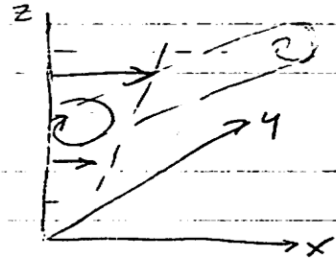
Parameters affecting thunderstorms

Role of wind shear

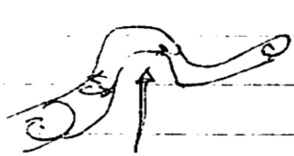
Simplest case is for <u>unidirectional</u> <u>shear</u>: consider a case with a storm growing in constant unidirectional shear. Such a sheared profile has a component of vorticity that is pointed <u>perpendicular</u> to the shear vector:



$$\zeta_y = \frac{\partial u}{\partial z} - \frac{\partial w}{\partial x} \approx \frac{\partial u}{\partial z} \quad \left[\text{since } w \text{ usually small and } \delta x \text{ large} \right]$$

So we have "vortex tubes" that are oriented along the y-axis; i.e., perpendicular to the shear vector.

Now suppose we have an updraft of finite extent. The updraft lifts the center part of the vortex tube:



This lifting acts to <u>tilt</u> <u>the</u> <u>horizontal</u> <u>vorticity</u> <u>into</u> <u>the</u> <u>vertical</u>. Mathematically, the tilting term in the tendency eq. for the vertical vorticity is:

$$\frac{\partial \zeta_z}{\partial t} = \frac{\partial w}{\partial y}\zeta_y = \frac{\partial w}{\partial y}\left(\frac{\partial u}{\partial z} - \frac{\partial w}{\partial x}\right) \approx \frac{\partial w}{\partial y}\frac{\partial u}{\partial z}$$

Then the <u>perturbation</u> (i.e., convective) updraft acts on the <u>environmental</u> vorticity due to the sheared profile to create <u>two</u> vortices: viewed from behind

- a <u>cyclonic</u> (CCW) vortex to the <u>right</u> of the original updraft, where $\frac{\partial w}{\partial y} > 0$, and
- an <u>anticyclonic</u> (CW) vortex to the <u>left</u> of the updraft, " where $\frac{\partial w}{\partial y} < 0$

Each of the vortices is associated with a low-pressure perturbation which typically is strongest in the mid levels. (cyclostrophic balance is helpful to understand why <u>both</u> the cyclonic and anticyclonic vortices have low pressure)

$$-\frac{1}{\rho}\frac{\partial p}{\partial n} = \frac{V^2}{R}$$

$$-\frac{\partial p}{\partial n} = \frac{\rho V^2}{R}$$

$$\frac{\partial p}{\partial n} = -\frac{\rho V^2}{R}$$

$$V = \sqrt{-\frac{R}{\rho}\frac{\partial p}{\partial n}}$$

(discussed some by Klemp & Weisman article)

The effect of <u>directional</u> shear is much more complicated to interpret. Operational forecasting has recently focused on two shear-related parameters:

- bulk Richardson number
- helicity

if $R > 0$

$\frac{\partial p}{\partial r} < 0$

if $R < 0$

$\frac{\partial p}{\partial n} > 0$

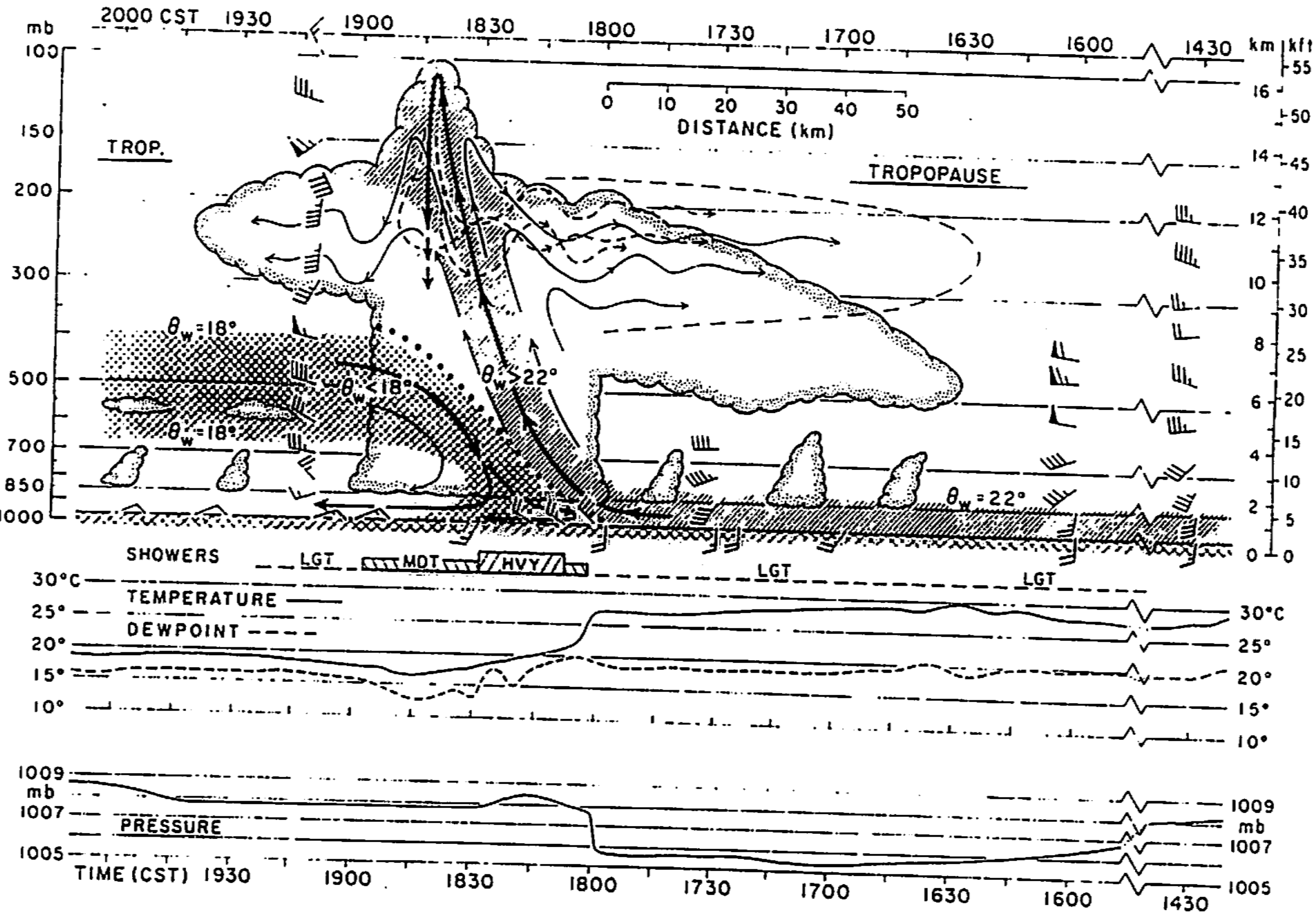


Figure 16.5b. Cross section through squall line of 21 May 1961 as it passed Oklahoma City. Hatching indicates probable extent of high θ_w air of low-level origin; cross-hatching indicates location of low θ_w air of probably middle-level origin. Heavy arrows are axes of main drafts; thin arrows are streamlines, dashed where air emanates from core of stratospheric tower. Long dashes suggest outline of air plume originating in storm; at lower altitude cloud plume consists of small precipitation particles. (After Newton, 1966.)

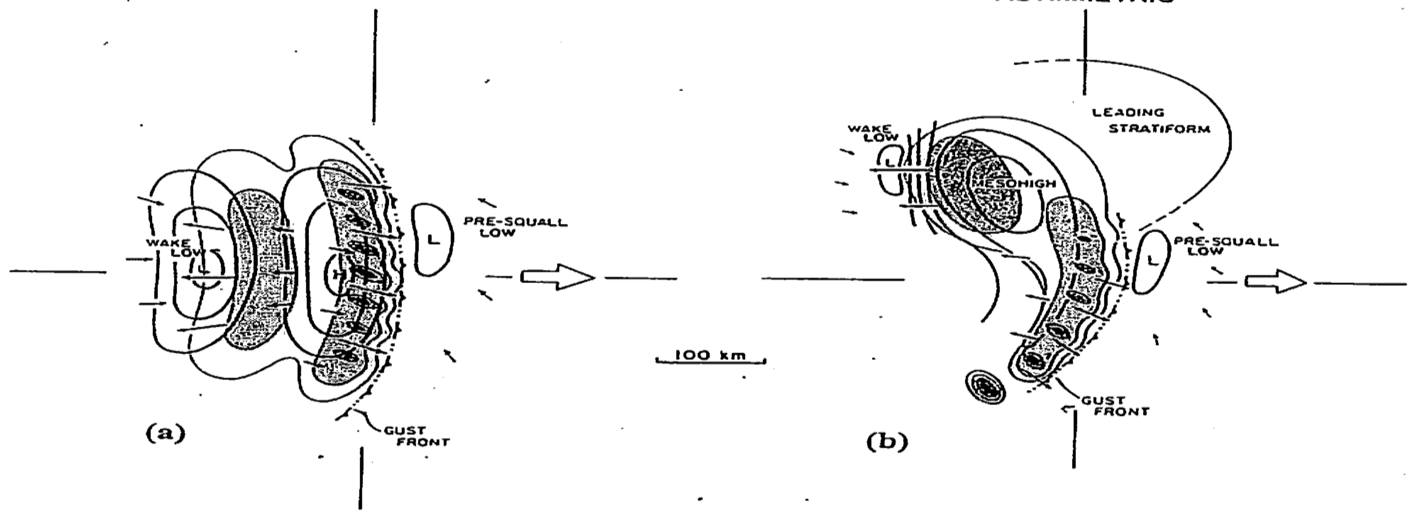


FIG. 9.19. Conceptual model of the surface pressure, flow, and precipitation fields associated with (a) symmetric and (b) asymmetric stages of the MCS life cycle. Levels of shading denote increasing radar reflectivity, with darkest shading corresponding to convective cell cores (adapted from Houze et al. 1990). Pressure is in 1-mb increments. Small arrows represent the surface flow. Length of the arrow is proportional to the wind speed found at its center. The large arrows represent the storm motion. From Loehrer and Johnson (1995).

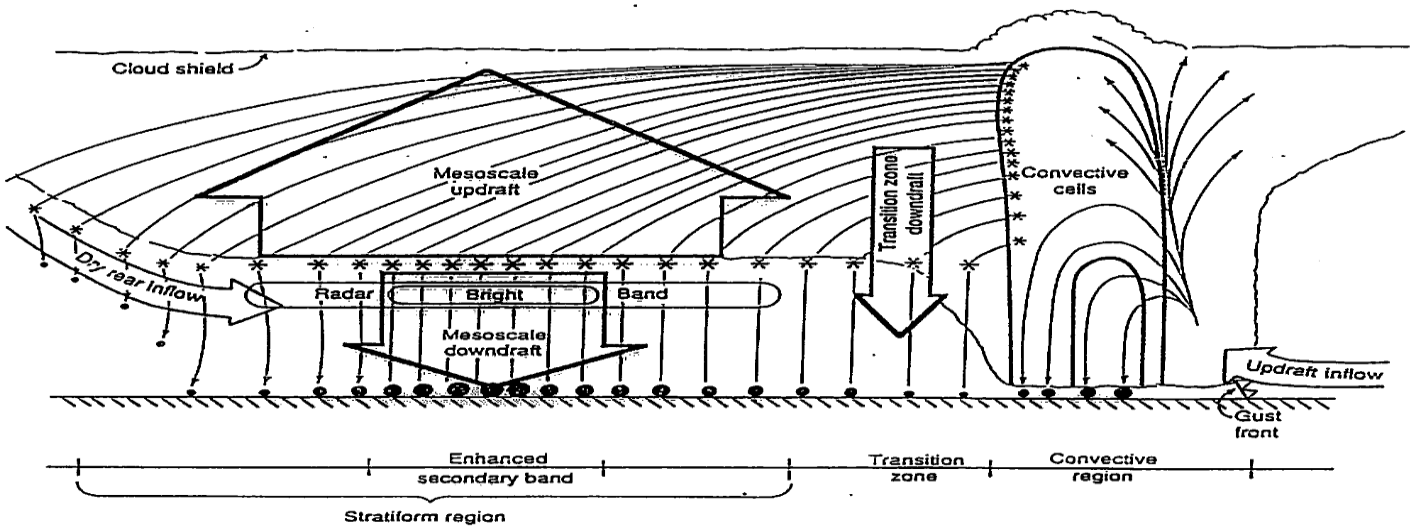


FIG. 3.26. Conceptual model of the two-dimensional hydrometeor trajectories through the stratiform region of a squall line with trailing stratiform precipitation. From Biggerstaff and Houze (1991).

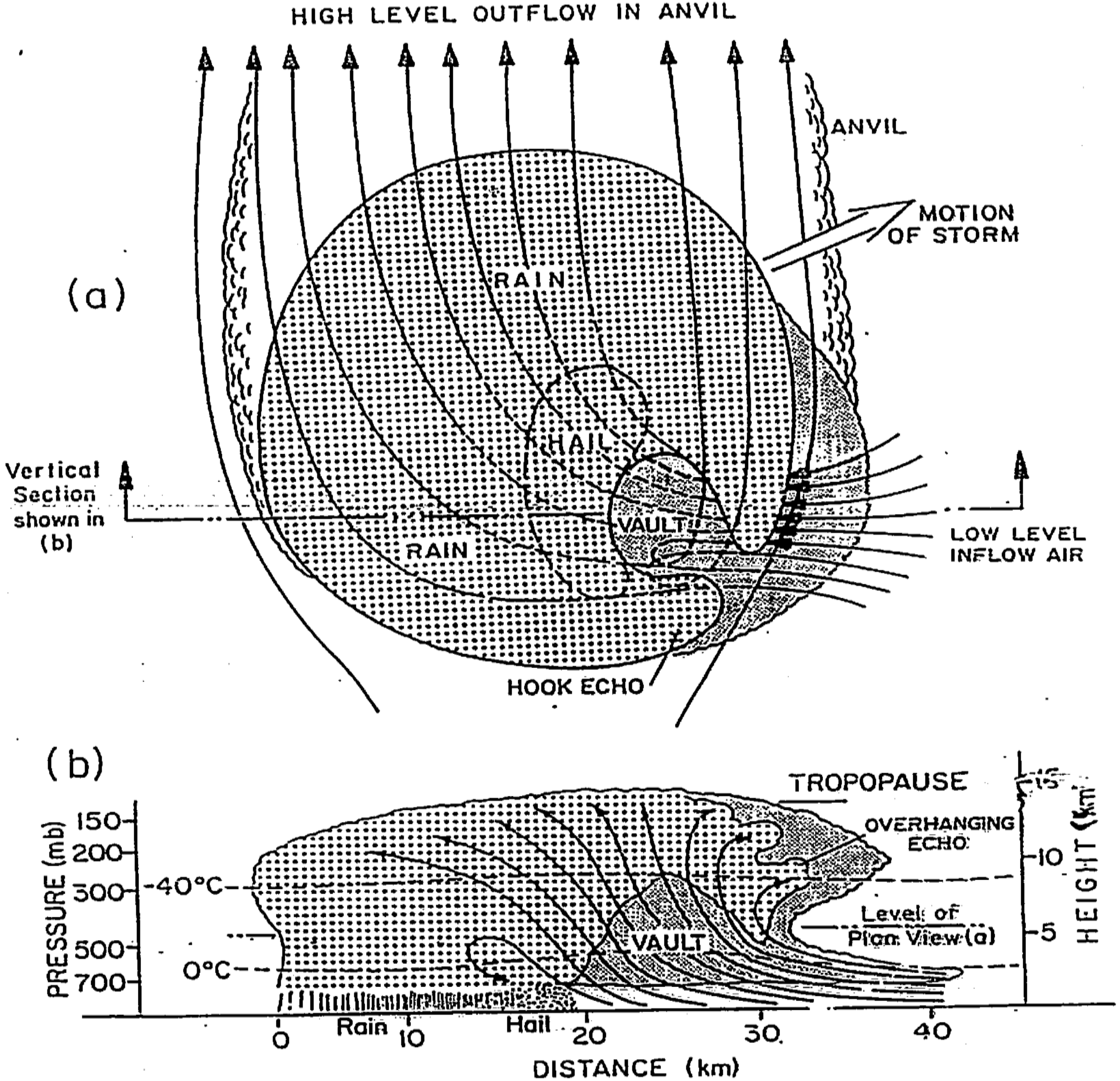


Fig. 5.27a,b Structure of a supercell storm (a) as seen from above and (b) as seen in a vertical cross section looking downstream in the direction of the midtropospheric winds. Areas of heavy stippling correspond to radar echoes; lighter stippling denotes clouds composed of small (non-reflective) cloud droplets. Arrows show projections of air motions in the updraft relative to the storm. Note that the arrows in (a) represent motion of air entering the storm at low levels and leaving at high levels, and that in (b) there is a component of air motion perpendicular to the page. [Adapted from *Quart. J. Roy. Met. Soc.* 102, 499 (1976).]

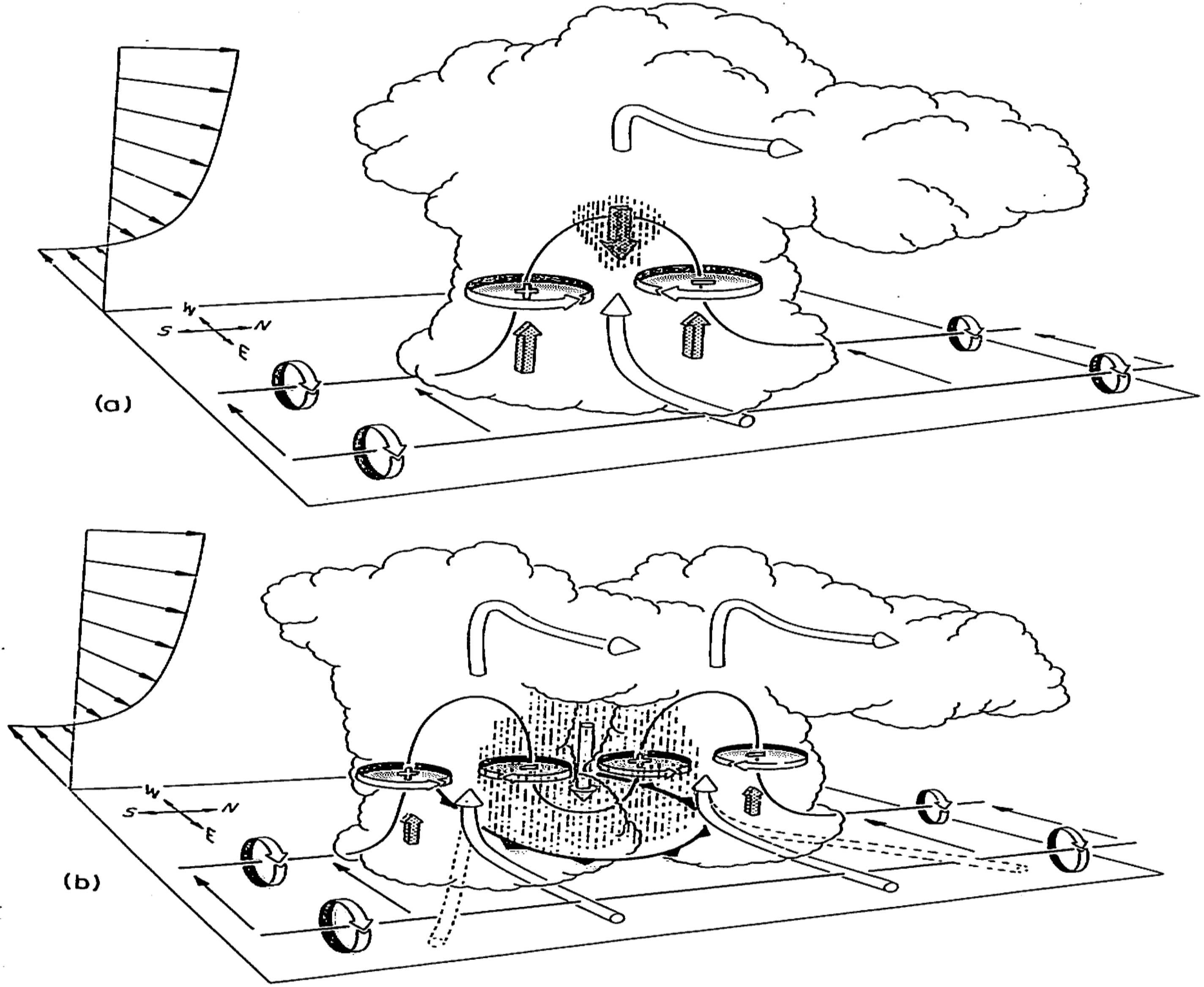


Figure 3.23 Schematic diagram depicting how a typical vortex line (streamline of three-dimensional vorticity vector) contained within (westerly) environmental shear is deformed as it interacts with a convective cell (viewed from the southeast). Direction of cloud-relative airflow (cylindrical arrows); vortex lines (solid lines), with the sense of rotation indicated by circular arrows; the forcing influences that promote new updraft and downdraft growth (shaded arrows); regions of precipitation (vertical dashed lines). (a) Initial stage: Vortex line loops into the vertical as it is swept into the updraft. (b) Splitting ...

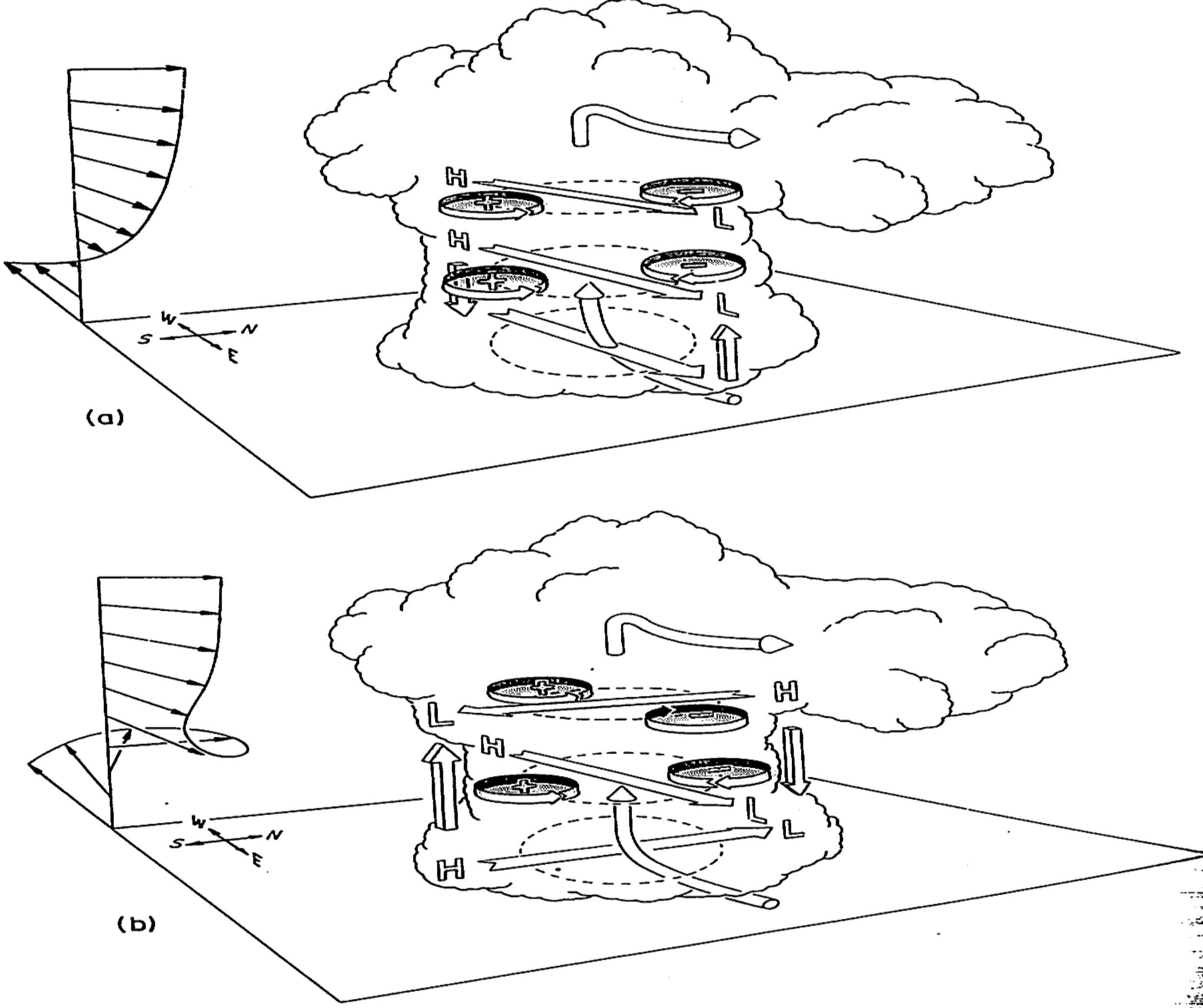


Figure 3.21 Schematic diagram illustrating the pressure and vertical vortic perturbations arising as an updraft interacts with an environmental vertical w shear vector that (a) does not change direction with height and (b) turns clockw with height. The high (H) to low (L) horizontal pressure-gradient forces parallel the shear vectors (flat arrows) are labeled along with the preferred location cyclonic (+) and anticyclonic (−) vorticity. Shaded arrows depict orientation resulting vertical pressure-gradient forces (from Klemp, 1987; adapted Rotunno and Klemp, 1982). (Courtesy of the American Meteorological Society)

Bulk Richardson number

The usual expression for the Richardson number is

$$Ri = \frac{\frac{g}{\theta}\frac{\partial \theta}{\partial z}}{\left[\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2\right]} = \frac{N^2}{S^2}$$

where $S = \left[\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2\right]^{1/2}$ is the magnitude of the vector wind shear. (Physical meaning re: TKE)

If we generalize the Ri to

$$\frac{\text{some measure of buoyancy}}{(\text{some measure of shear})^2}$$

we can define a type of Richardson number for deep convection as

$$R = \frac{PBE}{(\Delta U)^2/2} = \frac{CAPE}{(\Delta U)^2/2}$$

where

$$\Delta U = \bar{U}_{6000} - \bar{U}_{500}$$

each $\bar{U}$ is a density-weighted mean wind over some depth; i.e.

$$\bar{U}_{\hat{z}} = \frac{1}{\bar{\rho}} \int_0^{\hat{z}} \rho(z)\, u(z)\, dz$$

It is also worth noting that for simple undilute (non-entraining) parcel ascent,

$$\frac{1}{2} W_{ETL}^2 = PBE = \int_{LFC}^{ETL} g \frac{\theta'}{\theta_0}\, dz$$

i.e., at the ETL, the potential energy due to buoyancy has been completely converted to the kinetic energy of the rising parcel. We could also say

$$W_{ETL} = (2\ PBE)^{1/2}$$

So R could also be thought of as

$$R \sim \frac{W^2_{ETL}}{\Delta U^2}$$

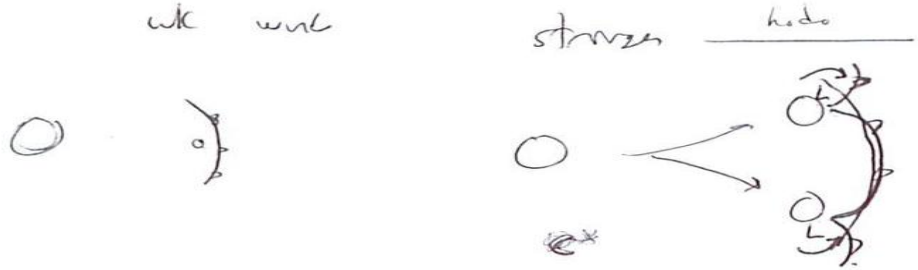
i.e., the ratio of the updraft kinetic energy to the shear.

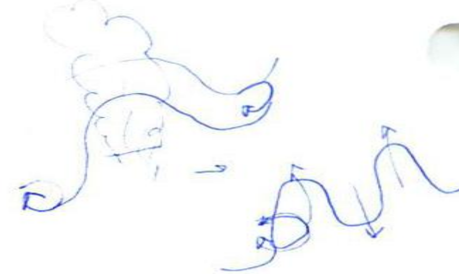
R helps to discriminate between different kinds of storms, but must also be considered along with the buoyancy itself.

- For very small PBE, precipitating convection is unlikely
- For "small" PBE, single-cell storms are produced if the shear is small (large R) while for strong shear (small R) the shear tends to break the storm apart.
- For moderate to large B ($\sim 1500-4000$ J kg^{-1} (or $m^2 s^{-2}$)), supercells are most likely if R is in the range of about 5-50. (40)

Evidence suggests that the main effect of R is in controlling whether supercells develop, i.e., critical value of R above which the environmental shear is insufficient to generate supercells. Can still have multicellular convection in the vicinity of supercells.

Physical explanations for the effect of R on storm morphology are not very clear. Weisman + Klemp discuss role of shear in creating supercells (tilt horiz. vorticity into vertical → dynamically induced pressure gradient → acceleration of updraft) wk wind   stronger hodo

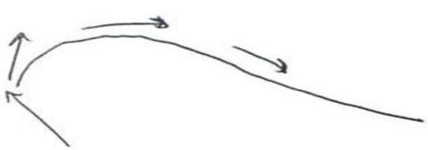




curved hodo

A   B

veering winds w ht ~~are good~~ permit ~~[illegible]~~ long-lived, organized, severe thunderstorms in general
but it is the veering of wind shear vector that determines
right-moving vs left moving storms



vs

A) R = 89, curved hodo, wk shear → short-lived multicell
B) R = 22 curved, but twice shear → supercell on south end of multicell line
C R = 15 curved low, straight high → right flank supercell, wk left flank storms
* R = 12 straight hodo → mirror image storms
D) R = 14 – shear is straight (like) but over bigger depth → right-flank supercell (clearer look) as rain goes farther downstream
2.5
E) shear stops at low levels → wk squall line
twice as much
F) V. strong lowlevel only shear → bow echo type storm with rotating comma head

Helicity

from Latin, Greek *helix*; cf. *helicopter*

- a measure of the tendency for the flow to be aligned with the vorticity
- several measures of "helicity":

formally, it is the volume integral

$$\text{Helicity} = \iiint (\vec{V} \cdot \vec{\omega})\, dx\, dy\, dz$$

the helicity density (often called just the "helicity") is

$$H = \vec{V} \cdot \vec{\omega} = \vec{V} \cdot (\nabla \times \vec{V})$$

notice

- H is a scalar, unlike vorticity (dot product). No "direction" to the helicity; doesn't matter e.g. whether H is due to horizontal or vertical vorticity-velocity correlations.
- H is maximized when the vorticity is directed along the velocity vector. By the definition of the dot product, we can define the relative helicity as

$$H_{rel} = \frac{H}{\underbrace{|\vec{\omega}||\vec{V}|}_{\text{magnitudes, not vectors}}} = \cos(\vec{\omega}, \vec{V})$$

Notice that $-1 \le H_{rel} \le +1$; $H_{rel} = 0$ if vorticity and velocity are perpendicular (angle between is $\frac{\pi}{2}$). H_{rel} is, H normalized by its maximum possible value.

In fluid dynamics there is a special type of flow called Beltrami flow in which the velocity vector is always aligned with the vorticity vector; i.e., helicity is maximized, so that

$$\vec{\omega} = c_\psi \vec{V}$$

where c_ψ is a constant along a given streamline.

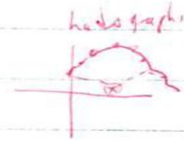
Notice helicity is not a Galilean invariant. In particular, the storm-relative helicity can differ markedly from the helicity as viewed from a fixed point in space. $\vec{v}$

$\vec{\omega}$

if $\vec{v} = \vec{c}$, then $(\vec{v} - \vec{c}) \cdot \vec{\omega} = 0$
but $\vec{v} \cdot \vec{\omega} \neq 0$

[example / cartoon: storm moving exactly in the direction of the vorticity vector]
~speed.

Then we can define the storm-relative helicity (not to be confused with the "relative helicity") as



$$H_{SR} = (\vec{V} - \vec{c}) \cdot \vec{\omega}$$

Typically this is integrated through some depth of the profile, i.e.,

$$H_v = \int_0^{\hat{z}} (\vec{V} - \vec{c}) \cdot \vec{\omega} \, dz$$ [What are units?]

Here $\vec{\omega}$ is the "streamwise" vorticity,

$$\vec{\omega} = \hat{k} \times \frac{d\vec{V}}{dz}$$

$$\left.\begin{array}{ccc} i & j & k \\ 0 & 0 & 1 \\ \frac{\partial u}{\partial z} & \frac{\partial v}{\partial z} & \frac{\partial w}{\partial z} \end{array}\right\} \quad -\frac{\partial v}{\partial z} i + \frac{\partial u}{\partial z} j$$

- an anticyclonic (CW) vortex to the left of the updraft - i.e. the center where $\frac{\partial w}{\partial y} < 0$

TABLE 5.2. Comparison of the properties of three forecast parameters (ignoring dependence of BRN on CAPE because other two parameters may be nondimensionalized by CAPE and $\sqrt{CAPE}$ if desired).

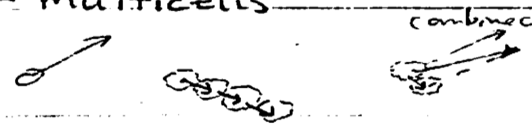
Property	SRH	BRN	Length of HODO.
Predictor of	Net updraft rotation (including sign)	Supercells, splitting storms	Tornadic storms, mesocyclones
Physics contained	Potential for midlevel updraft rotation	Buoyant energy/inflow kinetic energy	Mean magnitude of shear in 0–4-km layer
Geometric interpretation on hodograph diagram	$-2 \times$ signed area swept out by storm-relative wind between 0 and 3 km	BRNKE $\equiv (\Delta\bar{U})^2/2$ $\Delta\bar{U} \equiv \lvert\bar{\mathbf{v}}_{0-6} - \bar{\mathbf{v}}_{0-.5}\rvert$	Arc length
Dependence on hodograph shape	Highly dependent	Less dependent	Independent
Function of storm motion c?	Linear function of observed (or forecast) c	No	No
Sensitivity to c (nearly straight hodographs)	Highly sensitive	None	None
Sensitivity to c (highly curved hodographs)	Insensitive. Substituting mean wind for c gives good estimate of helicity	None	None
Differentiates storms in same environment by their motion?	Yes, may warn on storm that deviates to right (e.g., along a boundary)	No	No
Mirror image storms	Opposite signs	Same value	Same value
Sensitivity to surface-inflow wind	Moderate	High (important for supercell vs multicell)	Low
Volatility with respect to hodograph changes	More volatile	Less volatile	Less volatile
Galilean invariant?	Yes (no if estimate of c is not invariant)	Yes	Yes
Winds included	0–2 or 3 km explicitly but higher winds affect c	0–6 km	0–4 km
Sensitive to fractal hodographs	No	No	Yes, needs prescription for smoothing hodographs

Storm motion

Factors affecting storm motion are still not well understood. Two main ways in which the storm moves:

1. continuous propagation - single + super cells
2. discrete propagation - multicells

[draw cartoons]



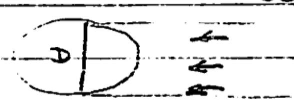
Various mechanisms have been proposed to explain storm motion (e.g., "steering level").

One idea that is fairly interesting assumes a balance between the rainfall rate (water used) for the total storm and the moisture supply.

rainfall: $R = c_1 \frac{\pi}{4} D^2$

moisture supply: $S = D V_{RL}$

where D = diameter of the storm



V_{RL} = storm-relative inflow rate

c_1 = proportional to rain rate

Then if $S = R$, (supply = rainfall), (and rain efficiency?)

$$D V_{RL} = c_1 \frac{\pi}{4} D^2$$

$$V_{RL} = c_1 \frac{\pi}{4} D$$

indicates:

- larger storms have greater relative motion (don't necessarily move fast; could be a stationary storm in a high-wind-speed environment)
- greater rainfall rate per unit area → greater storm-relative motion.