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Thunderstorms I

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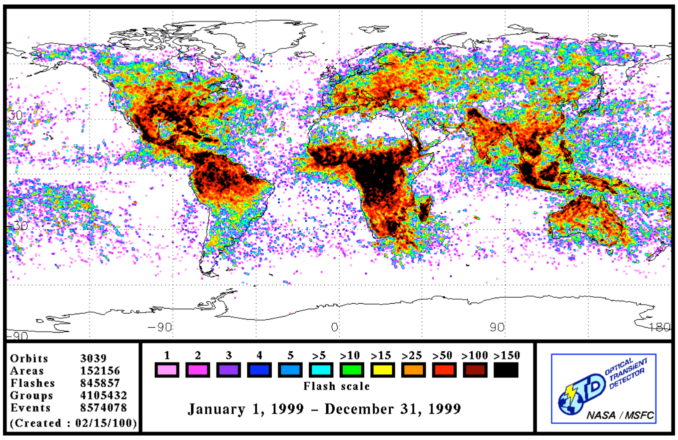
Motivation



(http://thunder.msfc.nasa.gov/)

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Development stages of a Cumulonimbus (Cb)

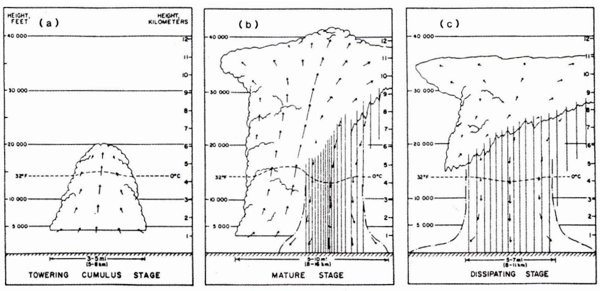


Fig. 9.1 Three stages in the evolution of common convective showers. (a) cumulus stage; (b) mature stage; (c) dissipating stage. Altitude in kilometers on right. Arrows denote air motion; dash-dot lines represent boundary of rain-cooled air. [*From Doswell (1985).*]

Figure: Three stages of cumulonimbus clouds (Emanuel, 1994)

Empirical model I

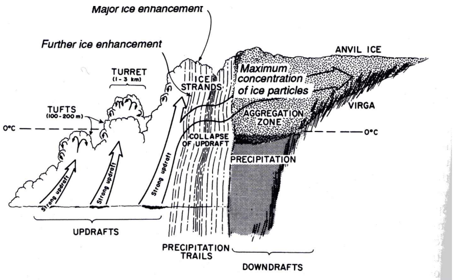


Figure 8.1 Empirical model of a small cumulonimbus cloud. Based on about 90 research aircraft penetrations of small cumulonimbus and large cumulus clouds. (From Hobbs and Rangno, 1985. Reprinted with permission from the American Meteorological Society.)

Houze (1993), Fig. 8.1

Characteristics:

- younger developing side (left)
- older glaciated side (right)

Empirical model II

Developing side:

- cumulus cell with updraft motion
- buoyant cores, leading to cloud turrets ($\Delta x \sim 1 - 3$ km)
- within turret:
  - overturning as in a thermal ($\rightarrow$ cumulus dynamics)
  - generating of horizontal vorticity (see below)
- on each turret: spherical tufts (diameter $\sim 100 - 200$ m)
- across the Cb the tops of turrets are formed at increasingly higher altitudes (i.e. new updrafts form systematically on the developing side of the cloud)

Empirical model III

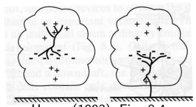
Glaciated side:

- after reaching above 0°C–level:
  some droplets are as large as 20μm$\rightarrow$ ice enhancement
- high ice concentrations ($1 - 100\mathrm{L}^{-1}$) in localized regions
- ice particles extend vertically, ice strands are formed at lower levels, graupel can be formed
- aggregation of ice particles just above of the 0°C–level $\rightarrow$ formation of stratiform precipitation
- decaying / collapsing of the updrafts:

$$B = g\left(\frac{T^*}{T_0} + \frac{p^*}{p_0} + 0.61 \cdot q_v^* \underbrace{-q_H}_{downdraft}\right) \quad (1)$$

Lightning in Cb I

- sequence of lightning typical for Cbs
- no lightning until cloudtop rises above $-15/-20°C$ –level
- two types of lightning:
  1. Intracloud
  2. Cloud–to–ground



Houze (1993), Fig. 8.4

Physics:

- positive and negative charges become separated within the region of cloud and precipitation
- lightning = transfer of charge
- $T \approx 30000$ K in narrow channel
- pressure enhancement by 1–2 orders of magnitude ⇒
  - supersonic shockwave
  - soundwave (thunder)

Lightning in Cb II

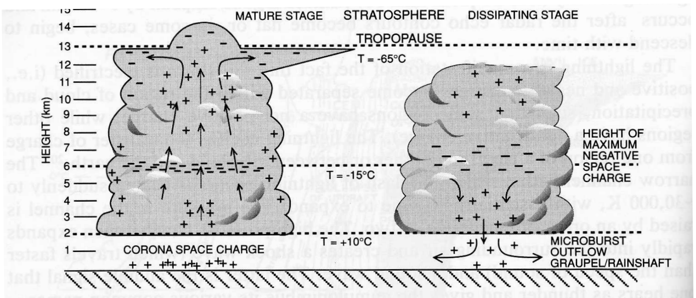


Figure 8.3 Schematic of the electrical structure of a cumulonimbus cloud. Positive and negative signs indicate the polarity of the charge at various locations. Streamlines indicate direction of airflow. (From Williams, 1988. © Scientific American, Inc. All rights reserved.)

Figure: Distribution of charge, Houze (1993)

Lightning in Cb III

- main negatively charged region is sandwiched between two positively charged regions
- characterisitics of negatively charged region:
  - vertical extension < 1 km
  - horizontal extension over several kilometres
  - located at $\sim -15°C$

Mechanism of electrification (current research)

- transfer of charge when graupel particles collide with small ice crystals (H^+ ions)
- polarity depends strongly on temperature and liquid water content
- critical temperature (“charged–reversed–temperature”, T_{cr}) in the range $-10 \geq T_{cr} \geq -20°C$
  - for $T < T_{cr}$: negative charge transferred to graupel
  - for $T > T_{cr}$: positive charge transferred to graupel

Lightning in Cb IV

Main conclusions/observations:

- lightning needs ice particles (in most cases)
- supercooled water droplets seem to play a role
- reversal temperature depends on liquid water content and relative humidity (Berdeklis and List, 2001)

Research is still going on ...

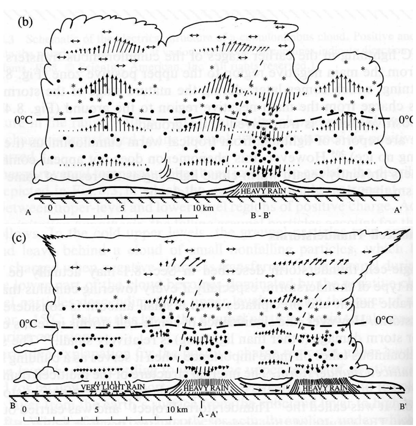
Remark: At the moment there is no (cloud resolving) model containing charged particles (→ solving Maxwell equations ...)

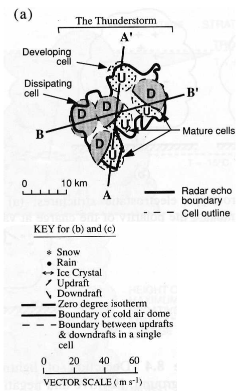
Overview multicell thunderstorms

Characteristics and properties:

- more frequent than single cell storms
- storm consists of a pattern of cells in various development stages (early/mature/dissipation stage)
- different cells can trigger each other → self–organisation
- larger horizontal extension (several tens of km) than single cell storms

Multicell Thunderstorms I





Houze (1993), Fig. 8.5

Multicell Thunderstorms II

Under certain conditions of wind shear the multicell thunderstorm takes on a form of organisation

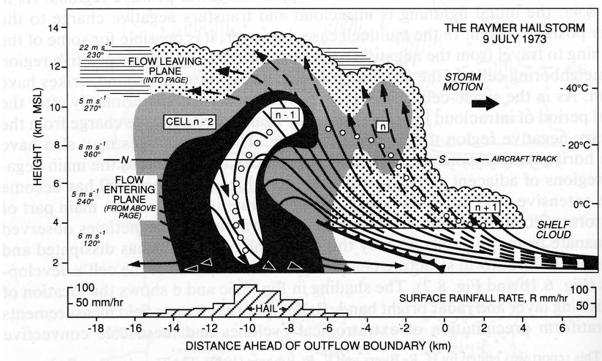


Figure 8.7

Houze (1993), Fig. 8.7

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Multicell Thunderstorms III

Cells:

- **n+1:** developing stage, strong updraft
- **n :** developing stage, collection of ice particles
- **n-1:** downdraft due to precipitation
- **n-2:** dissipating stage (precipitation)

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Supercell Thunderstorms - Overview I

Characteristics:

- same size as multicell thunderstorms
- only single storm–scale circulation of one giant updraft–downdraft pair
- strong vertical updrafts 10 – 40 m/s
- often producing hail
- often transition to tornadic phase → next lecture

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Supercell Thunderstorms - Overview II

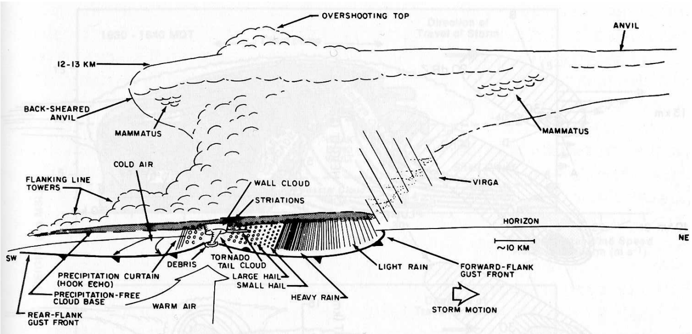


Figure 8.9 Schematic visual appearance of a supercell thunderstorm. (Based on U.S. National Severe Storms Laboratory publications and an unpublished manuscript of Howard B. Bluestein.)

Houze (1993), Fig. 8.9

Supercell Thunderstorms - Overview III

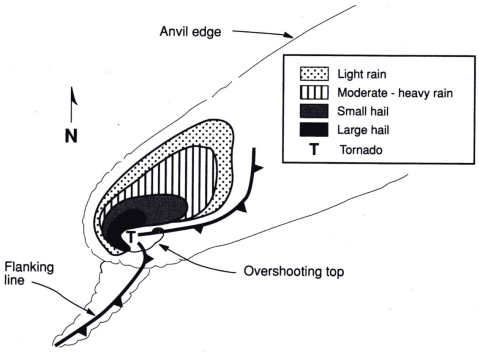
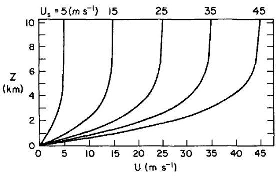


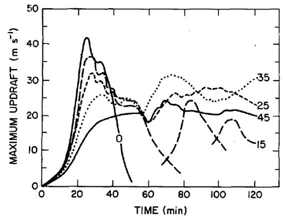
Figure: Idealized view from satellite, Houze (1993), Fig 8.10

Supercell vs. Multicell Thunderstorms

Environmental conditions favouring different types of storms:
Stability and wind shear

Variation of wind shear:





Weisman & Klemp (1982), Figs. 2&3

FIG. 3. Time series of maximum vertical velocities for the U_s = 0, 15, 25, 35 and 45 m s^{-1} wind shear experiments. q_{v0} = 14 g kg^{-1}.

- $u_s = 0$ m/s: single cell storm (convective shower)
- $u_s = 15$ m/s: multicellular storm structure
- $u_s \geq 25$ m/s: supercell dynamics (redevelopment of cells)

Basic equations (neglecting Coriolis term)

Incompressible Boussinesq equations:

$$\frac{D\vec{v}}{Dt} := \underbrace{\frac{\partial \vec{v}}{\partial t}}_{\text{time evolution}} + \underbrace{\vec{v}\cdot\nabla\vec{v}}_{\text{advection}} = -\frac{1}{\rho_0}\nabla p^* + B\vec{k} + \vec{F} \quad (2)$$

$$\nabla\cdot(\rho_0\vec{v}) = 0 \quad (3)$$

$$\text{Vorticity} \quad \omega \equiv \nabla\times\vec{v} = \eta\cdot\vec{i} + \xi\cdot\vec{j} + \zeta\cdot\vec{k} \quad (4)$$

$$\eta = \frac{\partial w}{\partial y} - \frac{\partial v}{\partial x}, \xi = \frac{\partial u}{\partial z} - \frac{\partial w}{\partial x}, \zeta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}$$

Vorticity equation (sources & sinks of vertical vorticity):

$$\frac{D\zeta}{Dt} = \underbrace{\zeta\frac{\partial w}{\partial z}}_{stretching} + \underbrace{\xi\frac{\partial w}{\partial y} + \eta\frac{\partial w}{\partial x}}_{tilting} + \underbrace{F_\zeta}_{mixing} \quad (5)$$

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Recap: Boussinesq approximation

Assumption: Density variations (ρ^*) are only regarded if they give rise to buoyancy forces

$$B = -g\frac{\rho - \rho_0}{\rho_0} = -g\frac{\rho^*}{\rho_0} \quad (6)$$

and they are ignored as they affect the fluid inertia or continuity

$$\nabla\cdot(\rho\vec{v}) = 0 \quad (7)$$

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Environment & initial state

Environmental conditions (horizontal wind $\vec{V}$ and wind shear $\vec{S}$):

$$\vec{V} \equiv U\cdot\vec{i} + V\cdot\vec{j}, \quad \vec{S} \equiv \frac{\partial\vec{V}}{\partial z} = \frac{\partial U}{\partial z}\cdot\vec{i} + \frac{\partial V}{\partial z}\cdot\vec{j}$$

Initial state:

- isolated cumulus in unidirectional wind shear, westerly velocity U increases with height ($\partial U/\partial z > 0$), $V = 0$.
- in early convective growth, the cloud moves with the westerly flow
- low level inflow from east (due to wind shear)
- upper–level outflow towards east (due to wind shear)

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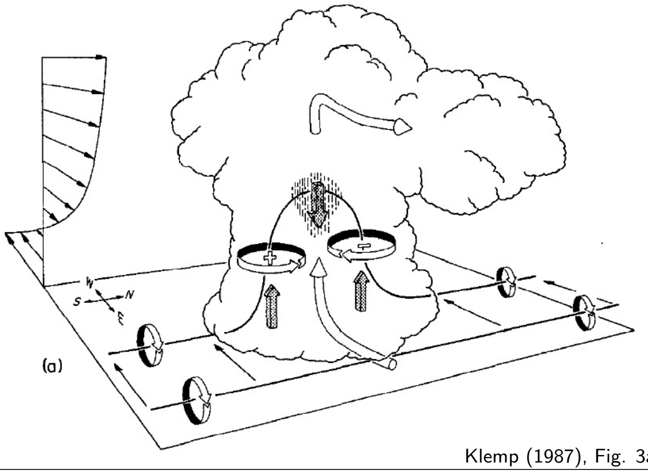
Early convective state



Klemp (1987), Fig. 3a

Circulation I

A vortex circulation develops due to the (linearized) vorticity equation:

$$\frac{D\zeta}{Dt} = \frac{\partial \zeta}{\partial t} + \underbrace{U\frac{\partial \zeta}{\partial x}}_{\text{neglegible}} = \xi \frac{\partial w}{\partial y} \approx \frac{\partial U}{\partial z}\frac{\partial w}{\partial y} \qquad (8)$$

- horizontal vorticity (ξ) generated by wind shear
- vortex tube lifted by convective updraft
- positive (cyclonic) vertical vorticity (ζ) is generated along the southern flank of the updraft ($\partial w/\partial y > 0$)
- negative (anticyclonic) vertical vorticity is generated along the northern flank of the updraft ($\partial w/\partial y < 0$)
- Remark: Entrainment enhanced by wind shear and vortex dynamics

Circulation II

As the cloud develops further, non–linear effects become important:

$$\frac{\partial \zeta}{\partial t} + U\frac{\partial \zeta}{\partial x} = \frac{\partial U}{\partial z}\frac{\partial w}{\partial y} + \underbrace{\zeta\frac{\partial w}{\partial z}}_{\text{stretching}} \qquad (9)$$

i.e. stretching of the vortex tube becomes important
→ enhancement of vorticity
two possibilities for the development of the storm:

1. dissipation of the storm
2. splitting and intensification of the storm (supercell thunderstorm)

Factors for splitting the storm:

- storm–relative low–level inflow from the east: preventing the cold air from moving out ahead of the storm
- lifting vertical pressure gradients (more important, see below)

Storm Splitting I

Momentum equations for the splitting stage:

$$\frac{\partial \vec{v}}{\partial t} = -\frac{1}{\rho_o}\nabla p^* + B\vec{k} - \vec{v}\cdot\nabla\vec{v} \tag{10}$$

multiplying by ρ_0, taking the 3D divergence, using the Boussinesq condition (as in last lecture, slide 7) and splitting $p^* = p_D^* + p_B^*$ we get:

$$\nabla^2 p_D^* = \Delta p_D^* \quad = \quad -\nabla\cdot(\rho_o \vec{v}\cdot\nabla\vec{v}) \text{ (dynamics)} \tag{11}$$

$$\nabla^2 p_B^* = \Delta p_B^* \quad = \quad \frac{\partial}{\partial z}(\rho_o B) \text{ (buoyancy)} \tag{12}$$

Storm Splitting II

Only regard the dynamical sources of pressure perturbations:

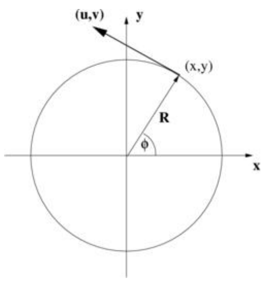
$$\Delta p_D^* = \underbrace{-\rho_0\left[\left(\frac{\partial u}{\partial x}\right)^2 + \left(\frac{\partial v}{\partial y}\right)^2 + \left(\frac{\partial w}{\partial z}\right)^2 - \frac{d^2 \log \rho_0}{dz^2} w^2\right]}_{fluid\ extension}$$

$$\underbrace{-2\rho_0\left[\frac{\partial v}{\partial x}\frac{\partial u}{\partial y} + \frac{\partial u}{\partial z}\frac{\partial w}{\partial x} + \frac{\partial v}{\partial z}\frac{\partial w}{\partial y}\right]}_{fluid\ shear} \tag{13}$$

Investigation of the shear term $\frac{\partial v}{\partial x}\frac{\partial u}{\partial y}$
Case of axis–symmetrical rotation: $\frac{\partial v}{\partial x} = -\frac{\partial u}{\partial y}$

$$\Rightarrow \frac{\partial v}{\partial x}\cdot\frac{\partial u}{\partial y} = -\left(\frac{\partial v}{\partial x}\right)^2 = -\frac{1}{4}\zeta^2$$

Axis–symmetrical rotation



$x = R\sin\varphi$, $y = R\cos\varphi$
Rotation with constant velocity V at radius R:
$u = -V\sin\varphi = -V\frac{y}{R}$, $v = V\cos\varphi = V\frac{x}{R}$

$$\Rightarrow \frac{\partial v}{\partial x} = -\frac{\partial u}{\partial y}$$

Storm Splitting III

Simplified assumption: in the interiour of a flow $\Delta\psi \propto -\psi$

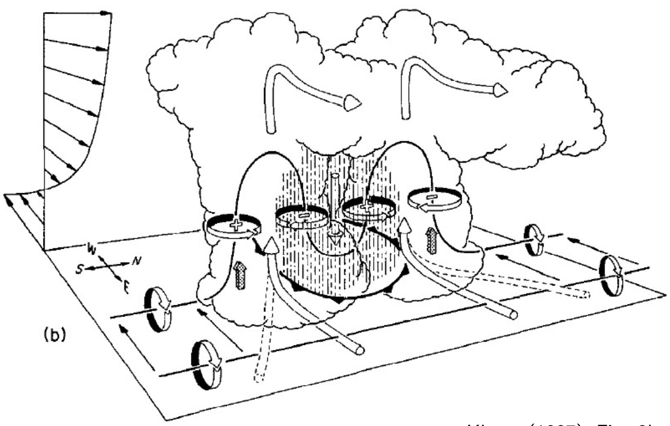
$$p_D^* \propto -\Delta p_D^* \propto \zeta^2 \tag{14}$$

Interpretation:

- dynamic pressure perturbation minimum associated with vortex
- strong midlevel rotation at the updraft flanks acts to lower pressure $\longrightarrow$ inducing updraft growth (lifting pressure gradient)
- second and third term in the fluid shear term of eq. 13 also contribute to lowering the pressure at the flanks
- fluid extension terms do not contribute to the lifting pressure gradients

As the splitting progresses and the two updraft centres move apart, the downdraft tilts the vortex lines downward
$\longrightarrow$ two vortex–pair circulations

Storm Splitting IV



Klemp (1987), Fig. 3b

Storm propagation

At this stage, each of the two vortex pairs will propagate transversely to the mean wind shear S

Remark: We will consider storm propagation more detailled within the next lecture.

Considering a steady updraft propagating transversely to wind shear (i.e. to the south with velocity v_c):

- in a coordinate framework relative to the moving updraft: flow approaches from the south with constant velocity $-v_c$
- linearized vorticity equation becomes

$$-v_c\frac{\partial\zeta}{\partial y} = \frac{\partial U}{\partial z}\frac{\partial w}{\partial y} \tag{15}$$

- this equation can be integrated, yielding: $\zeta = \frac{\partial U}{\partial z}\frac{w}{-v_c}$

Interpretation: Vertical vorticity is coincident with vertical velocity

Preferential enhancement of cyclonically rotating storms

From above we get two different types of supercell thunderstorms:

- right–moving cyclonally rotating storms
- left–moving anticyclonally rotating storms

Data investigations show that most of the storms are right–moving cyclonally rotating storms. WHY?

Cyclonically rotating storms – simple model

updraft perturbation in homogeneous fluid (i.e. no density perturbations → no buoyancy forces)
equations for vertical momentum and pressure perturbations (only dynamical sources):

$$\frac{Dw}{Dt} = -\frac{1}{\rho_o}\frac{\partial p_D^*}{\partial z} \qquad (16)$$

$$\Delta p_D^* = -2\rho_o \left(\underbrace{\frac{\partial U}{\partial z}\vec{i} + \frac{\partial V}{\partial z}\vec{j}}_{\vec{S}} \right) \cdot \nabla_h w \qquad (17)$$

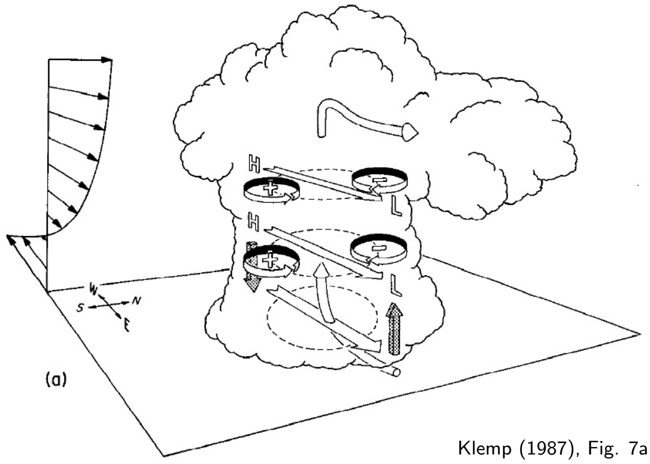
whereas $\nabla_h \equiv \partial/\partial x \cdot \vec{i} + \partial/\partial y \cdot \vec{j}$
We assume again $\Delta p_D^* \propto -p_D^*$ and get:

$$p_D^* \propto -\Delta p_D^* = 2\rho_0 \vec{S} \cdot \nabla_h w \qquad (18)$$

Unidirectional shear I

- shear produces high pressure on upshear (west) side of the updraft
- shear produces low pressure on downshear (east) side of the updraft
- this induces low–level lifting on downshear side → reinforcing storm inflow
- **no contribution to preferential growth on a flank**

Unidirectional shear II



Klemp (1987), Fig. 7a

Unidirectional shear III

This pressure effect increases in amplitude with height beneath the level of the maximum updraft velocity

simplest case: linear shear:

$$\vec{V} = (U, V) = (U, 0) = (U_o + S_x \cdot z, 0)$$

$$\vec{S} = \left(\frac{\partial U}{\partial z}, \frac{\partial V}{\partial z}\right) = (S_x, 0)$$

$$\Rightarrow \vec{S} \cdot \nabla_h w = \frac{\partial U}{\partial z}\frac{\partial w}{\partial x} = S_x \frac{\partial w}{\partial x}$$

only in x–direction (West–East–direction)

Turning shear I

- turning shear (clockwise with height) produces pressure gradients that favours ascent on the southern flank
- turning shear (clockwise with height) produces pressure gradients that favours descent on the northern flank

→ enhancement of development of right–moving storm
→ suppressing of development of left–moving storm

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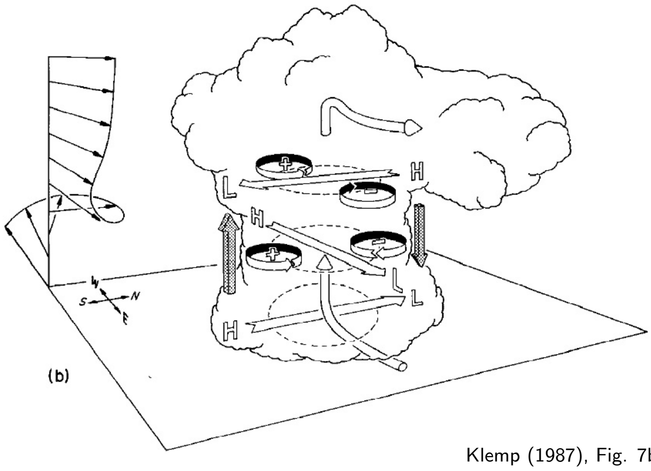
Turning shear II



Klemp (1987), Fig. 7b

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Turning shear III

Simple case:
Constant velocity $|\vec{V}| = V_o$, wind turns clockwise until it reaches level $z = z_o$

$$U = -V_o \cdot \cos(\varphi), V = -V_o \cdot \sin(\varphi), \varphi = 2\pi - \pi\frac{z}{z_o}$$

$$U(0) = -V_o, V(0) = 0; U(z_o) = V_o, V(z_o) = 0$$

$$\frac{\partial U}{\partial z} = -V_o\frac{\pi}{z_o}\sin(\varphi), \frac{\partial V}{\partial z} = V_o\frac{\pi}{z_o}\cos(\varphi)$$

$$\frac{\partial U}{\partial z}(0) = 0, \frac{\partial U}{\partial z}(z_o) = 0; \frac{\partial V}{\partial z}(0) = V_o\frac{\pi}{z_o}, \frac{\partial V}{\partial z}(z_o) = -V_o\frac{\pi}{z_o}$$