THUNDERSTORM DYNAMICS IN A SCALE INTERACTION

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ABSTRACT

This paper presents a review of thunderstorm dynamics through the use of the vorticity equation with a focus on the development of rotation, an indicator of storm severity. The processes involved in the scale interaction are basically the tilting of vortex tubes while the divergence (or ballerina) term, being highly non-linear, provides intensification of incipient vorticity. A brief account of numerically simulated storms in Brazil with well defined wind shear in the storm environment is presented.

1. INTRODUCTION

The scales of atmospheric motions have been defined by several authors but the more commonly used is the one defined by Orlanski (1975), reproduced in Fig. 1. Spatial and temporal scales are unified through a natural feature of atmospheric systems of being larger/smaller when long/short lived. Thunderstorms, or just storms for short, are atmospheric systems that last from about 30 minutes to a few hours with horizontal dimensions from 2 to 20 km (approximate horizontal diameter) and correspond to the meso-γ scale. Their existence is due to particular atmospheric conditions defined by larger scale systems, usually at the meso-α scale, and their evolution and lifetime, and their eventual growth into the meso-β scale, depend and impact on the smaller scale processes, typically of the micro scale. At the micro scale, the microphysical features of the formation of a spectrum of cloud droplets, raindrops and hail and the dynamic features such as the tornado and the wind gusts (downbursts and microbursts as defined by Fujita, 1981), are the end product of the energy cascade from larger to smaller dimensions with the energy input from phase change of water playing a major role in providing energy to the system.

Silva Dias (1999) presents a review of storms observed in all regions of Brazil, highlighting the damage to life and property associated. Because of their destructive capability it would be desirable to forecast these systems at least a few hours before their formation. However, some storms are more predictable than others. Lilly (1990) indicates that the storms that develop rotation are the more amenable to prediction. According to Browning (1986), rotation is the main feature that distinguishes supercell storms from multicell stroms.

This review focuses on the development of rotation in a thunderstorm. Rotation around the vertical axis measured by vorticity is an indication of possible thunderstorm severity. The interplay of scales that result in a thunderstorm, specifically the development of rotation, will be addressed through the vorticity equation.

2. THUNDERSTORM LIFE CYCLE

A conceptual model of the thunderstorm life cycle begins with some bubble of warm surface air being lifted through buoyancy or forced upwards due to flow over topography, cooled by expansion and thus driven to saturation and condensation to form cloud droplets. The release of latent heat of condensation heats the air parcel and maintains buoyancy further accelerating the air parcel upwards to the tropopause where local stability induces a strong deceleration and limits cloud top. Droplets grow in the updraft, ice may form and enhance the production of large particles, which then may fall either through the updraft, weakening it and even destroying it, and producing rainfall at the surface. When the basic state horizontal wind speed increases with height, the updraft gets tilted and raindrops may fall outside the updraft from mid tropospheric levels. In the latter case, evaporating raindrops cool the air making it negatively buoyant and generating strong downdrafts. In the former case, downdrafts are weaker since the evaporation usually begins below cloud base. The downdrafts hit the ground and become the gust front which propagates horizontally away from the storm behaving as a density current and lifting the warm surface air at the outflow boundary starting again the whole process of cloud formation. The first storm conceptual model was introduced by Ludlam (1966) based on data from the Thunderstorm Project (Byers and Braham,1949). Figure 2, from Houze (1993) shows a conceptual scheme of a mature thunderstorm with the main features outlined.

The development of storm rotation, basically in the updraft, has been studied extensively and conceptual ideas have emerged. Weisman and Klemp (1986) showed through numerical simulations that combinations of vertical wind shear of the environment wind and potentially thermal buoyancy in the environment can explain a wide variety of thunderstorm development from linear systems like squall lines, to isolated short lived and long lived rotating storms. Rotunno and Klemp (1982,1985) have studied the origin of rotation in storms. Their ideas will be explored further in the next section.

3. VORTICITY EQUATION FOR THE THUNDERSTORM SCALE

Most of the interest in the present case lies within the vertical component of the relative vorticity vector, hereafter called just vorticity, which quantifies rotation around the vertical axis z.

The vertical component of the relative vorticity vector is given by

$$\varsigma = \hat{k} \cdot \nabla \times \vec{V} = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \tag{1}$$

where $\vec{V} = u\hat{i} + v\hat{j} + w\hat{k}$ is the three dimension velocity vector with *u, v, w* being the velocity components, defined positive towards the east, north and for rising motion.

The vorticity equation for atmospheric motions is given by (e.g. Holton, 1992):

$$\begin{aligned} \frac{\partial \varsigma}{\partial t} &= -\vec{V} \cdot \nabla(\varsigma + f) - (\varsigma + f)\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) + \\ &\left(\frac{\partial w}{\partial x}\frac{\partial v}{\partial z} - \frac{\partial w}{\partial y}\frac{\partial u}{\partial z}\right) + \frac{1}{\rho^2}\left(\frac{\partial \rho}{\partial x}\frac{\partial p}{\partial y} - \frac{\partial \rho}{\partial y}\frac{\partial p}{\partial x}\right) \end{aligned} \tag{2}$$

On the left hand side of (2) the local change of vorticity is seen to depend on the processes defined by the four terms on the right hand side: the advection of absolute vorticity (*f* is the Coriolis parameter or vorticity due to the earth rotation), the divergence term, the tilting of vortex tubes term and the solenoidal term. The divergence term indicates an increase of vorticity when horizontal convergence concentrates absolute vorticity (the ballerina or stretching term). The tilting of vortex tubes denotes the transformation of rotation around the horizontal axis towards the vertical and the solenoidal term denotes the circulation that arises when density (ρ) gradients are not parallel to pressure (p) gradients.

The vorticity equation in this form is quite general and may be simplified to retain the dominant terms when a given scale of interest is defined. The orders of magnitude of each variable for the thunderstorm scale are given in Table 1 where values of the Coriolis parameter and its meridional variation are typical of mid-latitudes, Ω is the earth angular velocity, φ is the latitude (positive in the northern hemisphere and negative in the southern hemisphere) and a_T is the earth radius (6370 km). For tropical latitudes the results would be the same since the mid latitude case is more restrictive. The fractional horizontal density fluctuation is typical of a gust front with a 3 ^{o} C temperature fluctuation between the cold air in the thunderstorm outflow and the warm inflow air. For the horizontal pressure fluctuation an upper bound on observed pressure variations in the thunderstorm scale has been used.

Table 1: Orders of magnitude for thunderstorm scale analysis

Variable	Symbol	Order of magnitude
Horizontal velocity	U	10 $m.s^{-1}$
Vertical velocity	W	10 $m.s^{-1}$
Horizontal length scale	L	10 km = 10^4 m
Vertical length scale	H	10 km = 10^4 m
Horizontal pressure fluctuation	δp	10 hPa = 1000 Pa
Fractional density fluctuation	$\delta\rho/\rho$	0.01
Air density	ρ	1 $kg.m^{-3}$
Time scale	T = L/U	10^3 s
Coriolis parameter	$f = 2\Omega\sin\varphi$	10^{-4} s^{-1}
Beta	$\beta = df/dy = 2\Omega\cos\varphi/a_T$	10^{-11} m^{-1} s^{-1}
Vorticity	$\zeta \approx U/L$	10^{-3} s^{-1}

Using the values in Table 1, the orders of magnitude of the terms in (2) are:

$$\frac{\partial\zeta}{\partial t}, u\frac{\partial\zeta}{\partial x}, v\frac{\partial\zeta}{\partial y} \approx \frac{U^2}{L^2} = 10^{-6}s^{-2}$$

$$\beta v \approx \beta U = 10^{-11} s^{-2}$$

$$w\frac{\partial\zeta}{\partial z} \approx \frac{UW}{LH} = 10^{-6}s^{-2}$$

$$f\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) \approx f\frac{U}{L} = 10^{-7}s^{-2}$$

$$\zeta\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) \approx \frac{U^2}{L^2} = 10^{-6}s^{-2}$$

$$\left(\frac{\partial w}{\partial x}\frac{\partial v}{\partial z} - \frac{\partial w}{\partial y}\frac{\partial u}{\partial z}\right) \approx \frac{UW}{LH} = 10^{-6}s^{-2}$$

$$\frac{1}{\rho^2}\left(\frac{\partial \rho}{\partial x}\frac{\partial p}{\partial y} - \frac{\partial \rho}{\partial y}\frac{\partial p}{\partial x}\right) \approx \frac{1}{\rho}\frac{\delta\rho}{\rho}\frac{\delta p}{L^2} = 10^{-7}s^{-2}$$

Making the velocity scales in Table 1 one order of magnitude higher and reducing the horizontal length scale in half does not change the relative importance of each term in the vorticity equation described below.

Retaining only the higher order terms, i.e., the terms of order 10^{-6} s^{-2}, the vorticity equation suitable for the thunderstorm scale is,

$$\frac{\partial\varsigma}{\partial t} = -\vec{V}.\nabla\zeta - \zeta\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) + \left(\frac{\partial w}{\partial x}\frac{\partial v}{\partial z} - \frac{\partial w}{\partial y}\frac{\partial u}{\partial z}\right) \qquad (3)$$

or, after rearrangement,

$$\frac{\partial \varsigma}{\partial t} = -\nabla \cdot \left(\vec{V}\zeta\right) + \zeta \frac{\partial w}{\partial z} + \left(\frac{\partial w}{\partial x}\frac{\partial v}{\partial z} - \frac{\partial w}{\partial y}\frac{\partial u}{\partial z}\right) \tag{4}$$

At this point, a separation of scales may be performed. The large scale will be assumed stationary and horizontally homogeneous during the lifetime of the thunderstorm and will be referred to as the basic state which depends only on the vertical coordinate z and consequently has no vertical velocity and zero vorticity. This imposition is valid for the thunderstorm scale because the earth rotation has been neglected, otherwise a thermal wind balance would have to be given for the basic state and the horizontally homogeneous condition would not apply. Separation of the large scale is imposed by

$$\begin{aligned} u &= U_0(z) + u' \\ v &= V_0(z) + v' \\ w &= w' \\ \zeta &= \zeta' \end{aligned}$$

which substituted in (4) give,

$$\begin{aligned} \frac{\partial \varsigma'}{\partial t} &= -U_0 \frac{\partial \zeta'}{\partial x} - V_0 \frac{\partial \zeta'}{\partial y} - \nabla \cdot \left(\vec{V}'\zeta'\right) + \zeta' \frac{\partial w'}{\partial z} + \\ &\left(\frac{\partial w'}{\partial x}\frac{\partial V_0}{\partial z} - \frac{\partial w'}{\partial y}\frac{\partial U_0}{\partial z}\right) + \left(\frac{\partial w'}{\partial x}\frac{\partial v'}{\partial z} - \frac{\partial w'}{\partial y}\frac{\partial u'}{\partial z}\right) \end{aligned} \tag{5}$$

The next step is to perform a Reynolds average by defining the resolvable scale and the subgrid scale processes. The Reynolds average is basically a running average in volume $\Delta v = \Delta x \Delta y \Delta z$ and in the time interval Δt, which for the thunderstorm scale would be of the order of 500 m for the space interval and about a few tens of seconds for the time interval.

The running average $\overline{A}$ for a generic variable A' is defined by

$$\overline{A} = \frac{1}{\Delta x \Delta y \Delta z \Delta t} \int\limits_{\Delta t} \iiint\limits_{\Delta x \Delta y \Delta z} A' dz dy dx dt \tag{6}$$

with the subgrid scale contribution A'' introduced as the residual

$$A' = \overline{A} + A'' \tag{7}$$

and the usual operations

$$\overline{A''} = 0$$
$$\overline{A'B'} = \overline{A}\,\overline{B} + \overline{A''B''}$$
$$\overline{\frac{\partial A'}{\partial x}} = \frac{\partial \overline{A}}{\partial x}$$

All variables are separated according to (7) in their resolved scale and sub-grid scale components, substituted into (6) and the running average (6) is applied to the resulting equation, according to the Reynolds average procedure (e.g. Holton, 1992). The result is

$$\frac{\partial \overline{\varsigma}}{\partial t} = -U_0 \frac{\partial \overline{\varsigma}}{\partial x} - V_0 \frac{\partial \overline{\varsigma}}{\partial y} - \nabla \cdot \left(\overline{\vec{V}}\,\overline{\varsigma}\right) + \frac{\partial \overline{\zeta}\overline{w}}{\partial z} +$$
$$\left(\frac{\partial \overline{w}}{\partial x}\frac{\partial V_0}{\partial z} - \frac{\partial \overline{w}}{\partial y}\frac{\partial U_0}{\partial z}\right) + \left(\frac{\partial \overline{w}}{\partial x}\frac{\partial \overline{v}}{\partial z} - \frac{\partial \overline{w}}{\partial y}\frac{\partial \overline{u}}{\partial z}\right) + SG \tag{8}$$

or

$$\frac{\partial\bar{\varsigma}}{\partial t} = -\left(\vec{V}_0 + \vec{\bar{V}}\right)\cdot\nabla\bar{\zeta} - \bar{\zeta}\left(\frac{\partial\bar{u}}{\partial x} + \frac{\partial\bar{v}}{\partial y}\right) +$$

$$\left(\frac{\partial\bar{w}}{\partial x}\frac{\partial V_0}{\partial z} - \frac{\partial\bar{w}}{\partial y}\frac{\partial U_0}{\partial z}\right) + \left(\frac{\partial\bar{w}}{\partial x}\frac{\partial\bar{v}}{\partial z} - \frac{\partial\bar{w}}{\partial y}\frac{\partial\bar{u}}{\partial z}\right) + SG$$

I II III IV (9)

where

$$SG = -\nabla\cdot\overline{\vec{V}''\zeta''} + \overline{\frac{\partial w''}{\partial x}\frac{\partial v''}{\partial z}} - \overline{\frac{\partial w''}{\partial y}\frac{\partial u''}{\partial z}} \tag{10}$$

From (9) it is seen that the vorticity in the thunderstorm scale responds to:

I - the advection of vorticity by the basic state and thunderstorm winds.

II - The non-linear interaction between thunderstorm scale convergence and vorticity. This term acts in intensifying vorticity through convergence.

III - The interaction between basic state vertical wind shear and thunderstorm scale vertical velocity gradients; this term creates vorticity in the thunderstorm scale from the basic state vertical wind shear. Rotunno and Klemp (1982) discuss this term and show how it can be used to explain some observed features of moving storms. This will be discussed further below.

IV - The tilting of thuderstorm vortex tubes; this term also creates/destroys vorticity as soon as the vertical circulations and horizontal structure of the thunderstorm begin to dominate.

V - The effect of sub-grid scale turbulent motions.

From (9) an evolution of the thunderstorm rotation can be conceptually described. Starting from an environment with vertical wind shear, the only available vorticity is in the form of horizontal vorticity defined by the vertical variation of $U_0(z), V_0(z)$. As soon as a first cell is formed as described in the previous section, a $\overline{w}(x, y, z)$ field of vertical velocity is defined with positive values in the updraft region and negative values in the downdraft region. To develop vorticity in the thunderstorm scale, (9) shows that term III, the tilting of large scale horizontal vortex tubes to the vertical by thunderstorm scale vertical motions is fundamental (see Fig. 3). This is in accordance with Rotunno and Klemp (1982) and with Davies-Jones (1984, et al.,1994). As soon as term III starts to act to produce thunderstorm scale vertical vorticity, term II begins to strengthen it. Term II is a non-linear term. At the beginning, while vorticity is zero, thunderstorm convergence associated with the updraft cannot produce vorticity. However, as soon as term III extracts vorticity from the basic state and transforms it into the thunderstorm scale vorticity, term II increases the updraft rotation. Rotunno and Klemp (1982) show that term III explain the formation of two vorticity centers in the right and left flanks of the updraft. When the wind shear vector rotates with height, Rotunno and Klemp (1982) show that one of the vorticity centers is enhanced by the vertical pressure gradient force producing vertical acceleration (see the Appendix for the equation for the vertical velocity). For the northern hemisphere, it is common that the environment wind shear associated with storms is such that the wind shear vector turns clockwise with height. In this case the right side vorticity center is favored and it rotates counterclockwise (cyclonically in the northern hemisphere). In the southern hemisphere, the wind shear vector typically turns counterclockwise with height is storm days, and this favors the left side vorticity center which turns clockwise being also cyclonic in the southern hemisphere. The favored cyclonic center enhances its rotation through term II and evolves to be a strong rotating thunderstorm, or supercell storm, while the other decays.

Term IV is also a non linear term. The tilting of vortex tubes within the thunderstorm indicates that storm created horizontal vortex tubes due to, for example, the spreading out of the gust front at low levels, may continuously produce more vertical vorticity.

Arguments concerning the predictabilty of storms with well defined rotating updrafts point to term III as the main physical process that gives a certain behavior. In the absence of environmental shear, rotation may develop through term IV in later stages when the storm produces the acceleration at low levels and at cloud top. However, mid-level rotation will hardly be developed through this process.

4. NUMERICAL SIMULATION OF STORMS IN BRAZIL

Among the several systems described by Silva Dias (1987, 1989, 1999) some have been numerically simulated in the thunderstorm scale with some success. Well defined large scale wind shear is a common feature of these systems. The cases described below have been numerically simulated using the – Regional Atmospheric Modeling System - RAMS (Pielke et al, 1992). The successful numerical modeling of these cases, measured by resemblance to observed features, is an indication of predictability.

Cohen and Silva Dias (1997) focus on the squall lines that develop at the northern coast of Brazil and propagate to the Amazon Basin with lifetimes that frequently are in excess of 24 hours. Cohen et al (1995) show that the environment of the Amazonian squall lines is dominated by a low level easterly jet and a westerly upper level jet giving a strong (by equatorial standards) wind shear.

Menezes and Silva Dias (1996, 1998 a,) focus on two cases of severe storms that generated tornados and damaging winds in southeast

Brazil, hitting the cities of Itú (30 September, 1992) and Ribeirão Preto (14 May, 1994). The Itú case was of a squall line in unidirectional shear while the Ribeirão Preto case was of supercell in directional shear. Both cases follow Weisman and Klemp (1986) conceptual model of the effect of wind shear in thunderstorm development. Menezes and Silva Dias (1998b) perform a vorticity budget of the two cases. They calculated term II and terms III+IV of Eq. (9) noting that in both cases the sum of III+IV, the tilting of vortex tubes by large scale plus storm scale winds, is the main process and its magnitude is larger the divergence term II. In the Ribeirão Preto case vorticity and vertical velocity extremes are co-located indicating the updraft rotation; at low levels the main source of rotation though the tilting terms is the gust front region. In the Itú case, vorticity couplets appear in the flanks of the vertical velocity maxima, splitting the updraft and generating the multicelular structure typical of a squall line.

5. CONCLUDING REMARKS

The evolution of storm scale vorticity is strongly linked to larger scale vertical wind shear and the development of storm scale rotation is an indication of dominance of a dynamical process that supports predictabilty. This is the basic conclusion of the previous sections with the support of extensive work in the literature.

The interplay of scales in the development of a thunderstorms is not complete without looking into the smaller scales, or subgrid scale processes, which certainly have an effect either in enhancing or destroying rotation. Vorticity budgets performed for larger scales in the tropics (e.g. Sui and Yanai, 1986) have shown that subgrid scale processes including convective development are sometimes dominant in the vorticity equation. In the thunderstorm scale vorticity budget, subgrid scale processes are dominated by turbulence and by condensation/evaporation, freezing/melting heat inputs. A clear understanding of these, in the framework of storm scale vorticity, is still lacking.

Acknowledgements

Some of the research repported here has been supported by grants from FAPESP and CNPq.

Appendix

The equation for the vertical velocity couples the field of motion to the thermodynamic forcings in a thunderstorm. Following the same procedure used in section 3, and assuming a hydrostatic balance in the basic state

$$\frac{dp_0}{dz} = -\rho_0 g \, ,$$

where g is the acceleration due to gravity, the vertical acceleration in the thunderstorm scale is given by

$$\frac{\partial \overline{w}}{\partial t} = -\left(\vec{V}_0 + \vec{\overline{V}}\right) \cdot \nabla \overline{w} - \frac{\overline{\rho}}{\rho_0} g - \frac{1}{\rho_0}\left(1 - \frac{\overline{\rho}}{\rho_0}\right)\frac{\partial \overline{p}}{\partial z} + SG_w \tag{A1}$$

where

$$SG_w = -\overline{\vec{V}'' \cdot \nabla w''} + \overline{\frac{\rho''}{\rho_0^2}\frac{\partial p''}{\partial z}} \tag{A2}$$

As seen in (A1), vertical velocity may be advected by the basic state and thunderstorm scale winds, and acceleration is affected by the buoyancy term (warm, relatively less dense air produces positive acceleration), the vertical pressure gradients towards the thunderstorm relative low pressure ($\overline{p}$ is perturbation upon the basic state $p_0(z)$ where the average profile of pressure decreasing with height is already taken into account) and the sub-grid scale turbulent processes.

The buoyancy term indicates that as soon as condensation takes place, the release of latent heat introduces a negative perturbation on the density field which in turn produces vertical acceleration. On the other hand, an evaporating population of cloud or rain drops are associated with a colder denser air parcel which is accelerated downward.

Since $\left(1-\frac{\overline{\rho}}{\rho_0}\right)>0$, the thunderstorm pressure gradient creates acceleration towards the relatively low pressure in the vertical.

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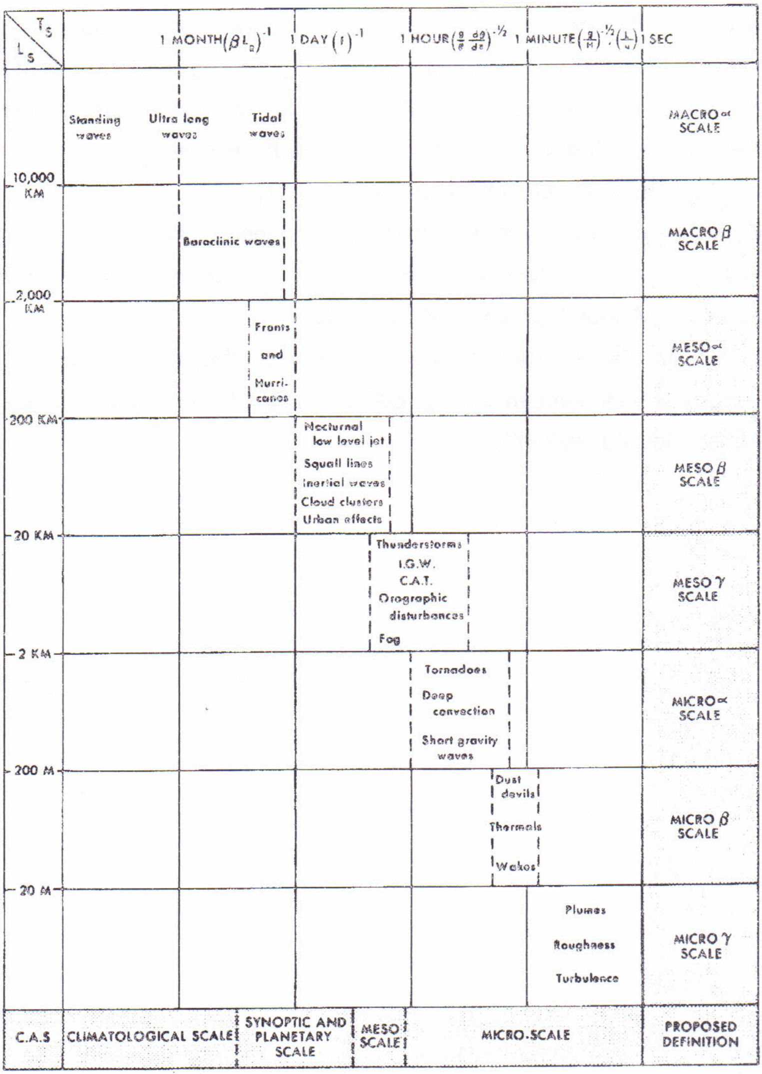


Figure 1: Scale definitions and different processes with characteristics time and space scales (from Orlanski, 1975).

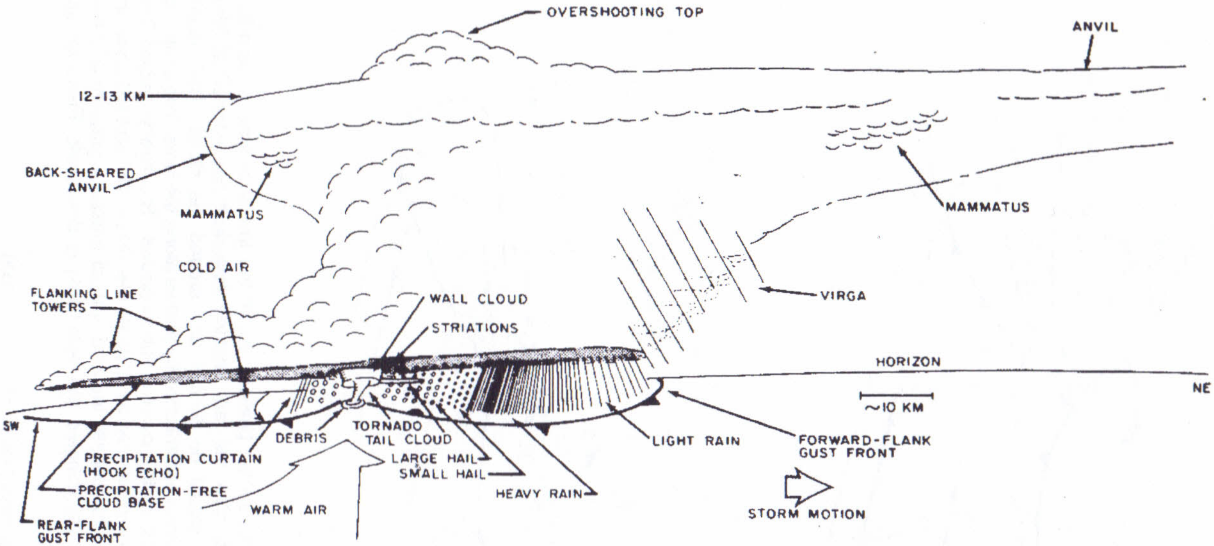
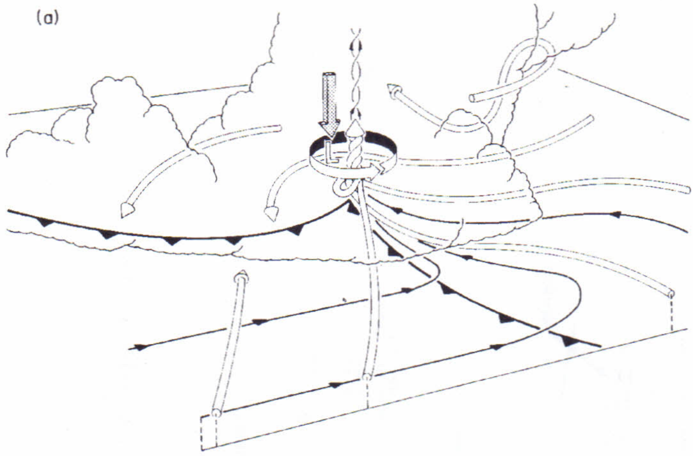


Figure 2: Schematic visual appearance of a thunderstorm (from Houze, 1993)

(a)



(b)

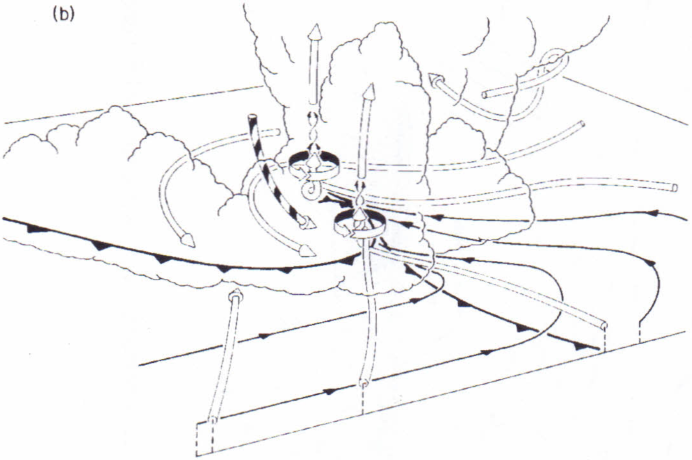


Figure 3. Expanded three-dimensional perspective, of the low-level flow (a) and (b) about 10 min later after the rear flanking downdraft has intensified. The cylindrical arrows depict de flow in and around the storm. The vector direction of vortex lines are inticated by arrows along the lines. The sense of rotation is indicated by the circular ribbon arrows. The heavy barbed line works the boundary of the cold air beneath the storm. The shaded arrow in (a) represents the rotationally induced vertical pressure gradient and the stripped arrow in (b) denotes the rear-flanking downdraft. From Klemp (1987).