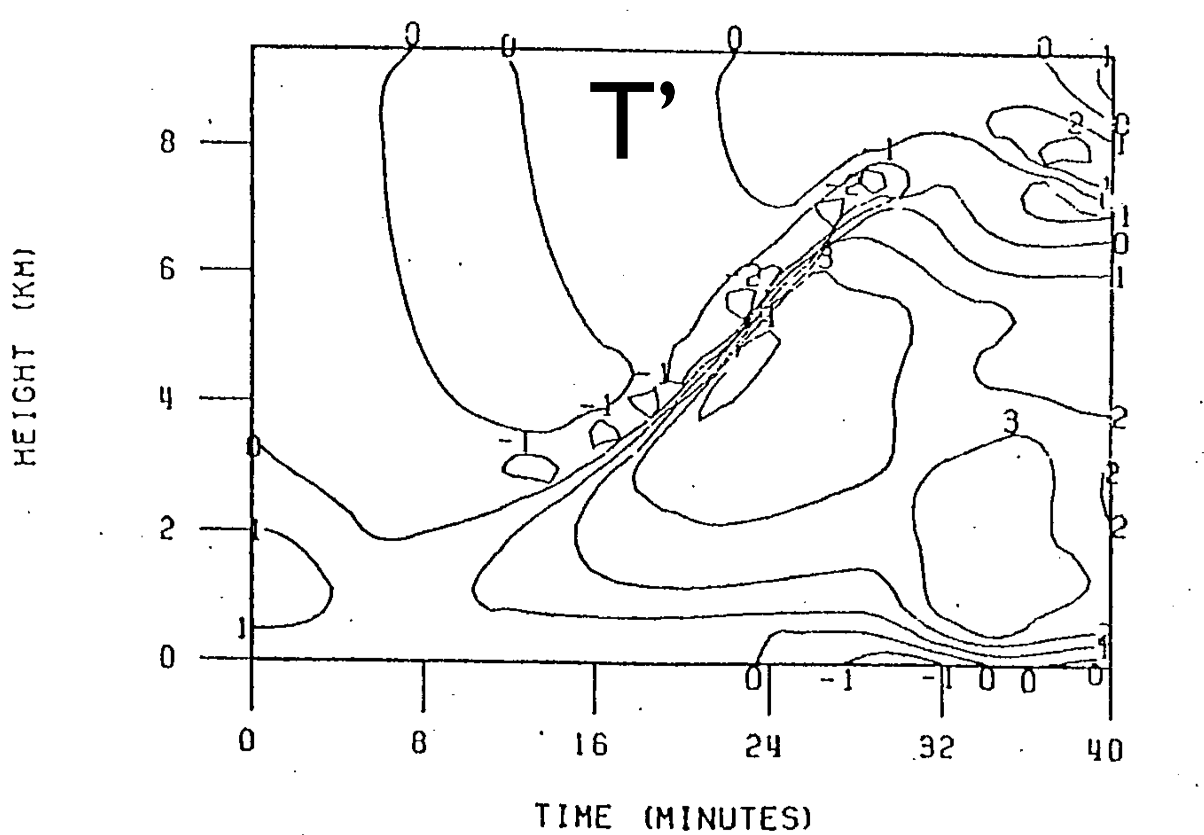


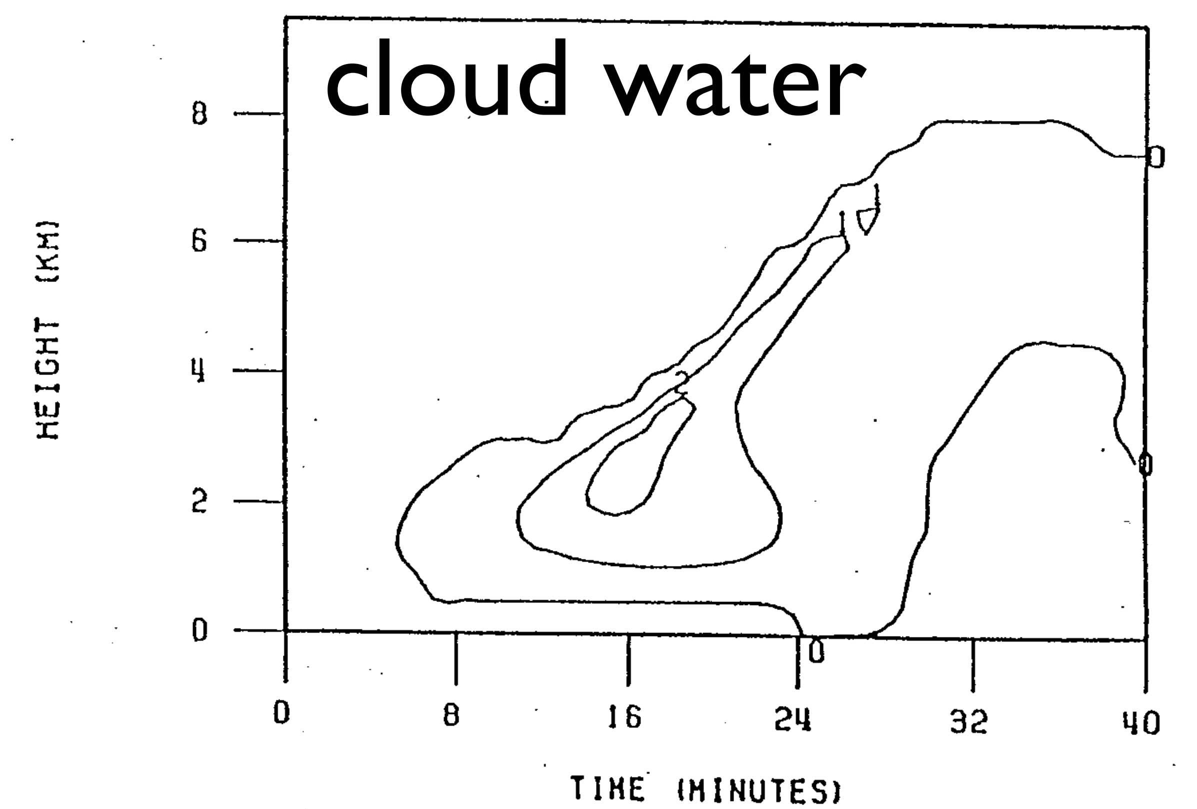
Fig. ***(a) The towering cumulus stage, (b) mature stage, and (c) dissipating stage of a short-lived convective cell (from Weisman and Klemp, 1986; adapted from Byers and Braham, 1949, by C. A. Doswell).***





Life cycle of a convective cell





Rainfall evaporation in thunderstorm downdraughts

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SUMMARY

It is explained that in air descending at speeds of some m sec^{-1} raindrop evaporation must proceed at a large subsaturation. Under some simplifying steady-state assumptions, the vertical temperature profiles in thunderstorm downdraughts of a range of speeds are estimated, and it is shown that the most favourable conditions for the production of strong draughts are a small raindrop size, and, especially, an intense rainfall rate and a general lapse-rate close to the dry adiabatic. The two latter conditions are often associated in severe storm situations, and the downdraught then arrives near the ground with a rather low relative humidity, and a temperature little lower than found generally at the same level.

THE DEPARTURE FROM THE SATURATED STATE IN UPDRAUGHTS AND IN DOWNDRAUGHTS

When condensation occurs in an updraught, the transfer of water from the vapour to the liquid state requires the presence of a supersaturation which increases with the rate of condensation, that is, with the speed of the updraught. However, the properties of the atmospheric aerosol particles which determine the microphysics of the condensation process are such that even in the strongest updraughts encountered, this supersaturation is small. Assuming that the supersaturation does not change rapidly during the rise of air, a concentration of about 1 g m^{-3} of water has to be condensed for each km of ascent.

If drops have mass m and number concentration N, then the LWC is

$$\mathrm{LWC} = Nm,$$

and the change of LWC with height z is

$$\frac{d\,\mathrm{LWC}}{dz} = N\frac{dm}{dz} = 1\ \mathrm{g\ m^{-3}\ km^{-1}} = 10^{-11}\ \mathrm{g\ cm^{-4}}$$

If the updraft speed is U, then

$$N\frac{dm}{dt} = N\frac{dm}{d}\frac{dz}{dt} = N\frac{dm}{dz}U = 10^{-11}U\ \mathrm{g\ cm^{-3}\ s^{-1}}.$$

The microphysics determines an average drop concentration N (cm^{-3}), radius r (cm) and mass m (g); the magnitude of the supersaturation s may be determined by using the familiar diffusion equation for the growth of individual drops :

$$dm/dt = C_v \, 4\pi r \, D\rho_v \, s, \qquad (1)$$

whence (previous slide)

$$10^{-11} \, U = N dm/dt = N \, C_v \, 4\pi r \, D\rho_v \, s \qquad (2)$$

where

D is the coefficient of diffusion of water vapour in air ($\approx 0{\cdot}4$ cm^2 sec^{-1})

C_v is an empirical ventilation coefficient needed to take into account the motion of the drop through the air,

ρ_v is the saturated vapour density ($\approx 10^{-5}$ g cm^{-3}), and

U is the updraught speed in cm sec^{-1}.

In this equation ($\rho_v \, s$) is strictly the effective vapour density difference which governs the diffusion, which is smaller by a factor of 2 to 3 than the value given by using the actual supersaturation s, because of the rise of droplet surface temperature above the vapour temperature during the liberation of the latent heat of condensation. This may be taken into account in an order of magnitude calculation by using an adjusted value of $0{\cdot}1$ cm^2 sec^{-1} for the diffusion coefficient D. Accordingly we have numerically

$$s \approx 10^{-6} \, U/N \, r \, C_v \, . \qquad (3)$$

Solve (2) for s:

$$s = \frac{10^{-11}U}{NC_v 4\pi r D\rho_v}.$$

Use $D = 0.1$ cm^2 s^{-1}, $\rho_v = 10^{-5}$ g cm^{-3}, and $4\pi \approx 10$:

$$s \approx 10^{-11}\frac{U}{NC_v(10)r(0.1)10^{-5}} = 10^{-6}\frac{U}{NC_v r}.$$

$$s \approx 10^{-6}\,U/N\,r\,C_v\,. \qquad . \qquad . \qquad . \qquad . \qquad (3)$$

Now the properties of atmospheric condensation nuclei are such that in an updraught of 1 m sec^{-1} or more, N is about 10^3 and r therefore about 10^{-3} cm, and C_v can be put equal to unity. Correspondingly, even in an updraught of 10 m sec^{-1}, s is of order only 0·1 per cent. Thus it arises that for all practical purposes updraughts can be assumed to be just saturated, and the state of the air within them compared directly with an adiabatic reference process without specification of the supersaturation.

For $N = 10^3$ cm^{-4}, $r = 10^{-3}$ cm, and $C_v = 1$,

$$s \approx 10^{-6}\frac{U}{10^3(1)10^{-3}} = 10^{-6}\,U$$

so that even for $U = 10$ m/s $= U = 10^3$ cm/s ,

$$s \approx 10^{-6}\,10^3 = 10^{-3}.$$

Some downdraughts occur inside clouds, but the deeper, and more intense and signicant are likely to develop when precipitation falls from inclined or horizontally-spreading clouds into the air of the middle troposphere, which generally has a minimum wet-bulb potential temperature (as for example in Fig. 3) and so is the most suitable for sustaining a cold descending current. In such a downdraught the evaporation of rain can be estimated with an empirically determined ventilation coefficient, which we shall use subsequently, found by Frössling (1938) :

$$C_v = 1 + 0{\cdot}229\,(Re)^{\frac{1}{2}} \qquad . \qquad . \qquad . \qquad . \qquad (4)$$

where (Re) is the Reynolds number, which reaches a value of about 10^3 or rather more for raindrops of millimetric size. Accordingly, for such drops a ventilation coefficient of magnitude 10 can be used. Eq. (3) then becomes numerically

$$s' \approx 10^{-7}\, U'/Nr \qquad . \qquad . \qquad . \qquad . \qquad . \qquad (5)$$

where s' is a persistent subsaturation in a downdraught of speed U' cm $\sec^{-1}$. Now with $r = 10^{-1}$ cm and even in an intense rain of about 4 in. hr^{-1}, for which $N \approx 10^{-3}\ \mathrm{cm}^{-3}$, and with U' only 1 m $\sec^{-1}$, from Eq. (5) $s' = 10^{-1}$, that is, the relative humidity is only 90 per cent. Clearly downdraughts are likely to be characterized by rather low humidities; it cannot be assumed that within them the changes of state of descending air follow closely any simple reference process. The speed attained by a downdraught and the degree to which apparently available potential energy is actually converted into kinetic energy, must depend markedly not only on the particular atmospheric stratification, but also upon the size-distribution and concentration of the raindrops.

next slide calculates s for heavy rain

For heavy rain: $N = 10^{-3}$ cm^{-4}, $r = 10^{-1}$ cm, and $C_v = 10$,

$$s \approx 10^{-6} \frac{U}{10^{-3}(10)10^{-1}} = 10^{-3}\, U$$

so that even for $U = 1$ m/s $= U = 10^2$ cm/s ,

$$s \approx 10^{-3}\, 10^2 = 10^{-1}.$$

Rain Evaporation Rate

$$r\,dr/dt = C_v\,D\,\Delta\,\rho \qquad (15)$$

where $\Delta\,\rho$ is the difference between the vapour densities in the surroundings of the drops and at their surfaces. A more convenient form of Eq. (15) is obtained if we substitute $dz/(V + w)$ for (dt) and $-\,10^3\,dp/g\,\rho_a$ for (dz), where p is pressure in mb. Then the change of radius of a falling drop with height is given by :

$$dr/dp = (10^3\,C_v\,D\,\Delta\,x)/g\,(V + w)\,r \qquad (16)$$

where $(V + w)$ is the speed of the drop relative to the ground and $\Delta\,x$ is the difference of the mixing ratios in the downdraught air and at the surfaces of the raindrops.

Thus the evaporation of a population of raindrops, introduced into the downdraught at the 500 mb level with an initial radius r_0 and concentration N_0 (determining the rainfall intensity), can be found from the set of three equations (12), (14) and (16). Their simultaneous solution gives the relationship between the raindrop size and concentration and the magnitude and temperature of the downdraught.

Eq. (12) expresses conservation of droplet number mixing ratio.

Eq. (14) expresses conservation of total water (water vapor and liquid water) mixing ratio.

The equations in section 2 have been solved to give the temperature distribution between 500 and 800 mb in weak, moderate and strong downdraughts when the initial values of N and r at the 500 mb level correspond to rainfall of intensity described as moderate (m; 5 mm hr^{-1}), heavy (H; 5 cm hr^{-1}), such as is typical of thunderstorms, and intense (I; 25 cm hr^{-1}), which is sometimes attained in severe local storms. In order to illustrate the influence of raindrop size these rains were considered to be composed of uniform droplets of initial radius either 0·5 mm (about the minimum size of drops described as rain) or 2 mm (about the largest encountered size of raindrop). In the eighteen calculations the wet-bulb potential temperature was chosen to be 18°C, a value typical of the middle troposphere in the thundery weather of middle latitudes, with an air temperature at 500 mb of − 12·2°C, corresponding to a state of saturation. Two additional calculations were made (for a strong downdraught with I, $r_0 = 2$ mm, and for a moderate downdraught with m, $r_0 = 0{\cdot}5$ mm) with the same 500 mb temperature, but a very low relative humidity, corresponding to a wet-bulb potential temperature of 15°C (Fig. 2). These showed, as might be anticipated, that the results are not particularly sensitive to the initial relative humidity at the 500 mb level. The calculated vertical temperature distributions are illustrated in Fig. 1.

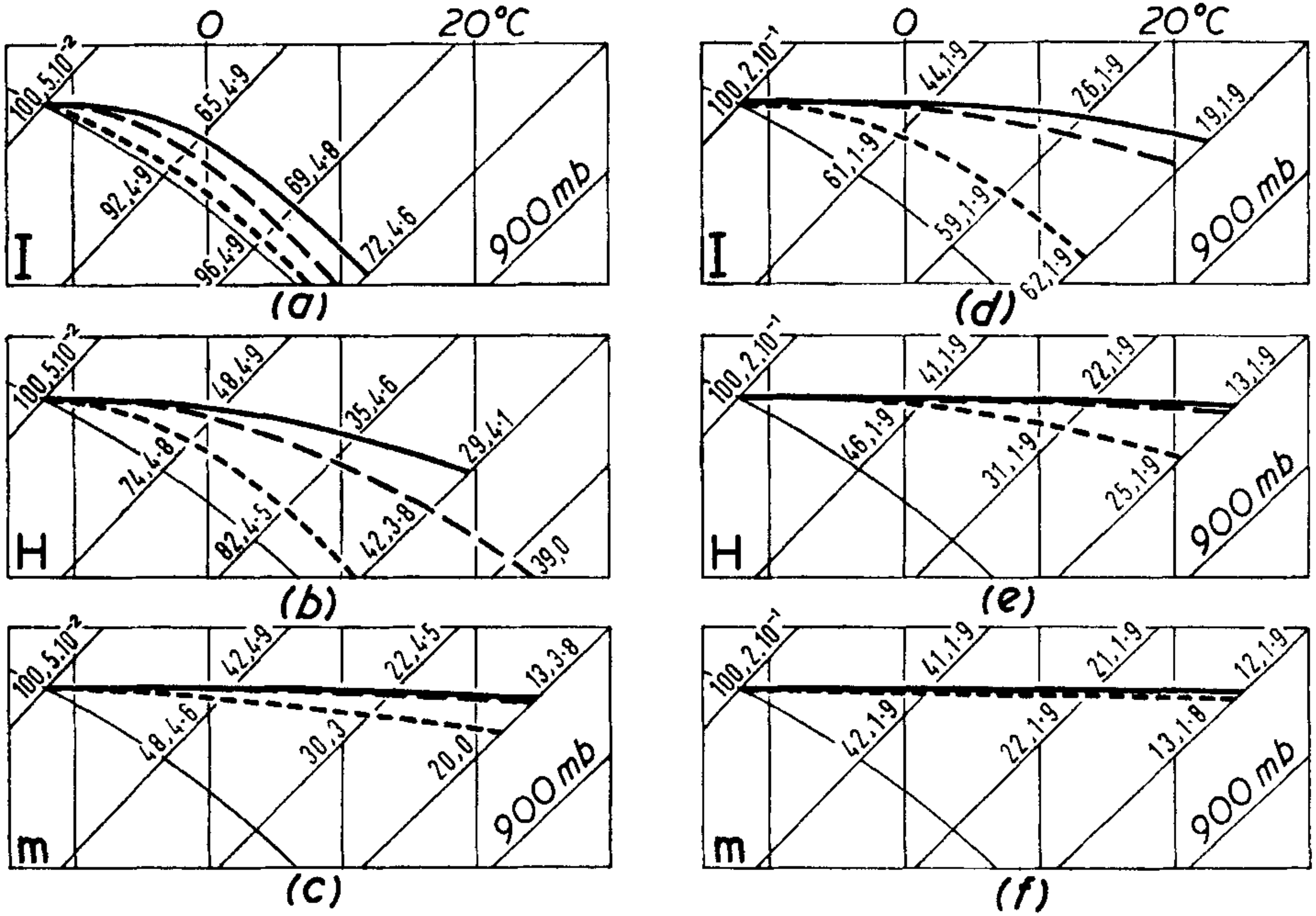
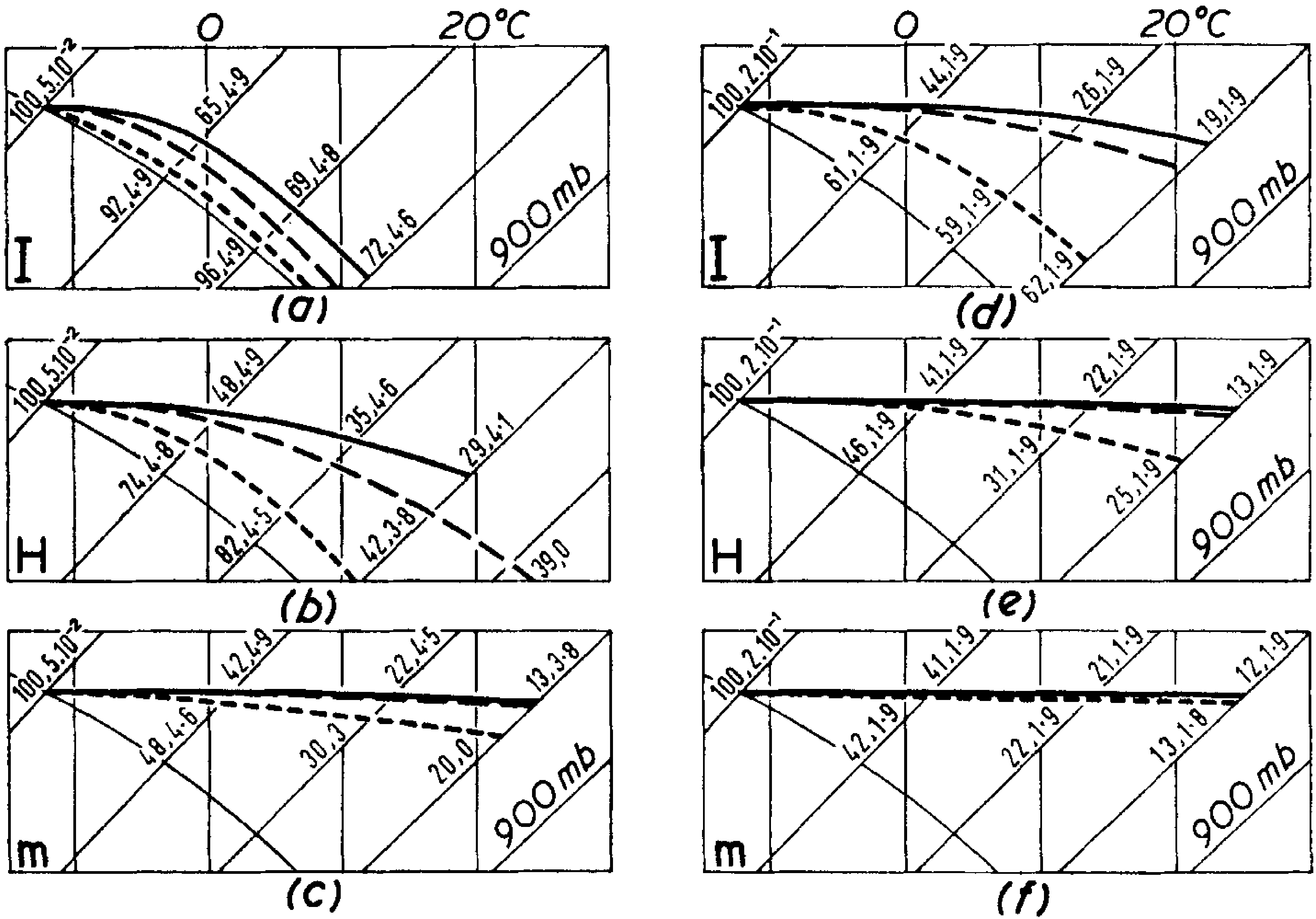


Figure 1. Temperature and humidity in *strong* (continuous lines), *moderate* (dashed lines) and *weak* (pecked lines) downdraughts, produced by the evaporation of *intense* (I), *heavy* (H) and *moderate* (m) rains, of uniform initial drop size 0·5 (left) and 2 mm (right). The rainfall intensities at 500 mb are respectively 250, 50 and 5 mm hr^{-1}. The temperature profiles are drawn upon sections of a tephigram limited by the horizontal dry adiabatics corresponding to potential temperatures of 30 and 50°C. The isobars are drawn at intervals of 100 mb. The lowermost curve is the saturated pseudo-adiabatic corresponding to the wet-bulb potential temperature of 18°C, which is preserved in the downdraughts. At each 100 mb level the pairs of figures entered beside the profiles for the strong and the weak downdraughts give the relative humidity in per cent, followed by the drop radius in units of 10^{-2} cm in diagrams (*a*) to (*c*), and in units of mm in diagrams (*d*) to (*f*). In all cases the downdraught air is assumed to be saturated at a temperature of − 12·2°C at the 500 mb level.

The vertical air speeds are respectively about 20 (24·8 to 16·8) m sec^{-1}, about 10 (12·4 to 8·4) m sec^{-1} and about 2 (2·5 to 1·7) m sec^{-1} in the strong, moderate and weak downdraughts. A more accurate specification of the results is given in Table 1.



0
20°C
100,5.10⁻²
65,4·9
69,4·8
72,4·6
92,4·9
96,4·9
900 mb
I
(a)
100,5.10⁻²
48,4·9
35,4·6
29,4·1
74,4·8
82,4·5
42,3·8
39,0
900 mb
H
(b)
100,5.10⁻²
42,4·9
22,4·5
13,3·8
48,4·6
30,3
20,0
900 mb
m
(c)
0
20°C
100,2.10⁻¹
44,1·9
26,1·9
19,1·9
61,1·9
59,1·9
62,1·9
900 mb
I
(d)
100,2.10⁻¹
41,1·9
22,1·9
13,1·9
46,1·9
31,1·9
25,1·9
900 mb
H
(e)
100,2.10⁻¹
41,1·9
21,1·9
12,1·9
42,1·9
22,1·9
13,1·8
900 mb
m
(f)

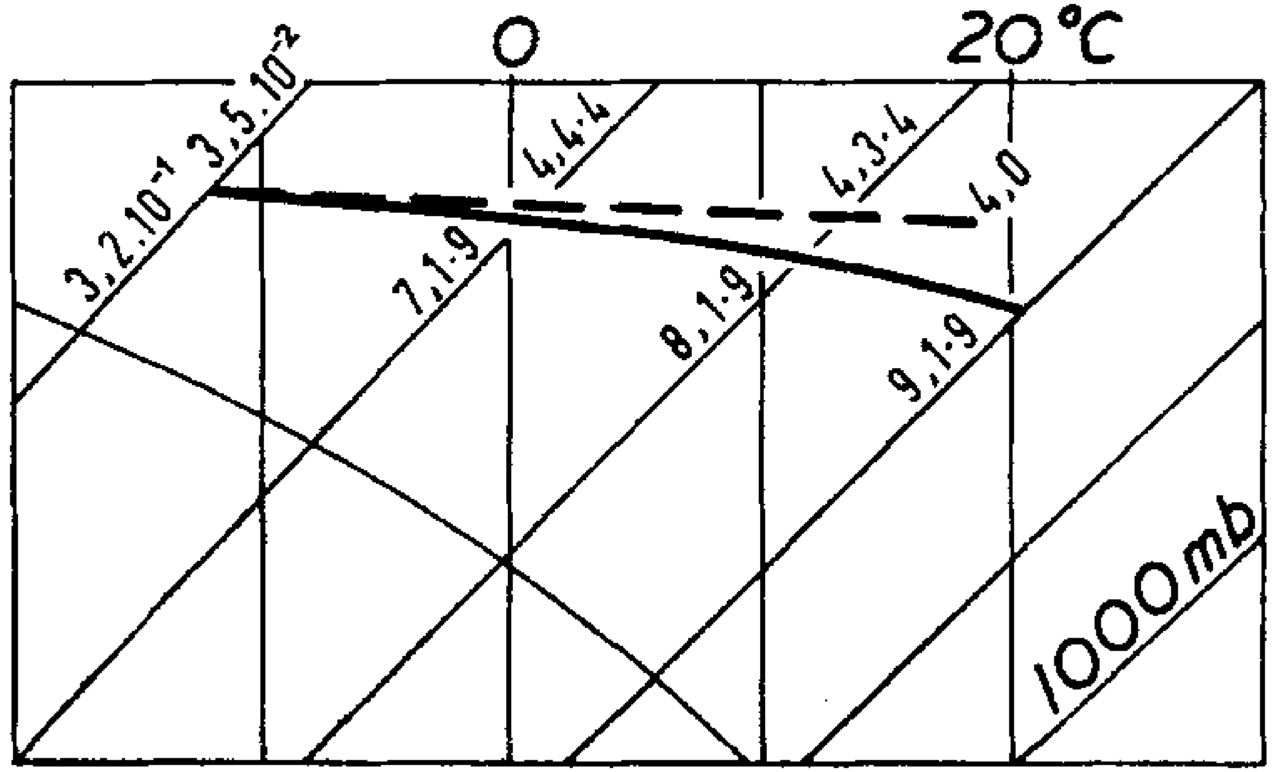


Figure 2. Temperature and humidity in downdraughts produced in initially very dry air by the evaporation of rain; dashed line : moderate downdraught (vertical speed about 10 m sec^{-1}, moderate rain of initial drop size 0·5 mm; continuous line: strong downdraught (speed about 20 m sec^{-1}), intense rain of initial drop size 2 mm. The two temperature profiles are drawn upon sections of a tephigram limited by the horizontal dry adiabatics corresponding to potential temperatures of 20 and 50°C. The isobars are drawn at intervals of 100 mb. The lowermost curve is the saturated pseudo-adiabatic corresponding to the wet-bulb potential temperature of 15°C, which is preserved in the downdraughts. At each 100 mb level the pairs of figures entered beside the profiles give the relative humidity in per cent, followed by the drop radius in units of 10^{-2} cm (upper curve) and in units of mm (lower curve). In both cases the downdraught of air is assumed to be very dry (R.H. = 3 per cent) at a temperature of − 12·2°C at the 500 mb level. The conditions at lower levels are also specified in Table 2, and can be compared with those in Table 1 and Figs. 1 (c) and (d).

4. Conclusion

The evaporation of populations of raindrops in the downdraughts characteristic of cumulonimbus convection has been estimated under some simplifying assumptions. The influence of raindrop size and rainfall intensity upon downdraught speed has been considered and it is concluded that the raindrop size has an appreciable influence, the evaporation proceeding more efficiently the smaller the size. However, the parameters of dominating importance are the rainfall intensity and the general lapse-rate. The most favourable conditions for the production of *strong* downdraughts are a general lapse-rate close to the dry-adiabatic, from the ground into the middle troposphere, and a great intensity of rainfall, in which the weight of the rain, as well as the enhanced cooling by evaporation, provides an important buoyancy force. These conditions are often associated, for example in the severe local storm situations of the North American mid-western States.

(4) *Damaging Wind*

A. *Microbursts and Macrobursts*

1. Extreme convective wind events tend to occur with relatively concentrated downdrafts and subsequently intense outflows. These ***"downburst" events are known as microbursts (macro-bursts); they are defined as having a horizontal dimension smaller (larger) than 2.5 nm.***

2. Concentrated downdrafts (and associated "outbursts" of strong horizontal winds) occur within regions of weaker overall descent. As a result, ***downbursts may be embedded within the outflow of an isolated storm or system of storms.***

3. ***Multiple downbursts may combine to form a larger outflow with a gust front at its leading edge.*** Strong surface outflows can occur for a wide variety of storms and systems evolving within diverse environments. Thus, ***downbursts have been associated with supercell storms, severe pulse storms, organized non-supercell storms, and MCS's (most notably the "bow echo" class of MCS).***

4. ***Downbursts can occur with very benign-appearing storms.*** In fact, the updraft may be so weak in some downburst-producing storms that thunder is absent!

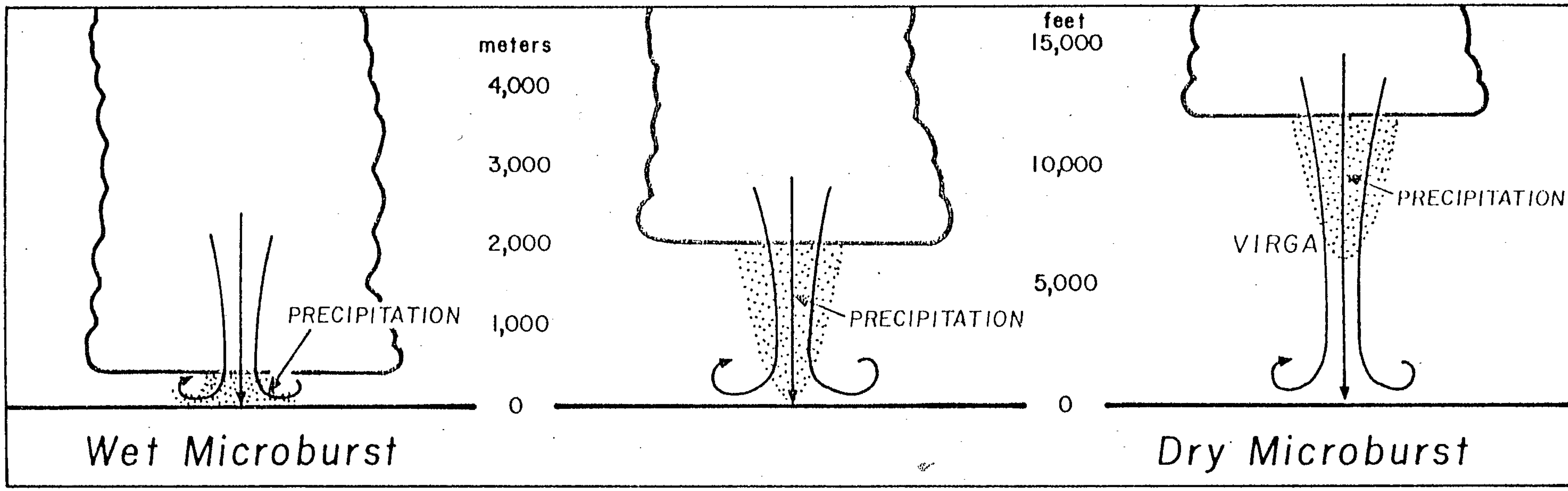


Fig. 5.1 Schematic views of wet and dry microbursts. Wet microbursts are expected to occur in the wet regions of the world, while dry microbursts are commonly seen in the dry regions with high bases of convective clouds.

B. Dry Microbursts

1. ***Dry microbursts (defined as a microburst accompanied by little or no rain during the period of high wind) typically develop from shallow, high-based cumulonimbus clouds.*** In general, this class of wind event occurs as a "pulse" and is associated with weak vertical wind shear and little if any synoptic scale forcing.

2. ***Environmental factors conducive to the formation of dry microbursts include a deep, dry adiabatic sub-cloud layer, with sufficient moisture aloft to sustain a downdraft all the way to the surface in the face of strong evaporation and adiabatic compressional heating.***

3. ***Dry microbursts are associated with very high LFCs and only marginal instability for updrafts. Thus the convection is usually weak and without electrical activity.*** This represents an extreme example of how downdraft and updraft instabilities differ. It would not be productive to apply ordinary convective stability indices to such an event since they are not designed to measure downdraft instability.

4. Although dry microbursts are observed most often over the high plains east of the Rocky mountains, they have also been observed in other parts of the country.

5. The downdraft associated with a dry microburst is initiated by precipitation loading within the elevated cloud layer. ***Strong surface winds result from negative buoyancy generated by evaporation, melting, and sublimation of precipitation below cloud base.***

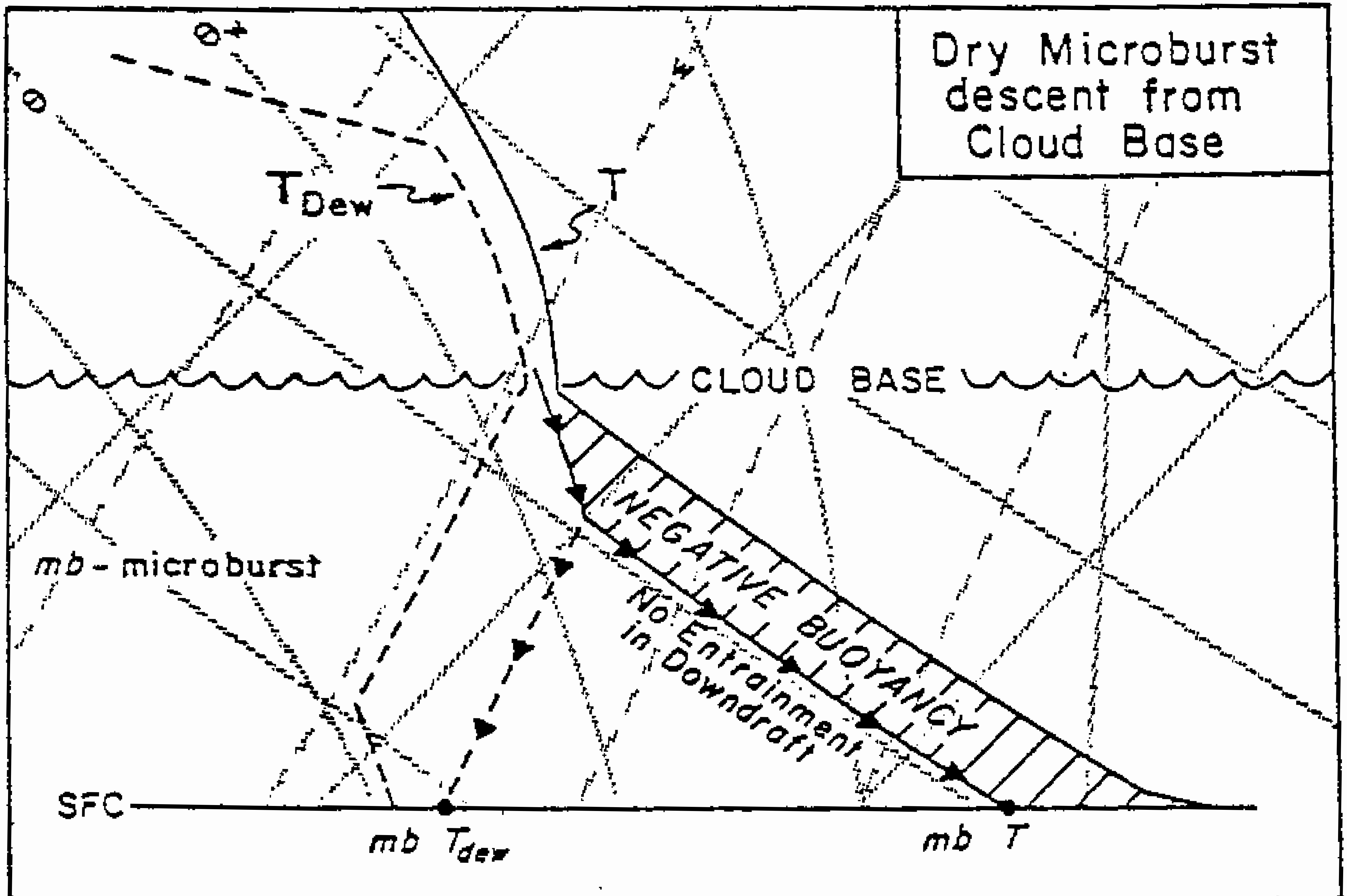


FIG. 10. Model of the thermodynamic descent of a dry microburst from cloud base. Surface temperature and dew-point temperature within the microburst are determined from PAM data. No entrainment into the downdraft is assumed.

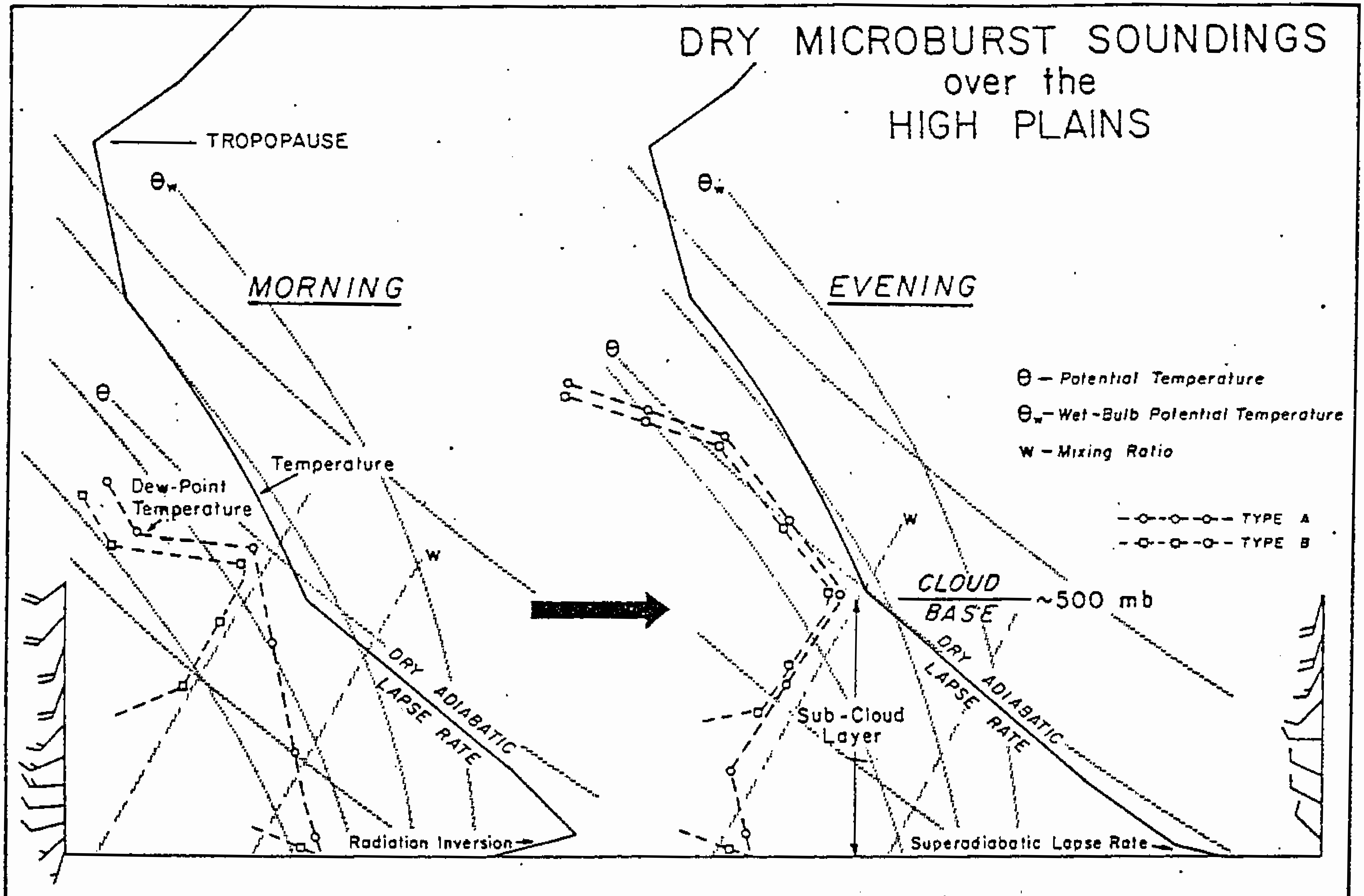


FIG. 8. Model of the characteristics of the morning and evening soundings favorable for dry-microburst activity over the High Plains.

C. *Wet microbursts*

1. ***Wet microbursts are often accompanied by heavy rain; they represent the dominant form of downburst for humid regions, such as the southeastern United States***. Environments associated with these wind events generally exhibit a nearly saturated boundary layer and a relatively shallow subcloud layer. Since wet microbursts are associated with heavy precipitation, high radar reflectivities are generally present.

2. ***Wet microburst environments are generally characterized by weak synoptic scale forcing and strong instability***. Precursor soundings typically indicate low LFCs, with little or no capping inversion, and a source of low humidity air aloft. Diurnal heating usually produces a dry adiabatic layer over the lowest 1.5 km during the afternoon.

3. ***Wet microbursts are driven primarily by melting and further enhanced by evaporative cooling, both in cloud and below cloud base.*** Entrainment and convergence of low theta-e environmental air near the level of minimum theta-e may also enhance the downdraft potential. Since wet-microbursts are associated with heavy precipitation, water loading is a significant contributor to the initiation and maintenance of the downdraft.

4. ***The lapse in theta-e (from the surface to some minimum value aloft) has been correlated with the production (or absence) of wet microbursts***. A lapse of theta-e greater than 20°K has been associated with microburst-producing storms; however, environments characterized by a lapse of theta-e less than 13° weren't associated with microbursts.

5. ***Wet microbursts are usually associated with severe pulse storms.*** However, ***microbursts also occur with well organized non-supercell and supercell storms***, which are capable of supporting repetitive episodes of damaging outflow winds.

Summary of Downdraft Processes

Conditions that contribute to an intense downdraft include:

(1) ***A deep subcloud layer with an environmental lapse rate close to dry adiabatic.***

(2) ***High precipitation content.***

(3) ***Microphysical conditions that produce a high concentration of small precipitation particles.***

A. Descending air continues to warm adiabatically under these conditions, but much of the heat produced is used to evaporate droplets or to melt ice particles. The net result is an accelerating downdraft.

B. If the environmental lapse rate within a deep subcloud layer is close to dry adiabatic (as it is with many dry microburst environments), evaporatively driven intense downdrafts can occur with practically any rainwater mixing ratio, enhanced possibly by melting below cloud base.

C. However, as the stability of the lapse rate increases and/or as the subloud layer becomes more shallow (as it does with wet many microburst environments), higher precipitation contents are needed to drive an intense downdraft. In the latter case, water loading and melting can become very important contributors to the production of an intense downdraft.