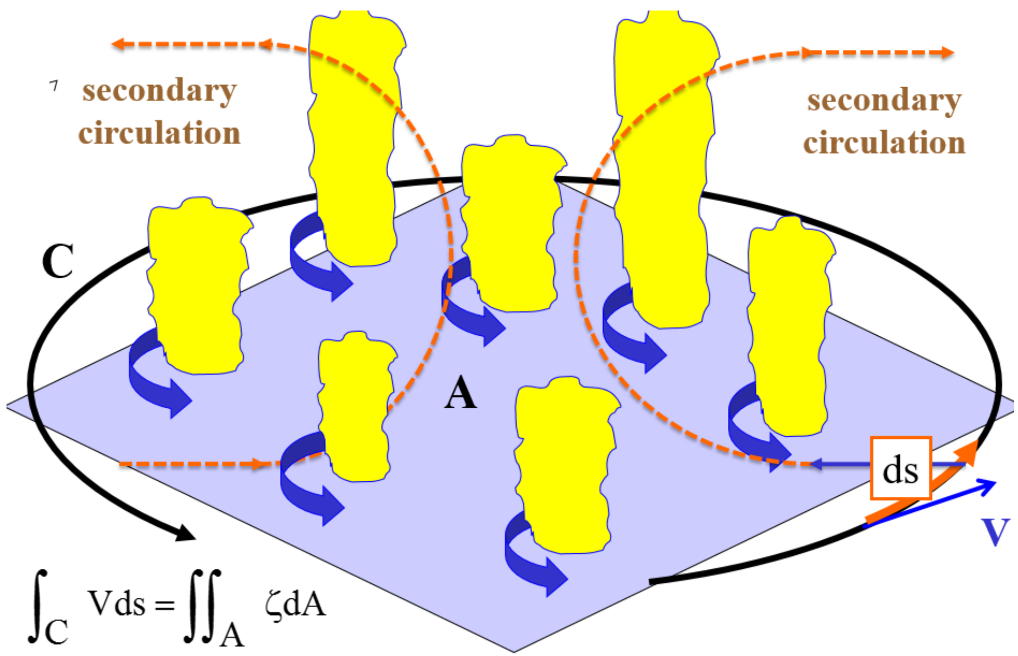
Rotating Convection During tropical Cyclogenesis

Gerard Kilroy

"Although some aspects of the transformation of atmospheric disturbances into tropical cyclones are relatively well understood, the general problem of tropical cyclogenesis remains, in large measure, one of the great mysteries of the tropical atmosphere."

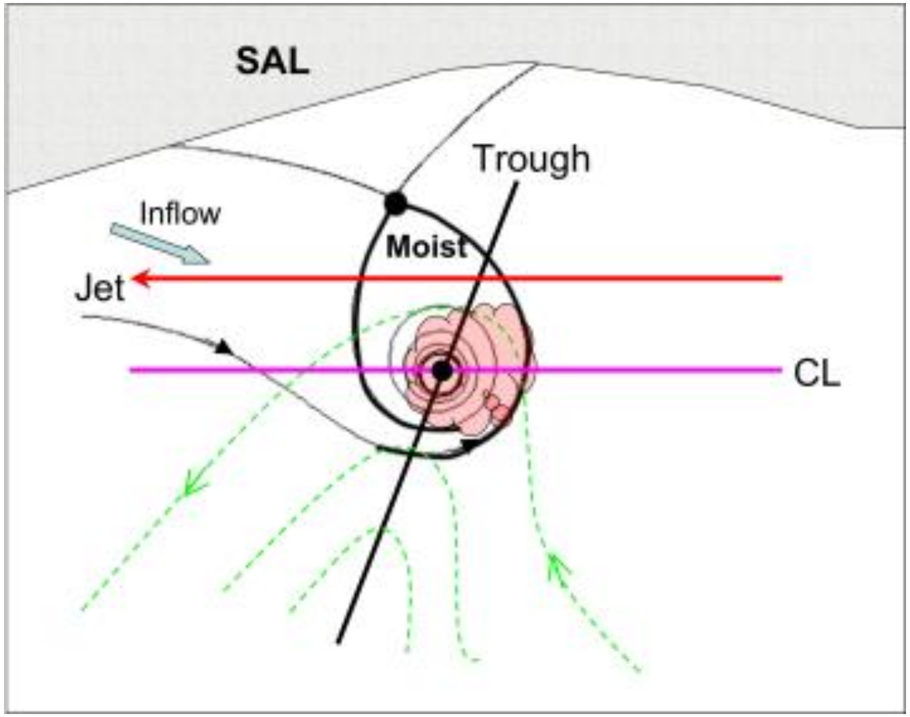
Kerry Emanuel, *Divine Wind*

- PREDICT Experiment 2010 (PRE-Depression Investigation of Cloud-systems in the Tropics)
- "**Marsupial paradigm**" for cyclogenesis (Dunkerton et al. 2009) and a related paradigm for tropical cyclone intensification (Montgomery and Smith 2014)
- This paradigm proposes that hurricanes form in a "pouch", consisting of a closed cyclonic circulation



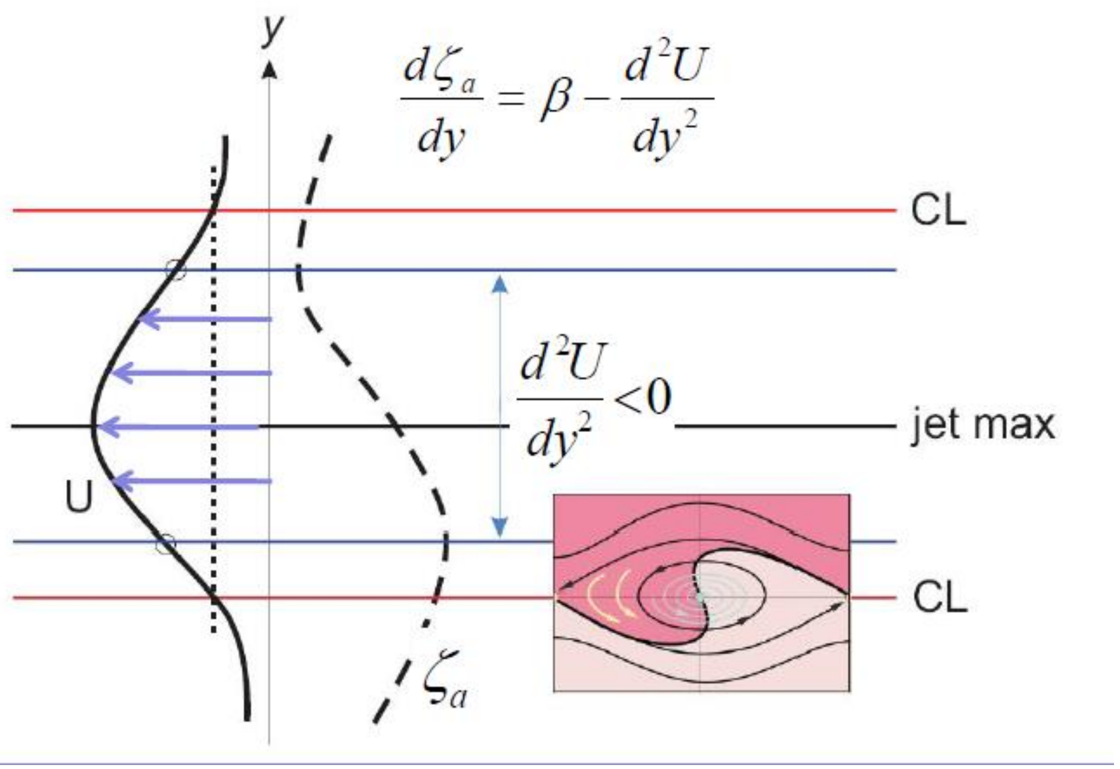
Marsupials are mammals in which the female typically has a **pouch** in which it rears its young through early infancy.

The hypothetical pathway for genesis via tropical waves may be regarded as a marsupial theory of tropical cyclogenesis in which the “young” proto-vortex is carried along by the “mother” wave until it is ready to be “let go” as an independent tropical disturbance.



- Hypothesis 1: Wave breaking or roll-up of the cyclonic vorticity in the lower troposphere provides a favored region for the aggregation of vorticity seedlings and TC formation;
- Hypothesis 2: The wave critical layer is a region of closed circulation, where air is repeatedly moistened by convection and protected from dry air intrusion;
- Hypothesis 3: The parent wave is maintained and possibly enhanced by diabatically amplified mesoscale vortices within the wave.

Dunkerton *etal*. (2009)



- Idealized profile of an easterly jet U north of the equator, on which stable easterly waves propagate westward having their critical latitudes just outside the jet's inflection points where the gradient of absolute vorticity or "effective " β is zero.

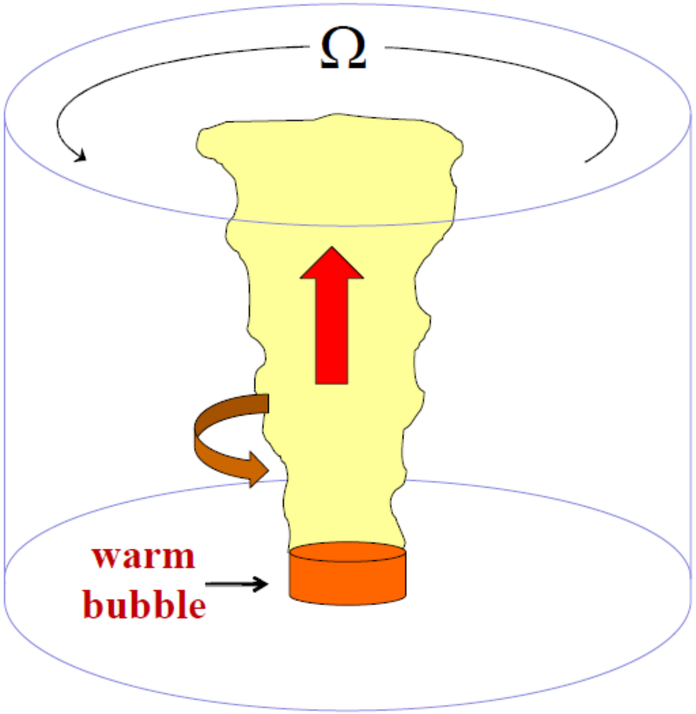
- Unstable waves may be imagined to have their critical latitudes just inside the latitude of inflection points.

The pouch protects and encourages development

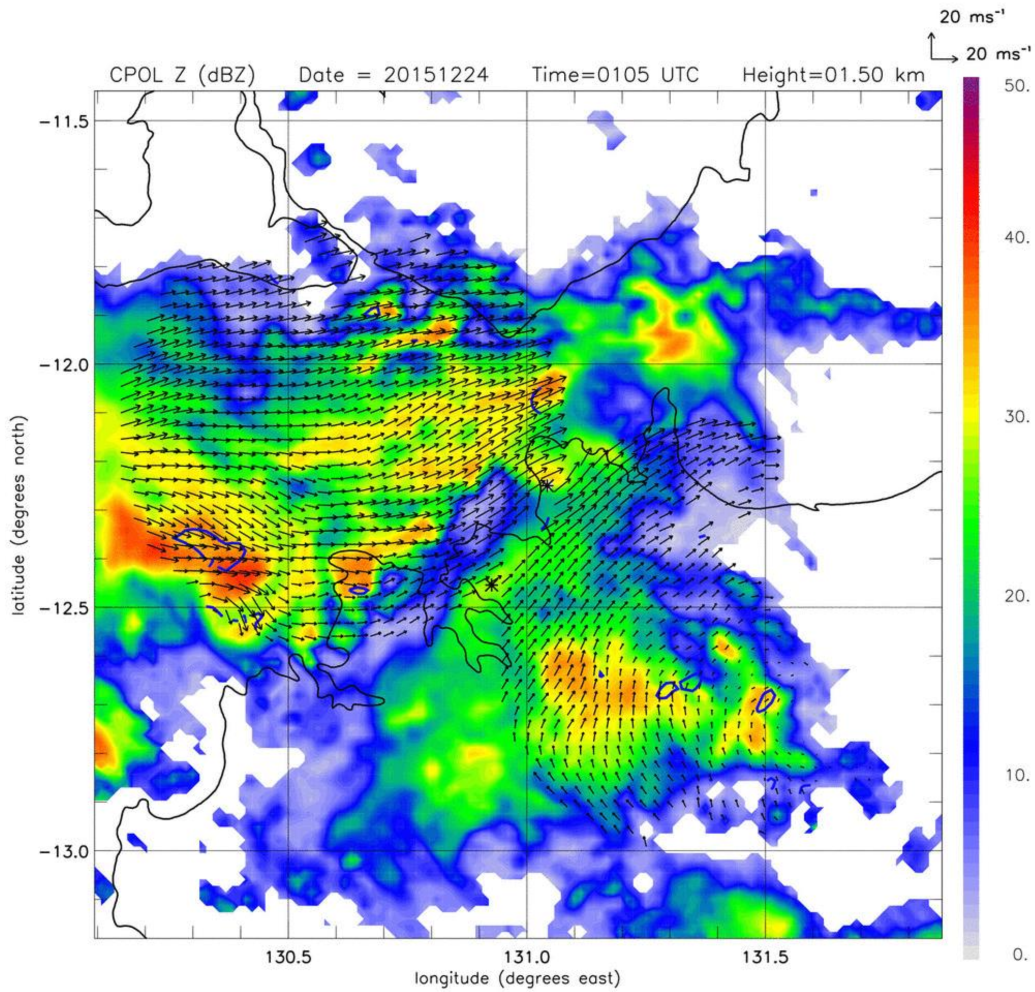
- Dry & dusty air (e.g., Saharan Air Layer) stifles development.
- Large vertical shear is generally detrimental to genesis.
- Cyclonic vorticity resides in the wave critical layer equator-ward of the easterly jet axis.
- This region is characterized by strong rotation and weak straining deformation. (Okubo-Weiss parameter)
- Vorticity and (if available) ambient moisture are entrained into the gyre and additional moistening by deep convection within the gyre will tend to remain in the gyre.
- In addition to its cyclonic vorticity the gyre provides a strong focus for spontaneous aggregation of vortical building-blocks on the mesoscale, and segregation of cyclonic & anticyclonic anomalies.
- Preliminary evidence suggests a predominantly convective, as opposed to stratiform, vertical profile of heating in the gyre.

“Vortical hot towers” (VHTs)

VHTs have been invoked as the fundamental coherent structures in the tropical cyclone genesis process (Hendricks et al. 2004, Montgomery et al. 2006, Braun et al. 2010) and the tropical cyclone intensification process (Nguyen et al. 2008, Shin and Smith 2008, Montgomery et al. 2009).



VHTs dominate the intensification period at early times and that the progressive aggregation, merger and axisymmetrization of the VHTs together with the low-level convergence they generate are prominent features of the tropical cyclone intensification process.


20 ms⁻¹
20 ms⁻¹
CPOL Z (dBZ)
Date = 20151224
Time=0105 UTC
Height=01.50 km
-11.5
-12.0
-12.5
-13.0
latitude (degrees north)
130.5
131.0
131.5
longitude (degrees east)
50.
40.
30.
20.
10.
0.

Dunkerton et al. (2009):

VHTs are deep moist convective clouds that rotate as an entity and/or contain updrafts that rotate.

These **hot vortical plume structures** amplify pre-existing cyclonic vorticity by low– to midlevel vortex tube stretching and generate local enhancements of cyclonic vorticity above that of the aggregate vortex.

Although early observations suggested that VHTs are neither necessary nor sufficient for tropical cyclogenesis, it is becoming clear from cloud representing numerical simulations that moist vortical updrafts are the essential building blocks of the tropical storm.

Theories for tropical cyclone intensification and structure

- CISK (Charney and Eliassen 1964)
- Cooperative Intensification Theory (Ooyama 1969).
- WISHE (Emanuel 1986 … , Holton and Hakim, 2012)
- Vortical deep convection paradigm whose mean field view provides an extended Cooperative Intensification Theory

The Bruce Morton Memorial Volume of the Australian Meteorological and Oceanographical Society Journal
http://www.meteo.physik.uni-muenchen.de/~roger/Publications/M8.pdf

Paradigms for tropical cyclone intensification

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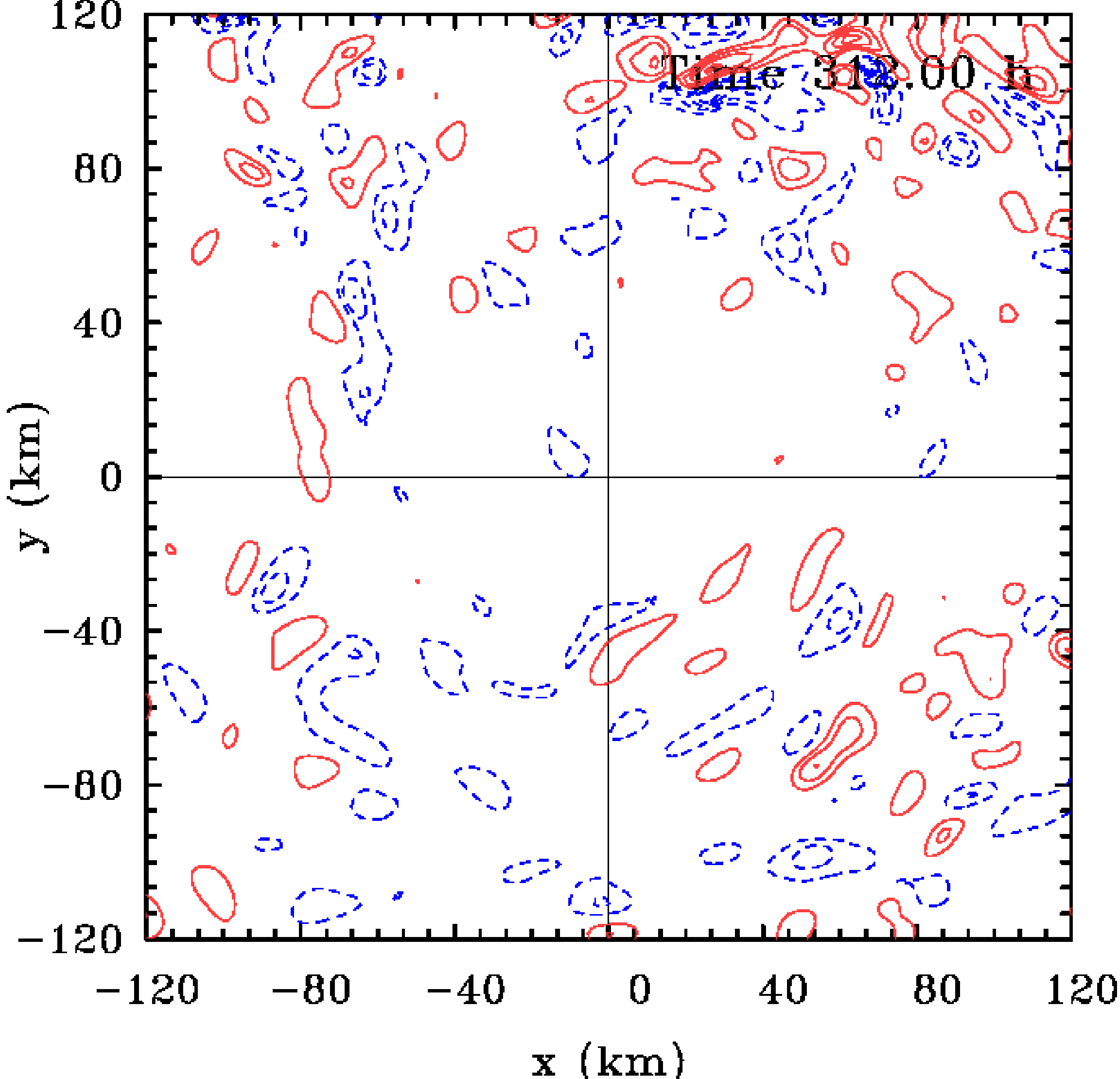
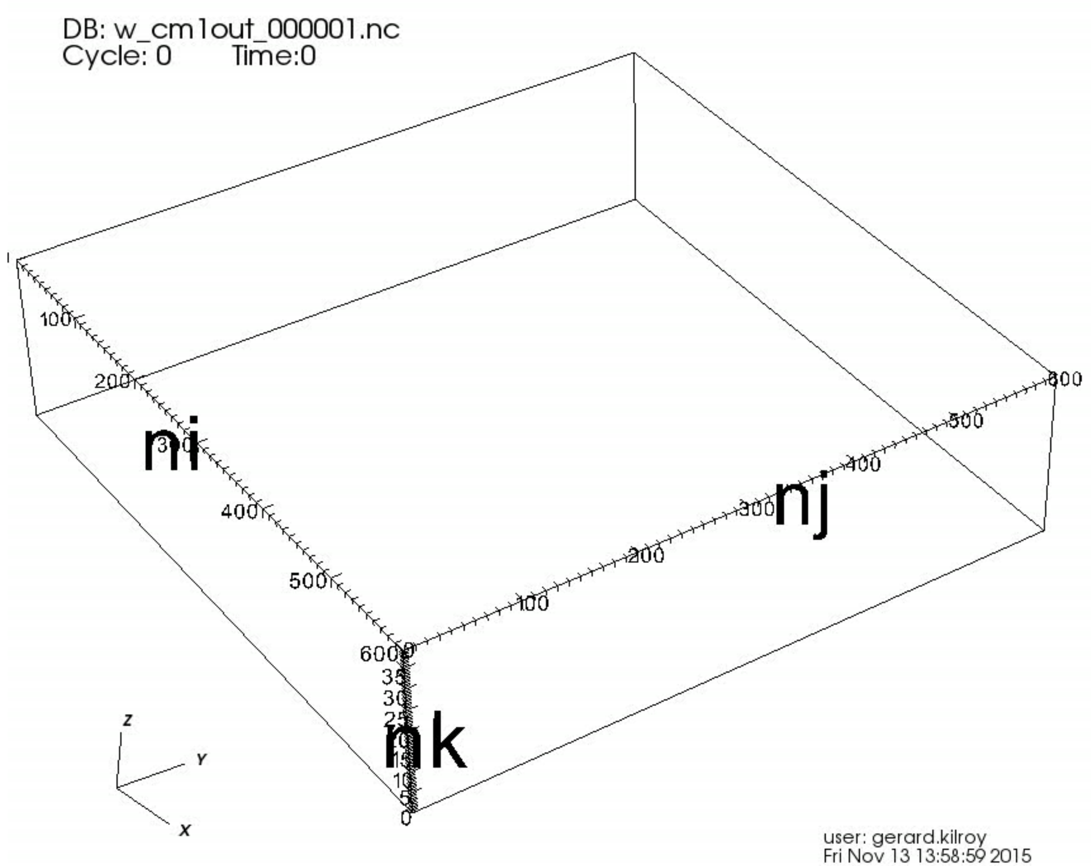
- Patches of enhanced vertical rotation from convective cells can interact and merge
- Wissmeier and Smith (2011) showed all convective clouds that form away from the equator have some vertical rotation **which outlives the cloud**
- Animation: an idealized MM5 simulation that we performed that captures a tropical cyclone genesis event.

Photograph: Roger K. Smith



Time 312.00 h
y (km)
x (km)
120
80
40
0
−40
−80
−120
−120
−80
−40
0
40
80
120

Genesis in 3D



Barotropic vortex merger

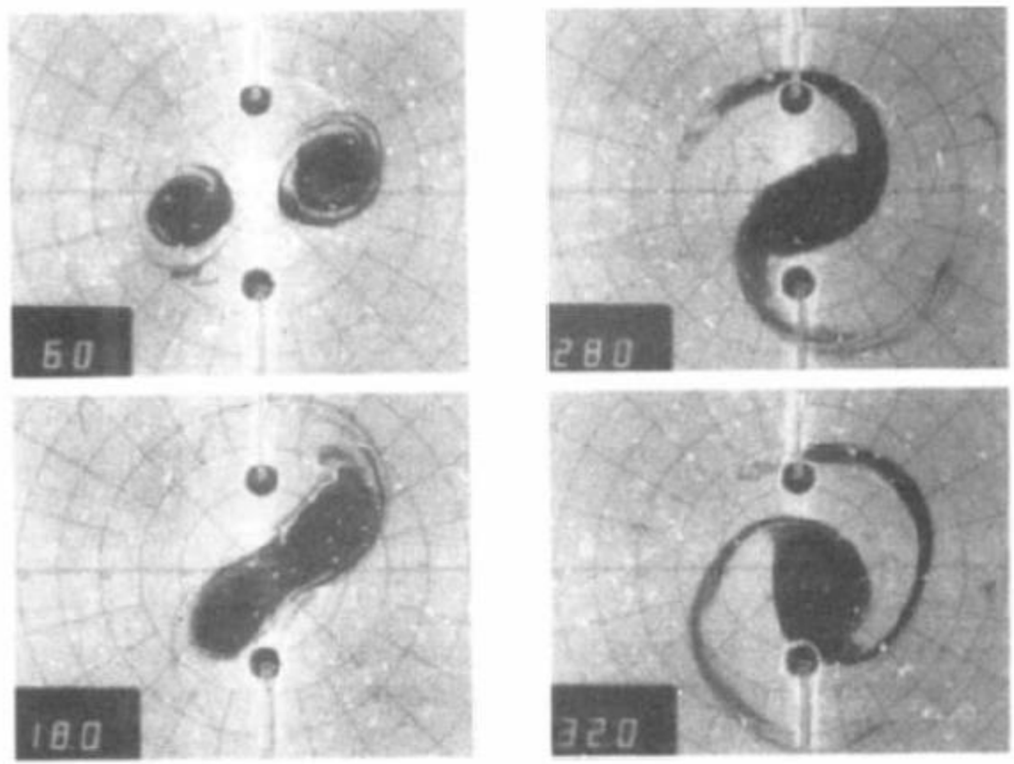


Figure 9 Photographs of two merging anticyclones in the top layer: $d = 18$ cm, $\Lambda = 10$ cm, $R' = 4.0$ cm, $s = 7.4$ cm^2 s^{-1}. The numbers indicate the elapsed orbital time in rotation periods. The orbital time is $44T_\Omega$.

In order to understand the merger process, we must first understand how vertical vorticity is generated in TC environments

Use idealized numerical model experiments with PREDICT soundings

The vorticity, or a measure of the local rotation, is the curl of the velocity vector $\mathbf{v}$:

$$\omega = (\xi, \eta, \zeta) = \nabla \times \mathbf{v},$$

$$= \left(\frac{\partial w}{\partial y} - \frac{\partial v}{\partial z}\right)\mathbf{i} + \left(\frac{\partial u}{\partial z} - \frac{\partial w}{\partial x}\right)\mathbf{j} + \left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}\right)\mathbf{k}$$

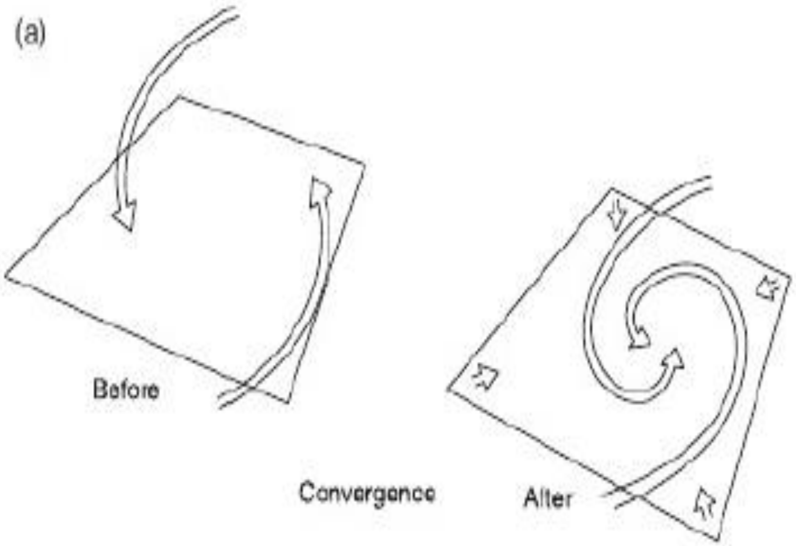
The time evolution of the absolute vertical vorticity

$$\frac{\partial \zeta_a}{\partial t} = \overbrace{-\mathbf{v}_h \cdot \nabla \zeta_a - w\frac{\partial \zeta}{\partial z}}^{\text{advective}} \overbrace{-\zeta_a \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right)}^{\text{stretching}} + \overbrace{\left(\frac{\partial w}{\partial y}\frac{\partial u}{\partial z} - \frac{\partial w}{\partial x}\frac{\partial v}{\partial z}\right)}^{\text{tilting}}$$
$$+ \underbrace{\frac{1}{\rho^2}\left(\frac{\partial \rho}{\partial x}\frac{\partial p}{\partial y} - \frac{\partial \rho}{\partial y}\frac{\partial p}{\partial x}\right)}_{\text{solenoidal}}$$

$$\frac{\partial v}{\partial x} = \Omega; \quad \frac{\partial u}{\partial y} = -\Omega; \quad \omega_z = 2\Omega$$

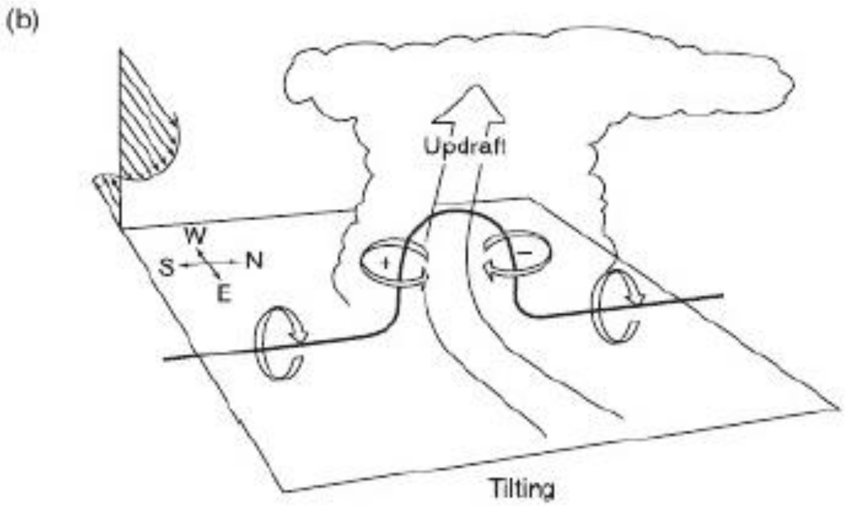
Can think of the vertical vorticity as twice the local angular velocity

Stretching



Cotton *et al.* (2011)

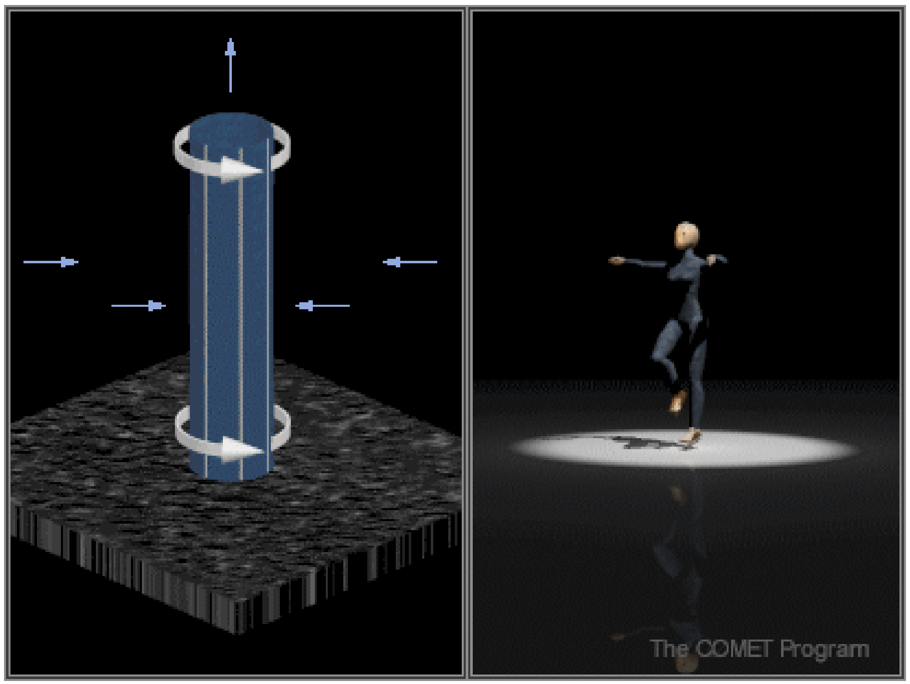
Tilting



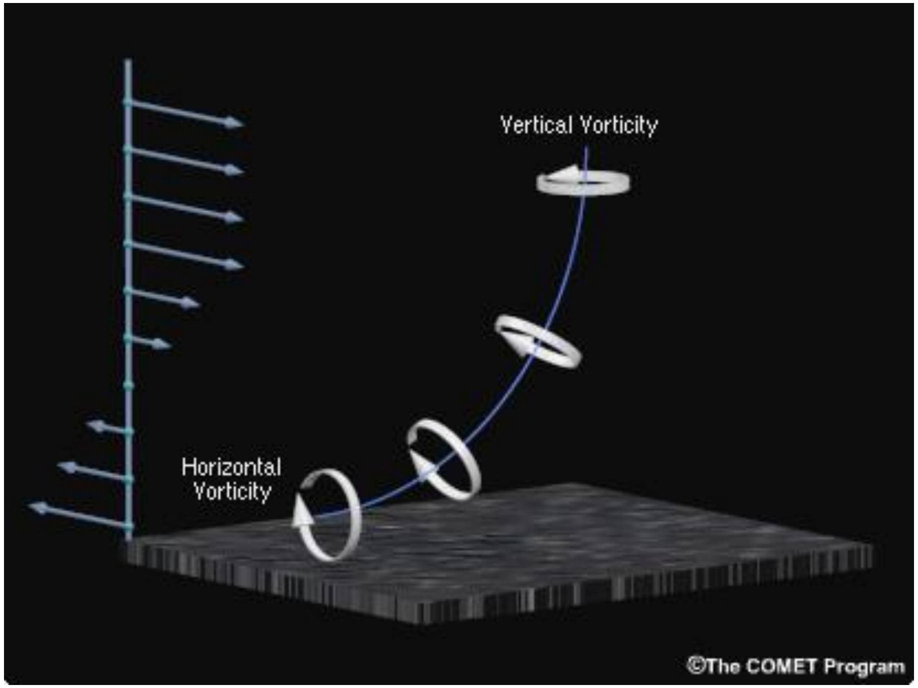
The environmental vertical shear is given by

$$\mathbf{S} = \frac{d\mathbf{v}_h}{dz} = \left(\frac{du}{dz}, \frac{dv}{dz}\right)$$

Stretching



Tilting



On the Evolution of Thunderstorm Rotation

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(Manuscript received 24 August 1980, in final form 21 November 1980)

The Dependence of Numerically Simulated Convective Storms on Vertical Wind Shear and Buoyancy

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(Manuscript received 9 October 1981, in final form 2 February 1982)

The Structure and Classification of Numerically Simulated Convective Storms in Directionally Varying Wind Shears

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(Manuscript received 16 April 1984, in final form 7 September 1984)

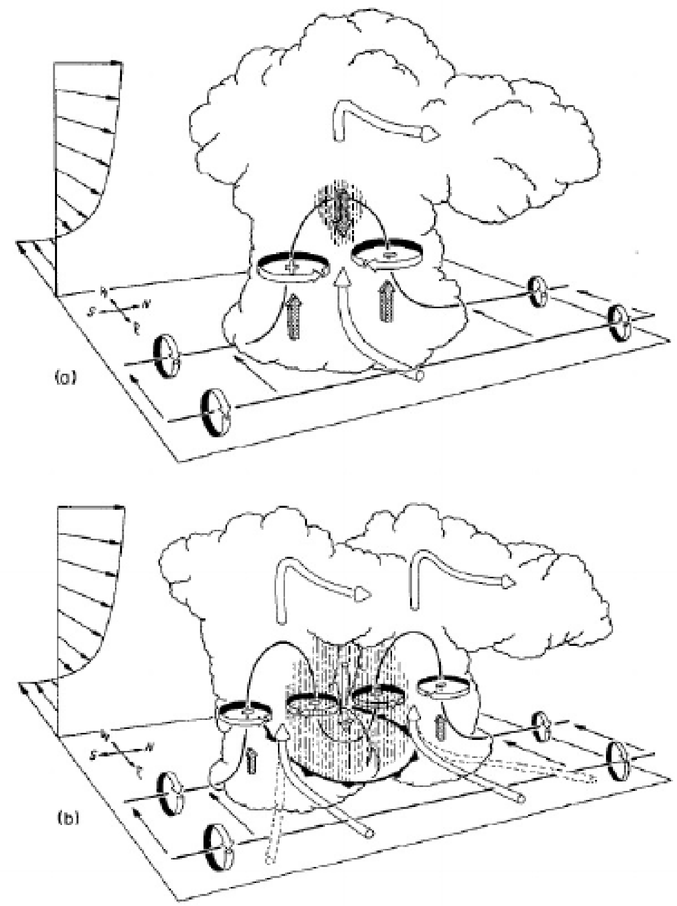
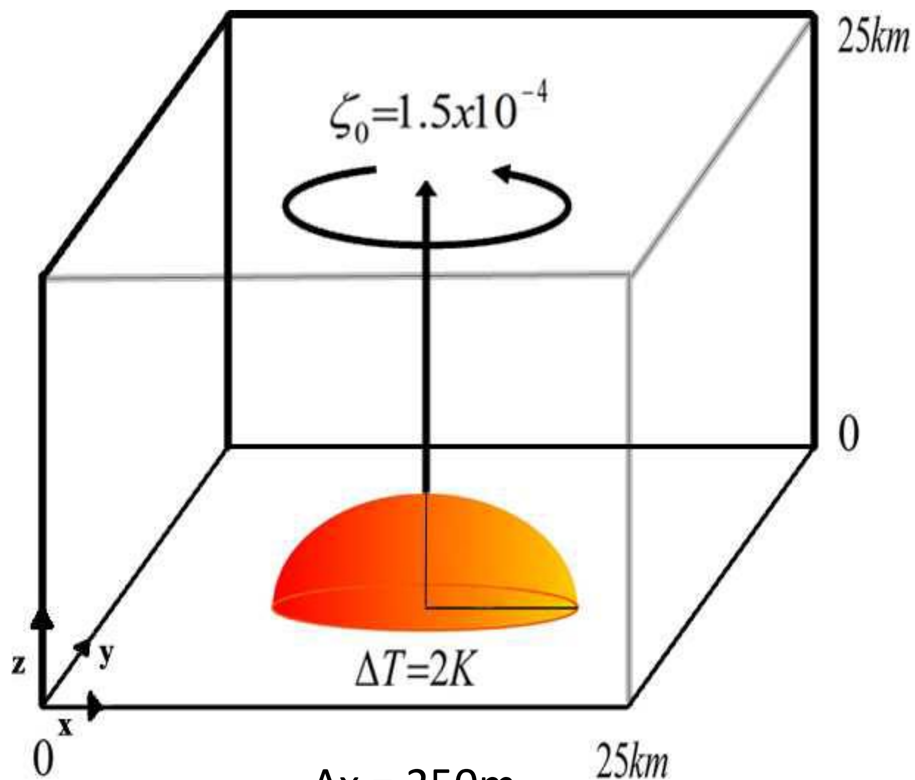


Figure 1.4: A schematic, adapted from Klemp (1987), which illustrates how vortex tubes are tilted in an environment with uni-directional vertical wind shear. The horizontal vorticity tubes (solid black lines) are tilted into the vertical by (a) the growing convective cell to form a vertical vorticity dipole. The rain induced downdraught (b) then tilts the horizontal vorticity downward and aids updraught splitting into two different cells. The white arrows show the direction of the inflow and outflow associated with the cell, while the shaded arrows represent the pressure gradient forces. The dash-dotted lines indicate regions of precipitation. Klemps schematic appears to ignore the effects of vertical shear in *slanting* the cell, which would cause the rain to fall out ahead of the updraught.

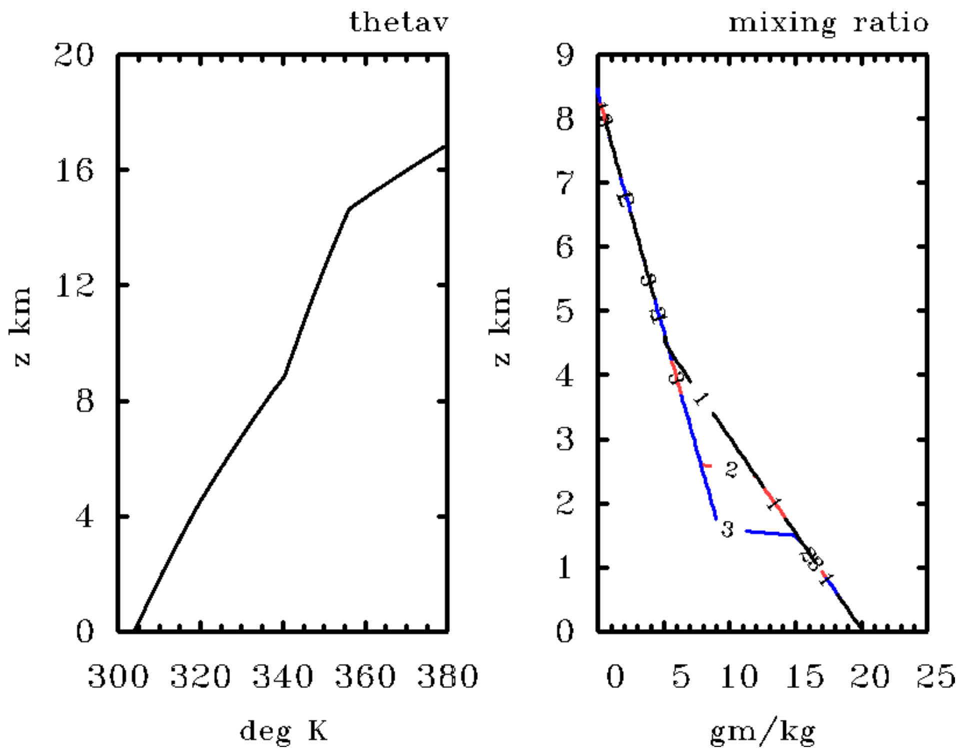
- George Bryan’s cloud model
- Three-dimensional, non-hydrostatic, non-linear, time-dependent numerical model
- Kessler/Gilmore microphysics scheme
- “Open” boundary conditions at lateral boundaries
- Integration time 2 hours
- No radiation, no surface fluxes, no friction



Δx = 250m

- Common perception among forecasters is that dry air aloft leads to stronger downdraughts
- PREDICT forecasters thought that the failure of ex-Tropical Storm Gaston to redevelop was because a dry Saharan air layer aloft suppressed the system, flooding the BL with cool downdraught air
- James and Markowski (2009) show dry air aloft exerts detrimental effects on convection, weakening updraughts and downdraughts
- Key question: How does dry air aloft affect vorticity generation?

- Idealised soundings approximating that from PREDICT on 5 September, 18:20 UTC.



No horiz shear

No vert shear

$f : \zeta_0 = 1.5 \times 10^{-4}\ s^{-1}$

CAPE: 2770 $J\ kg^{-1}$

CIN: 40 $J\ kg^{-1}$

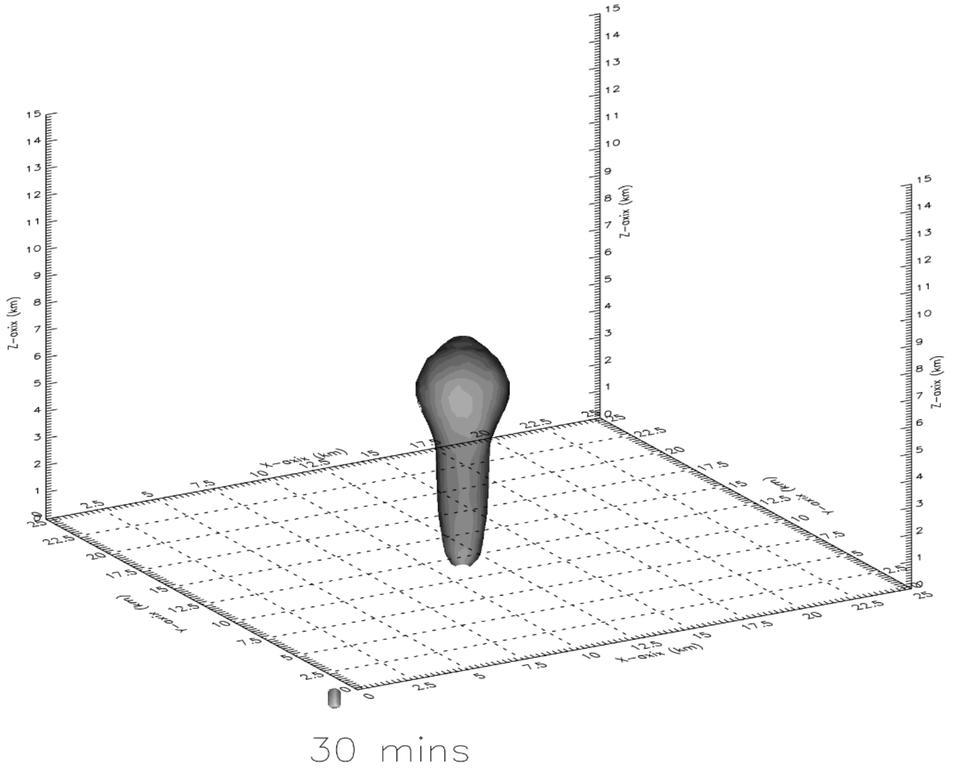
TPW Exp. 1: 62.3 $kg\ m^{-2}$

TPW Exp. 2: 59.3 $kg\ m^{-2}$

TPW Exp. 3: 54.8 $kg\ m^{-2}$

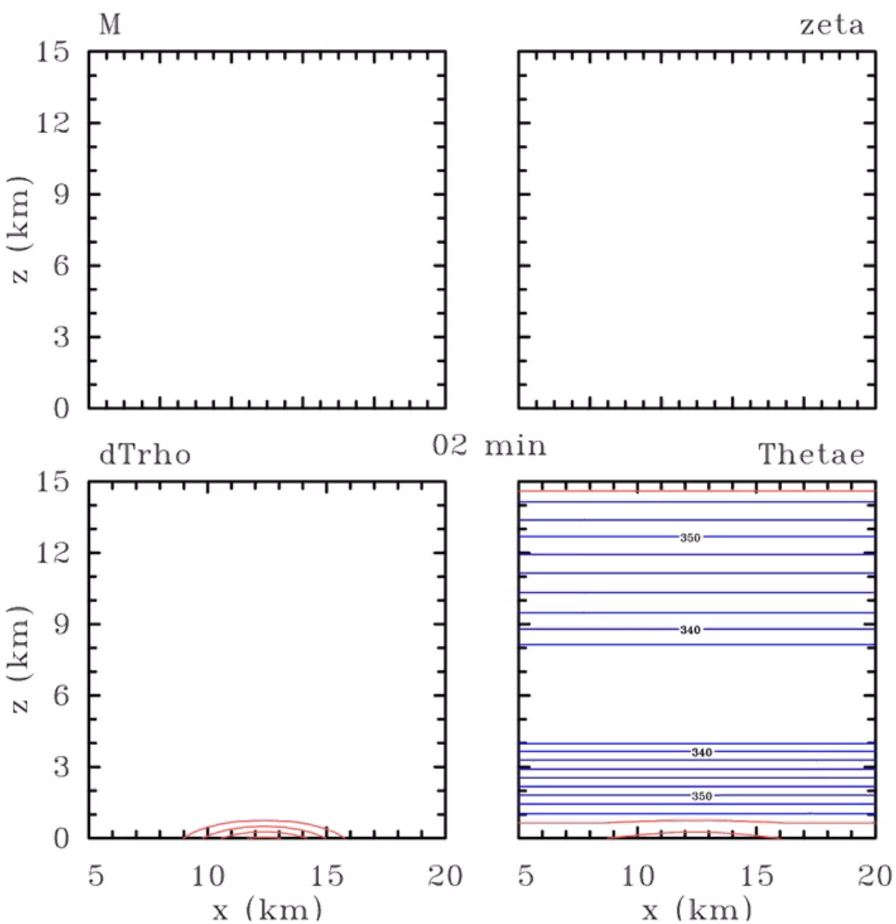
Thermal bubble: 2 K

3D cloud evolution



- Positive massflux
- Negative massflux
- Rain contours
- Cloud condensate

- Positive zeta
- Negative zeta



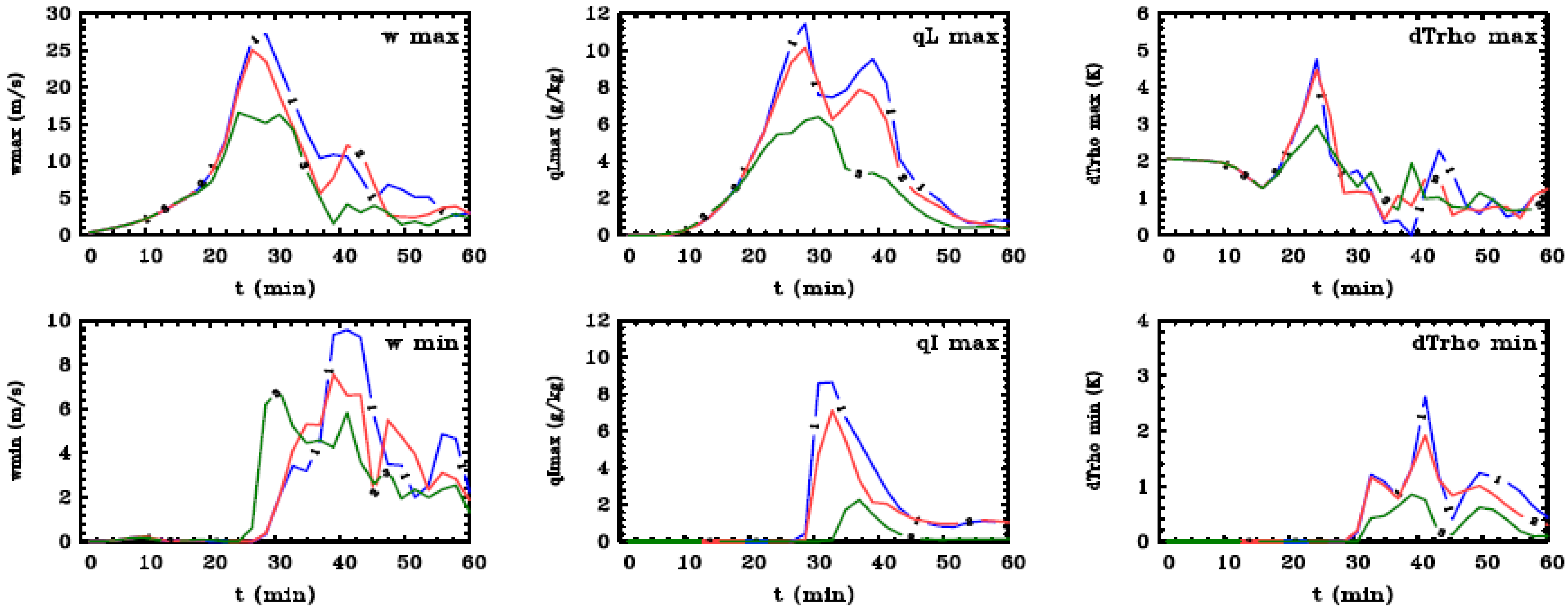
- Positive dTrho
- Negative dTrho
- Rain contours
- Ice particles

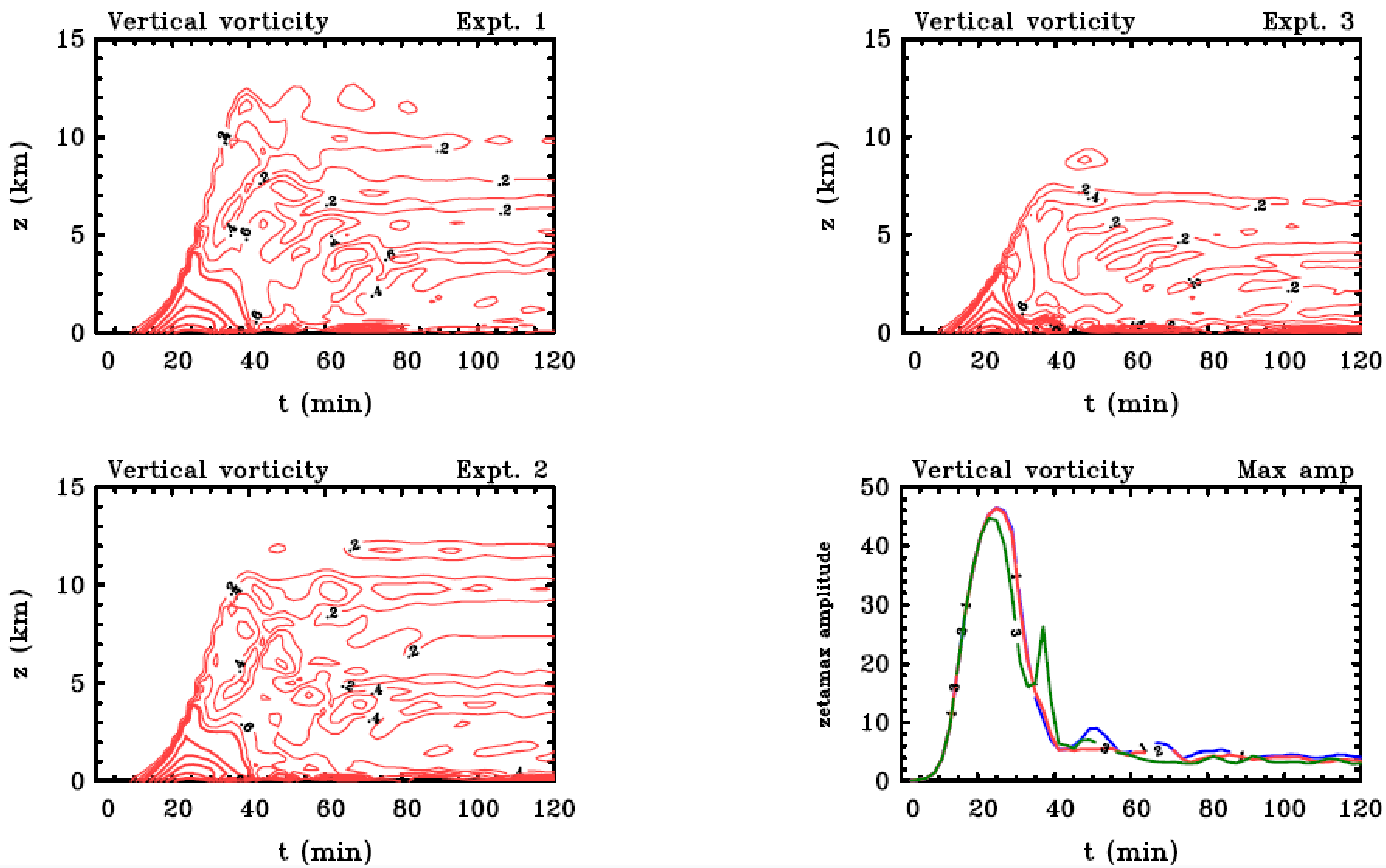
- Thetae > 354 K
- Thetae ⩽ 354 K
- Rain contours
- Cloud condensate

Dry air aloft

Evolution of maximum & minimum values

Expt. 1 Moist → **Expt. 2** Drier → **Expt. 3** Driest





Dry air aloft

Dry air aloft: conclusions

- Entrainment of dry air aloft weakens convective updraughts and downdraughts
- Dry air aloft may have weakened Gaston (2010) by weakening overall convection, not by strengthening convective downdraughts
- Convective cells amplify locally the ambient rotation by more than an order of magnitude at low levels
- This vorticity persists long after the initial updraught has decayed
- The maximum amplification of vorticity is insensitive to the presence of dry air aloft
- Dry air does reduce the depth to which there is significant amplification of vorticity

Uni-directional vertical shear with and without a boundary layer

- Mid-latitude numerical model studies typically have horizontal winds increasing in strength with height (Wilhelmson & Klemp 1978, Weisman & Klemp 1982 ...)
- Single sign of horizontal vorticity throughout the atmosphere
- In tropical depressions the wind speed decreases with height above a shallow BL (~ 1 km)
- The horizontal vorticity changes sign above the BL
- **Key question: How does a BL affect vertical vorticity generation?**

Weisman & Klemp 1982

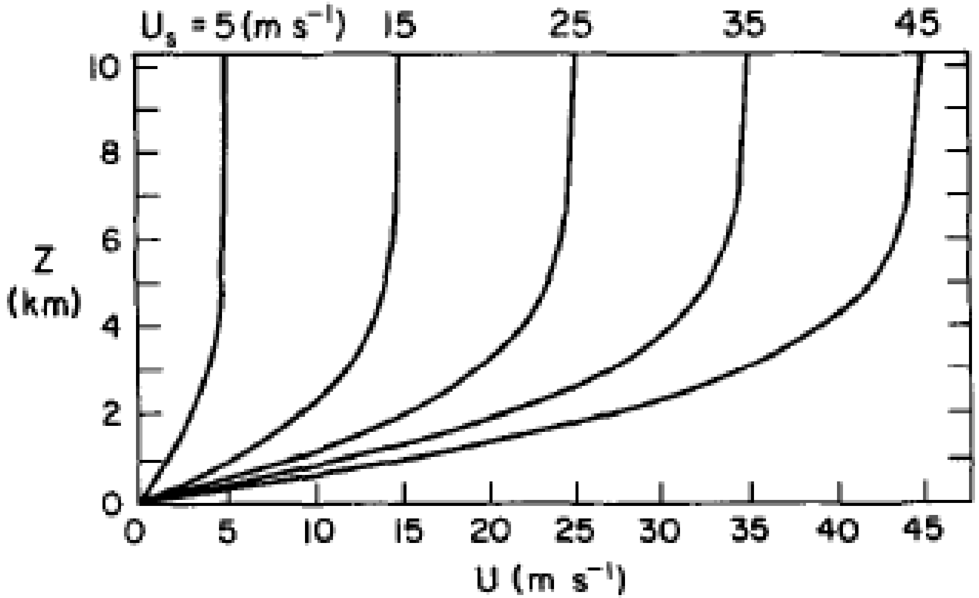
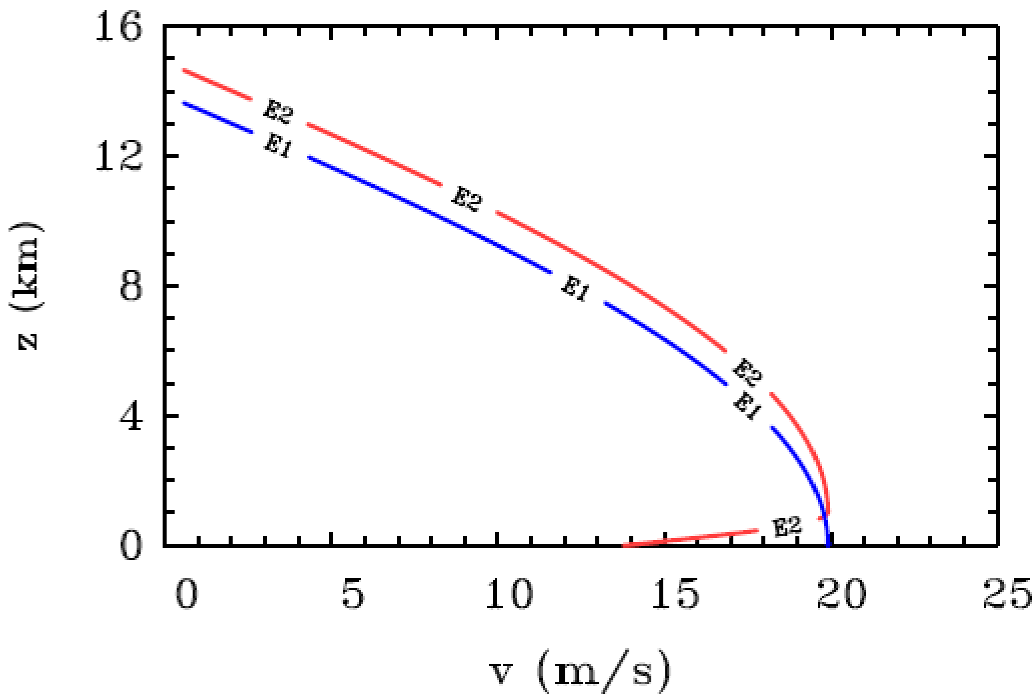


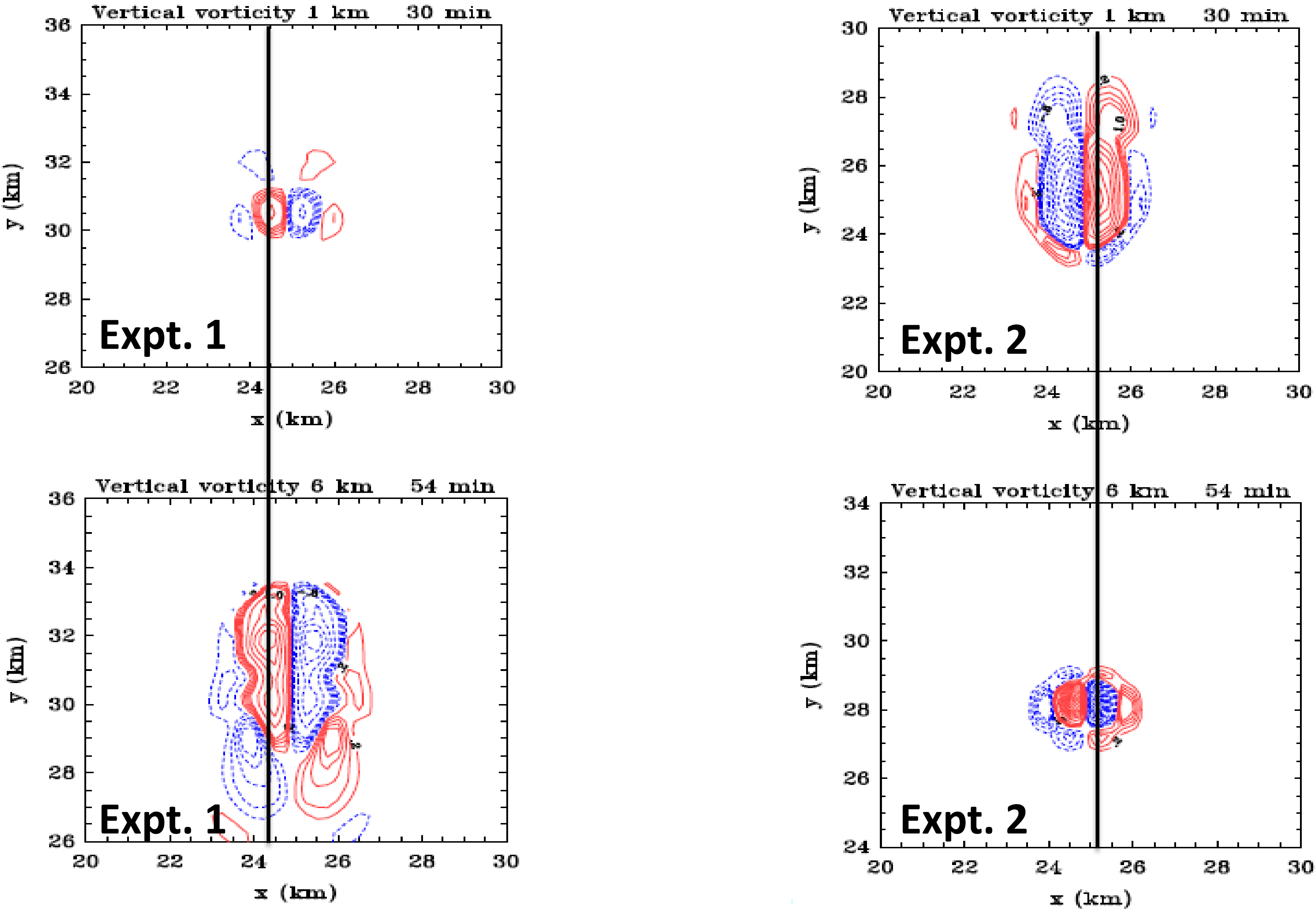
FIG. 2. Wind profiles as defined by Eq. (4).



No horiz shear
Vert shear **Expt 1: NoBL**
Vert shear **Expt 2: BL**
f : No background rotation

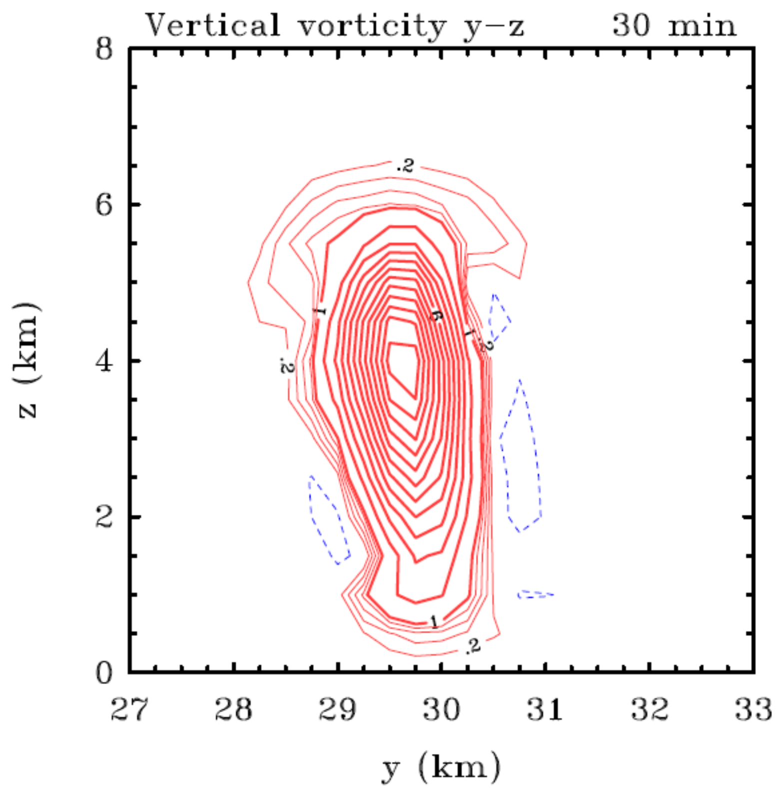
CAPE: 2080 $J\ kg^{-1}$
CIN: 40 $J\ kg^{-1}$
TPW: 59.1 $kg\ m^{-2}$

Thermal bubble: 2 K

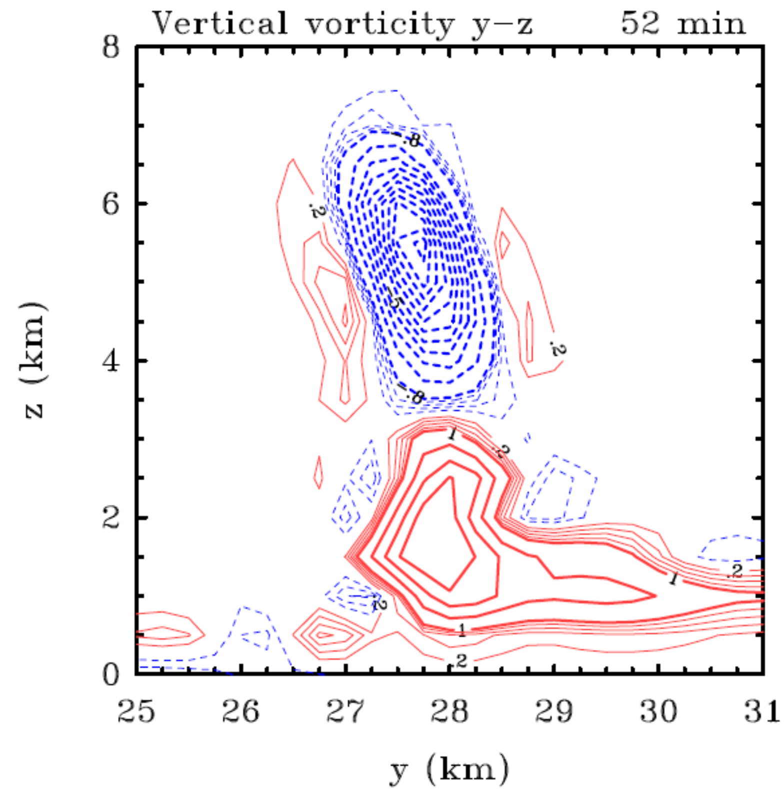


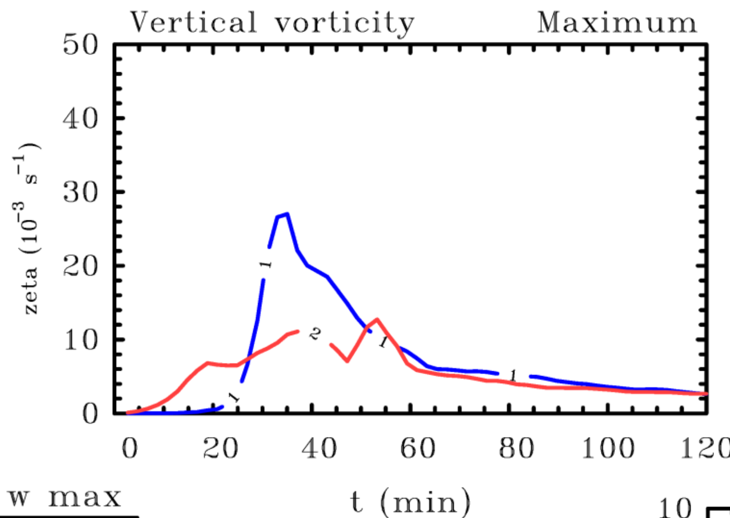
Uni-directional shear

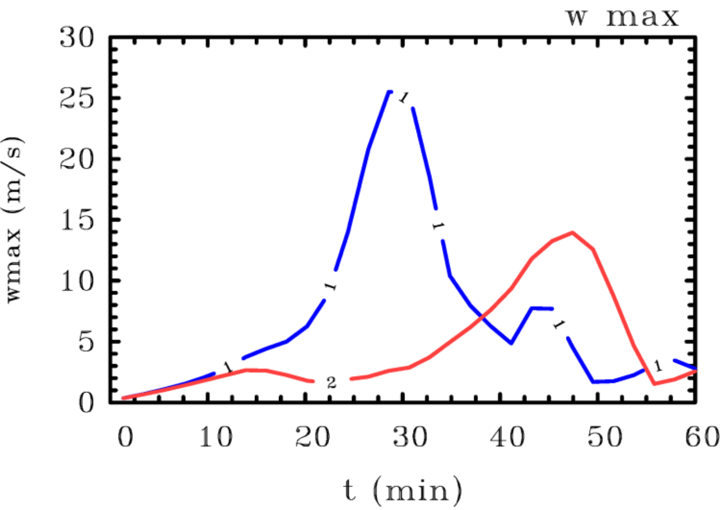
Expt. 1

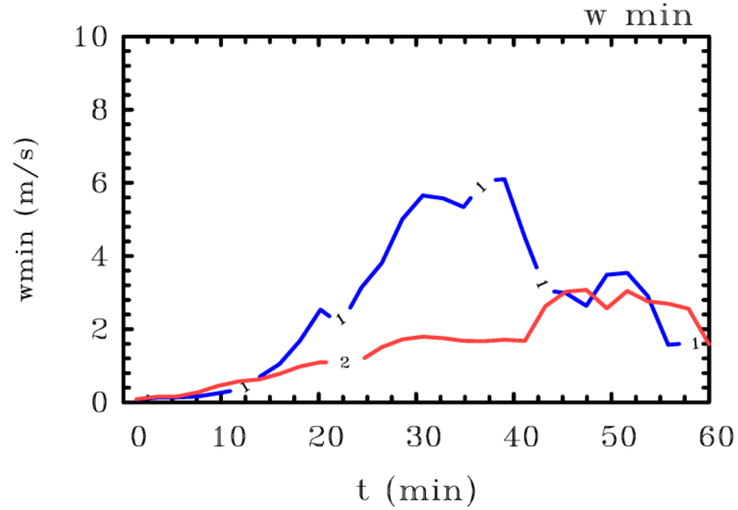

Vertical vorticity y–z 30 min
z (km)
y (km)
0
2
4
6
8
27
28
29
30
31
32
33
.2
1
.6

Expt. 2


Vertical vorticity y–z 52 min
z (km)
y (km)
0
2
4
6
8
25
26
27
28
29
30
31
.2
1
.2





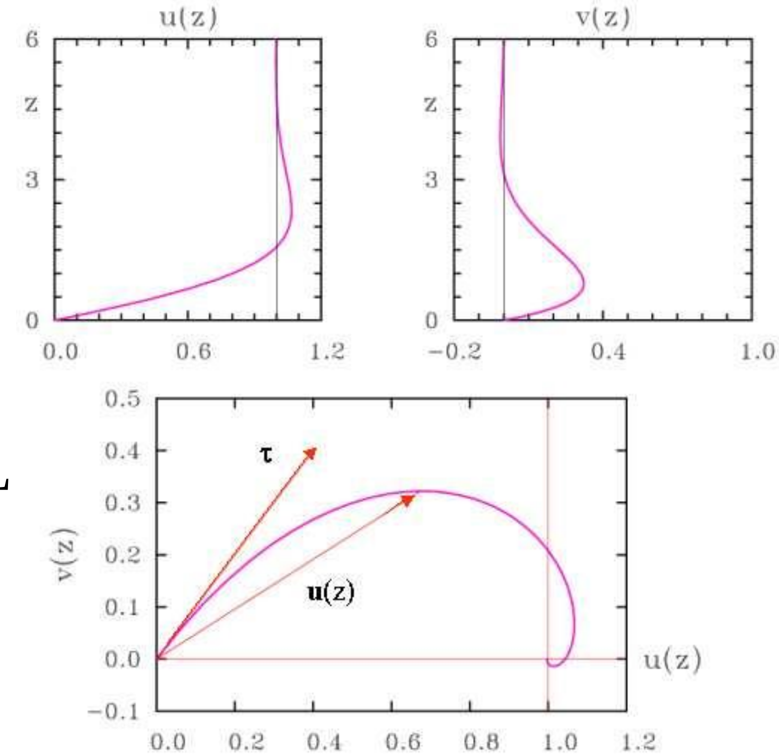


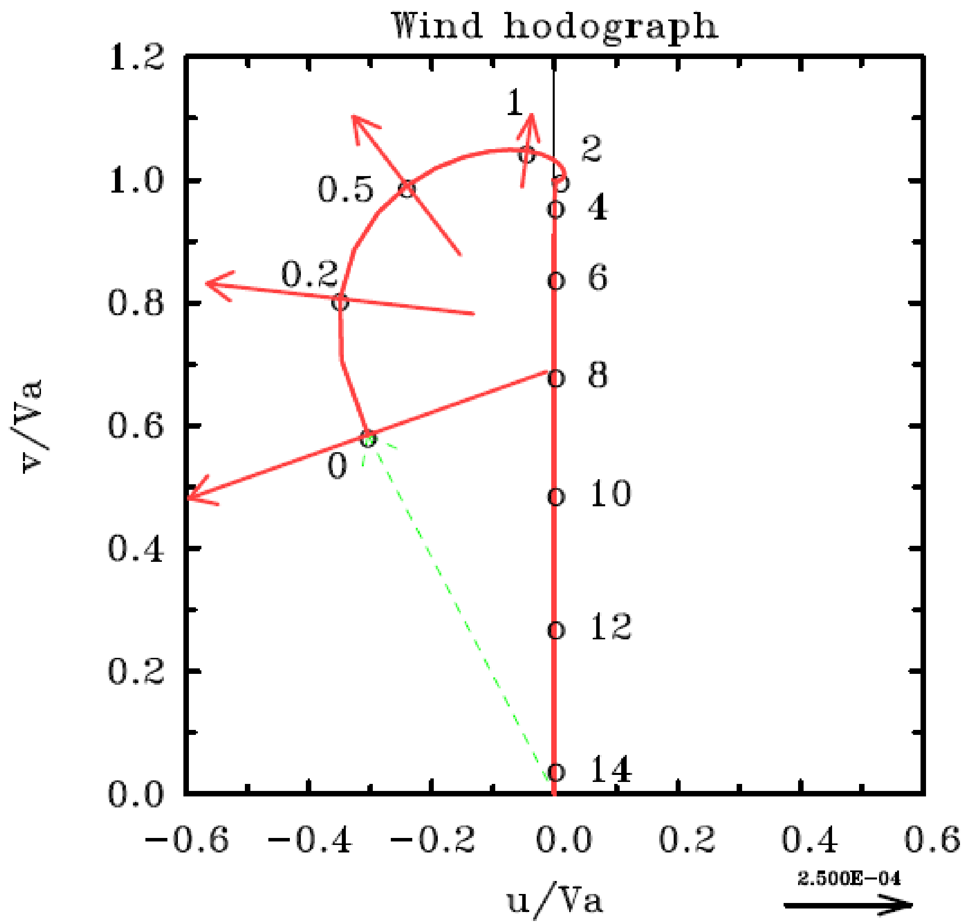
Uni-directional shear

Uni-directional shear: conclusions

- BL has a dramatic effect on convection, weakening convective updraughts and downdraughts
- Reduction in the vertical vorticity maximum with a BL profile
- Deformation of the initial thermal by low-level shear plays a large role
- Vertical vorticity dipole reverses sign with height when a BL is implemented
- Makes interpretations of vorticity anomaly merger more complicated

- Uni-directional wind shear within the BL is not very realistic
- Significant radial wind component in the BL, clockwise turning hodograph
- Previous numerical studies of experiments with clockwise turning hodographs did not include negative vertical shear above the BL
- **Key question: How does a cyclonically turning vortex-like BL affect vertical vorticity generation?**





No horiz shear
Vert shear < 2 km: Vortex BL
Vert shear > 2 km: Neg shear
$f : \zeta_0 = 3 \times 10^{-4}\ s^{-1}$

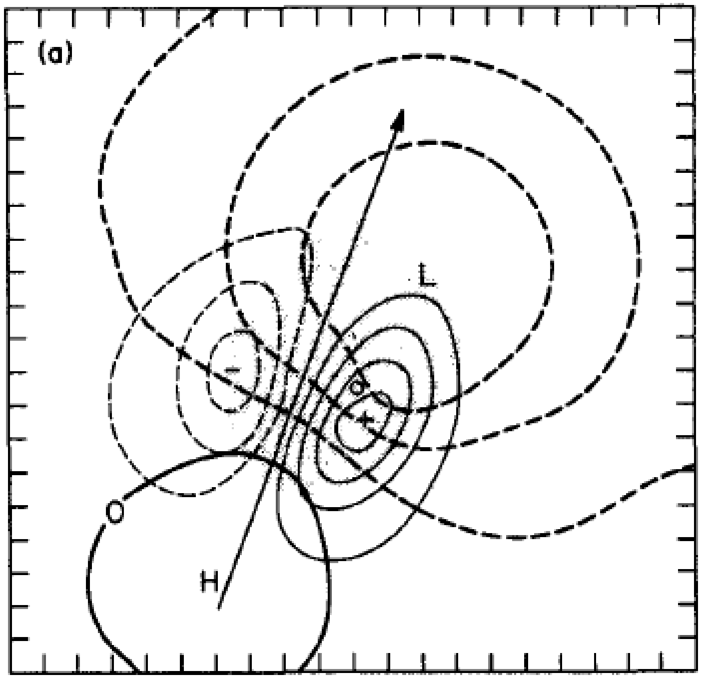
CAPE: 2080 $J\ kg^{-1}$
CIN: 40 $J\ kg^{-1}$
TPW: 59.1 $kg\ m^{-2}$

Thermal bubble: 3 K

What to expect?

- Clockwise turning hodograph favours cyclonic vorticity production (Klemp & Wilhelmson 1978, Rotunno 1982 Rotunno & Klemp 1982)
- Linear theory (Rotunno & Klemp 1982) – interaction of mean shear and updraught produces favourable vertical pressure perturbation gradients on one flank

Linear theory predicts that the vorticity dipole is aligned perpendicular to the shear vector, and a pressure perturbation dipole is aligned parallel to the shear vector

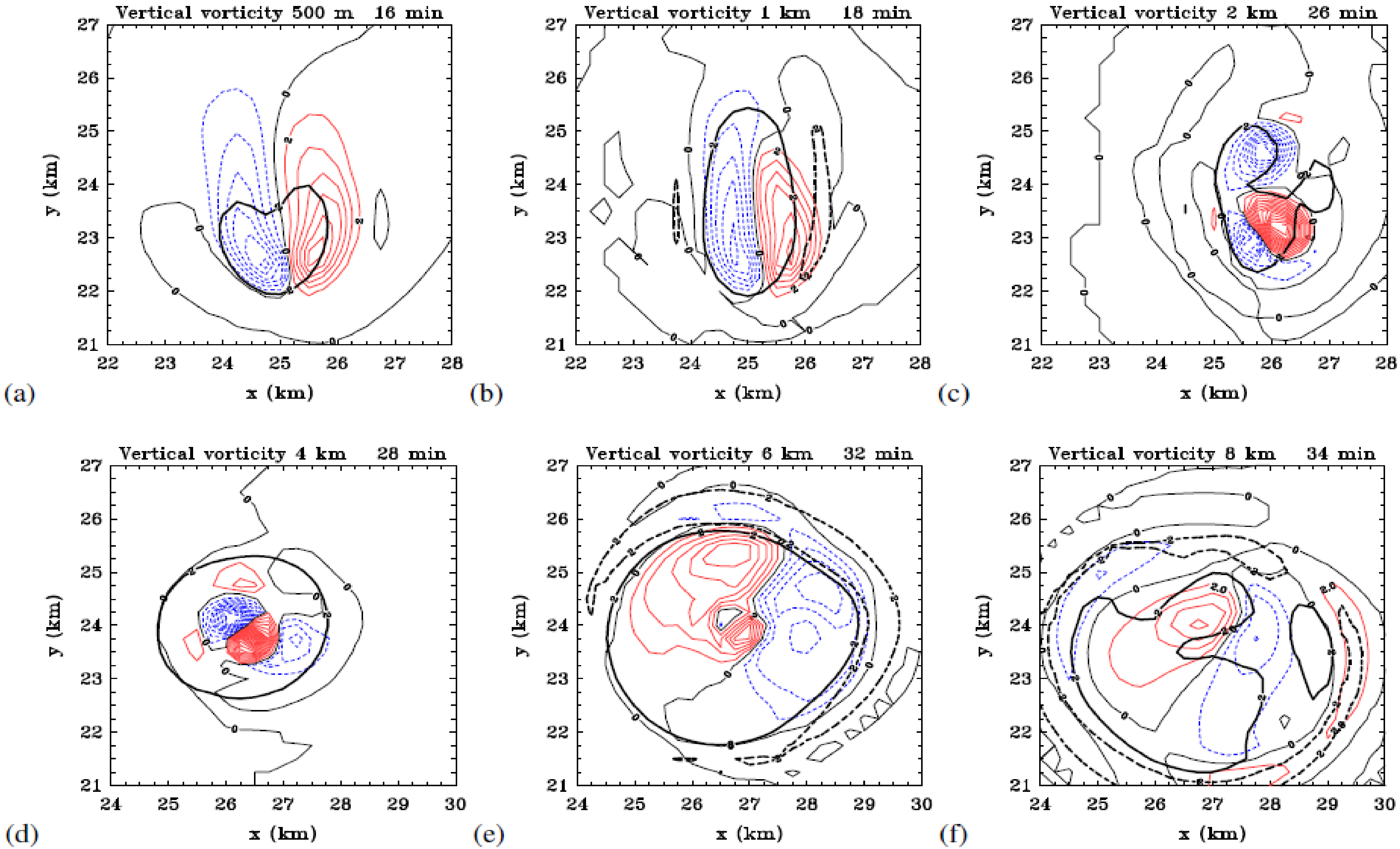


Rotunno & Klemp 1982

What to expect?

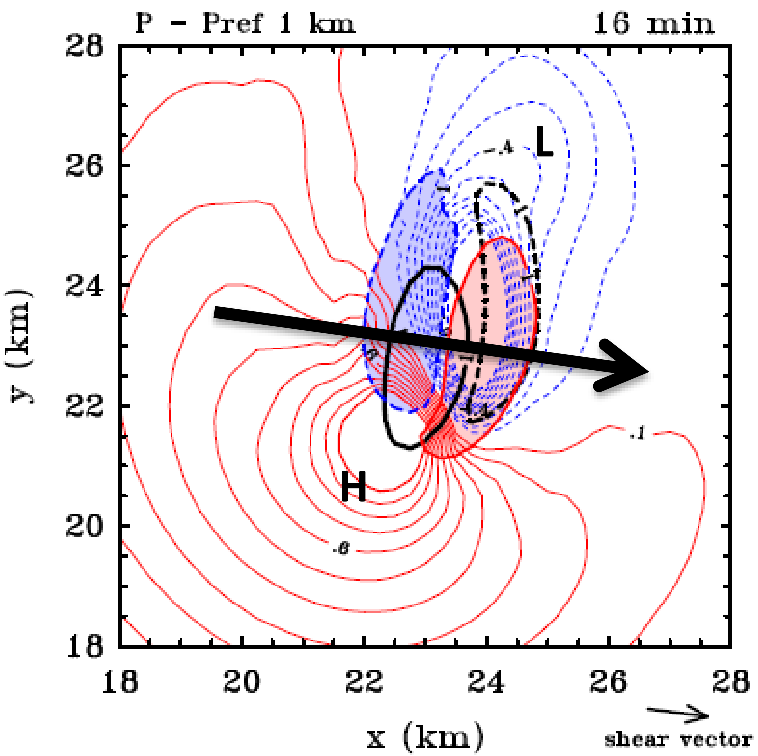
- Enhanced cyclonic vorticity production due to stretching of background vertical vorticity
- Vertical vorticity dipole that changes sign with height, as before

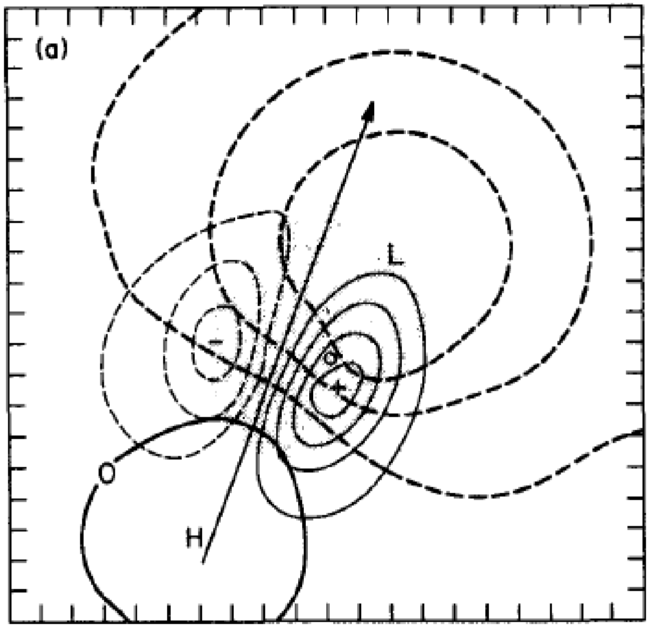
- Key question: How is vertical vorticity produced within the vortex BL and how does the result affect vortex merger?



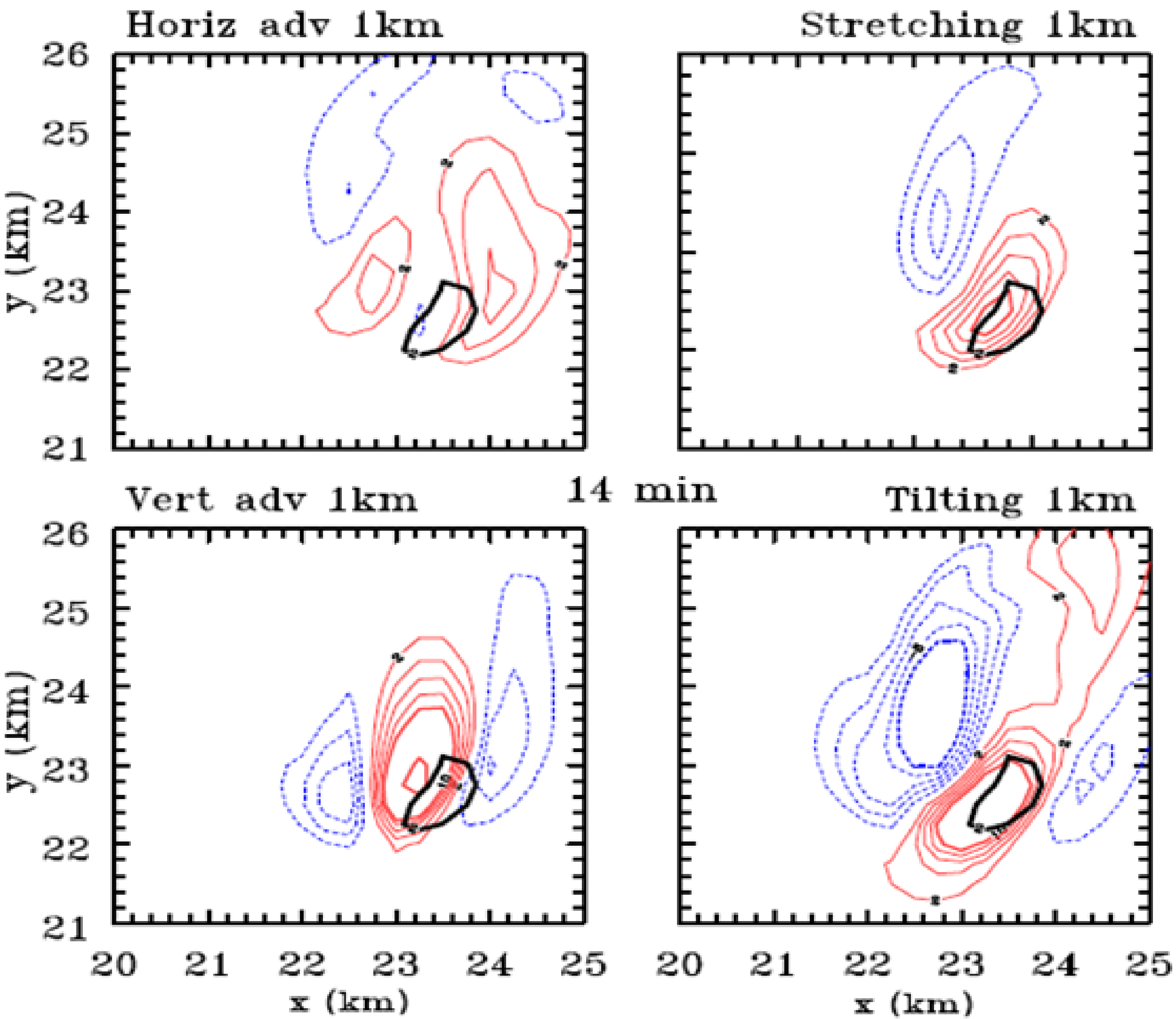
Directional shear

- Problem with linear theory? Vorticity dipoles are not aligned perpendicular to shear vector





$$\frac{\partial \zeta_a}{\partial t} = \overbrace{-\mathbf{v}_h \cdot \nabla \zeta_a - w\frac{\partial \zeta}{\partial z}}^{\text{advective}} \overbrace{- \zeta_a \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right)}^{\text{stretching}} + \overbrace{\left(\frac{\partial w}{\partial y}\frac{\partial u}{\partial z} - \frac{\partial w}{\partial x}\frac{\partial v}{\partial z}\right)}^{\text{tilting}}$$



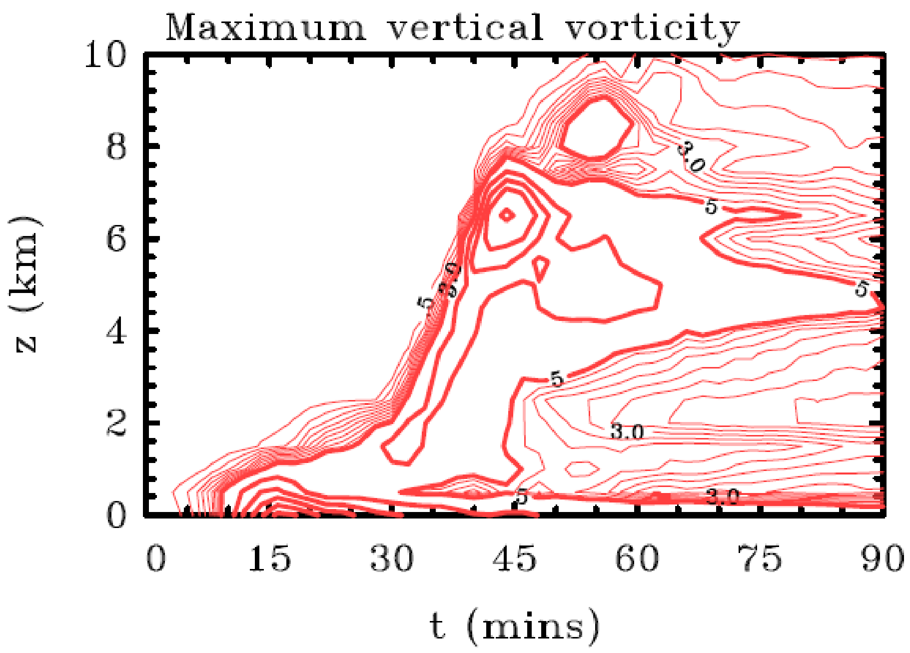
Directional shear

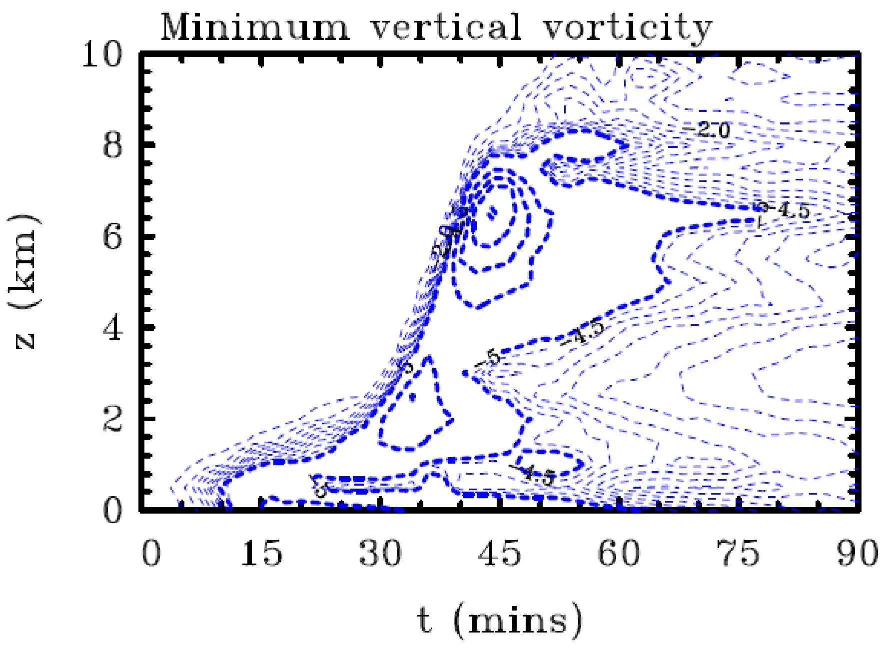
On the Evolution of Vorticity and Potential Vorticity in the Presence of Diabatic Heating and Frictional or Other Forces

P. H. HAYNES AND M. E. MCINTYRE

(I) *There can be no net transport of vorticity across any isobaric surface.*

(II) *Vorticity can neither be created nor destroyed, within a layer bounded by two isobaric surfaces.*


Maximum vertical vorticity
z (km)
t (mins)
0
15
30
45
60
75
90
0
2
4
6
8
10
5
3.0

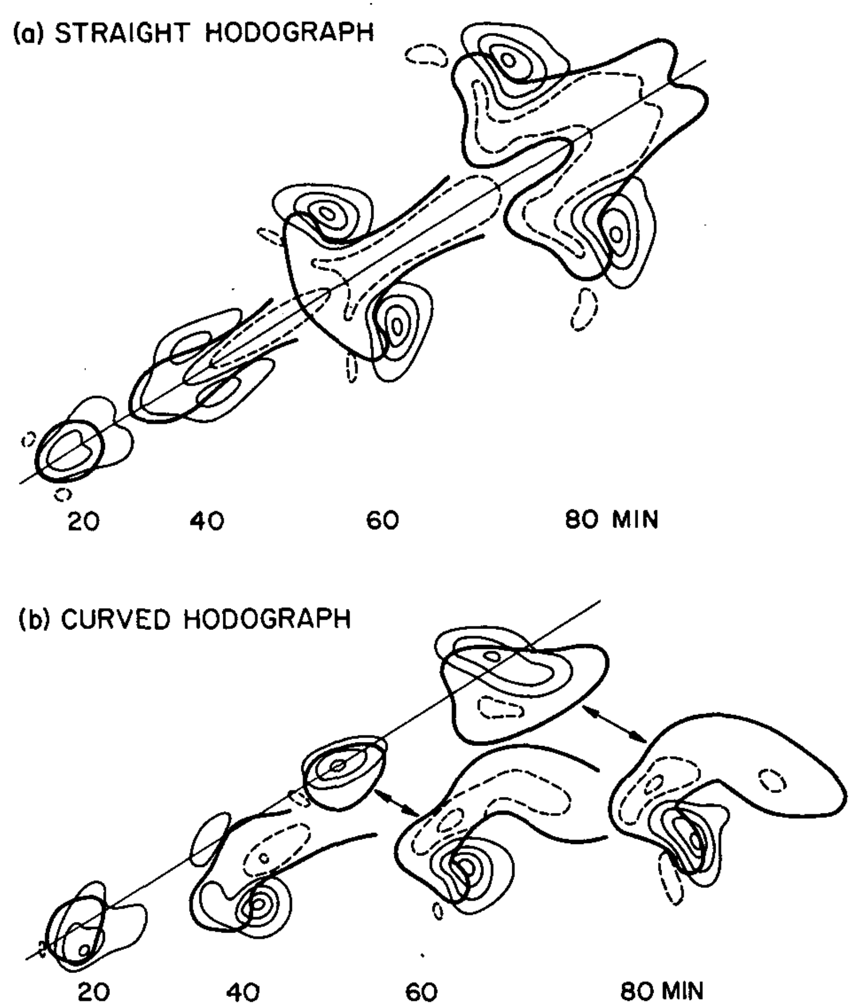

Minimum vertical vorticity
z (km)
t (mins)
0
15
30
45
60
75
90
0
2
4
6
8
10
-2.0
-4.5
-5

Directional shear: conclusions

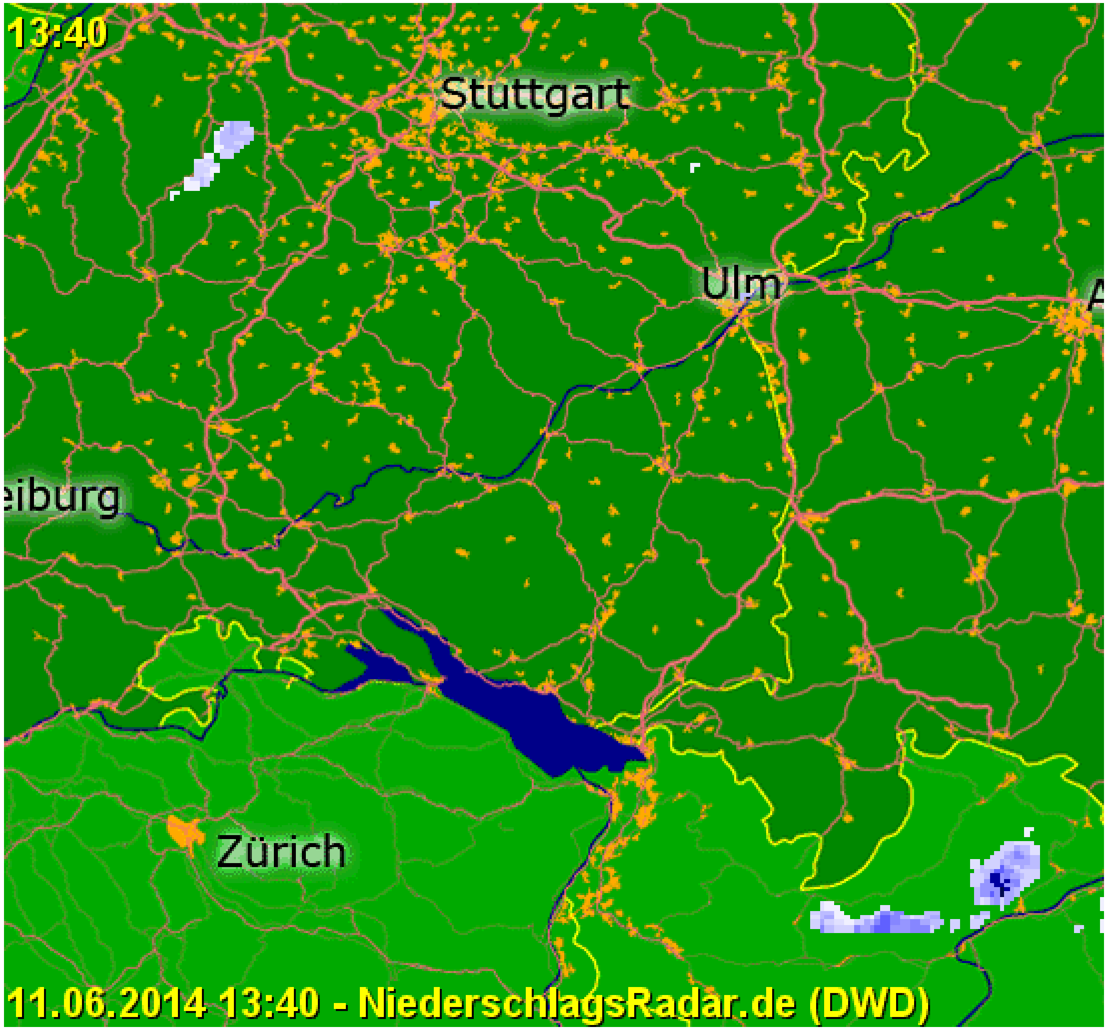
- Vortex BL has a dramatic effect on vorticity generation, leading to a vorticity dipole that rotates with height (following the shear vector)
- Linear theory is a poor representation, ignores vertical advection term which is dominant at early times
- Cyclonic vorticity production is favoured, especially at low levels and dominates over time
- Interpretations of vorticity anomaly merger?

Storm splitting

- Convective cells can split into two if there is sufficiently large enough shear - vertical or horizontal (Rozoff 2007)
- Cyclonic and anticyclonically rotating clouds
- Supercells are often the result of storm splitting

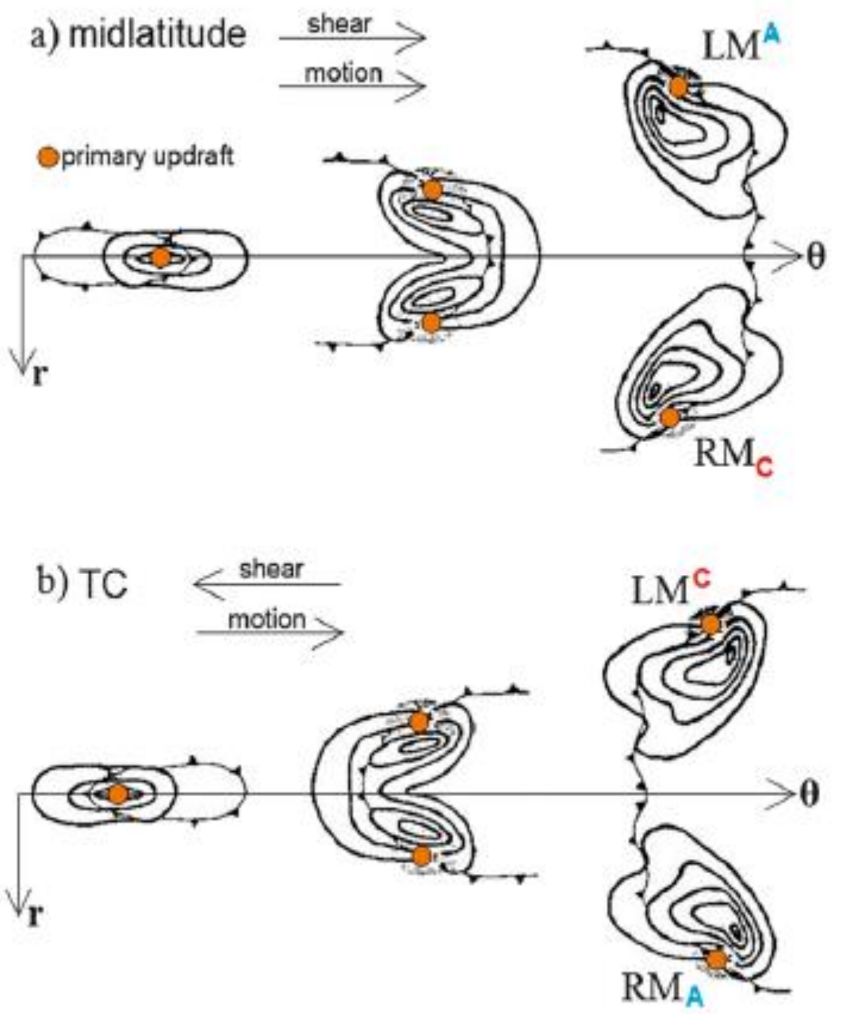


Rotunno and Klemp 1982


13:40
Stuttgart
Ulm
eiburg
Zürich
11.06.2014 13:40 - NiederschlagsRadar.de (DWD)

Hogsett and Stewart (2014) - conceptual model involves storm splitting to explain the inward contraction of eyewall convection in a TC

In mid-latitudes the cyclonic right mover of a split storm develops into the classic supercell, while in the tropics the left mover develops – the vertical shear is opposite in sign

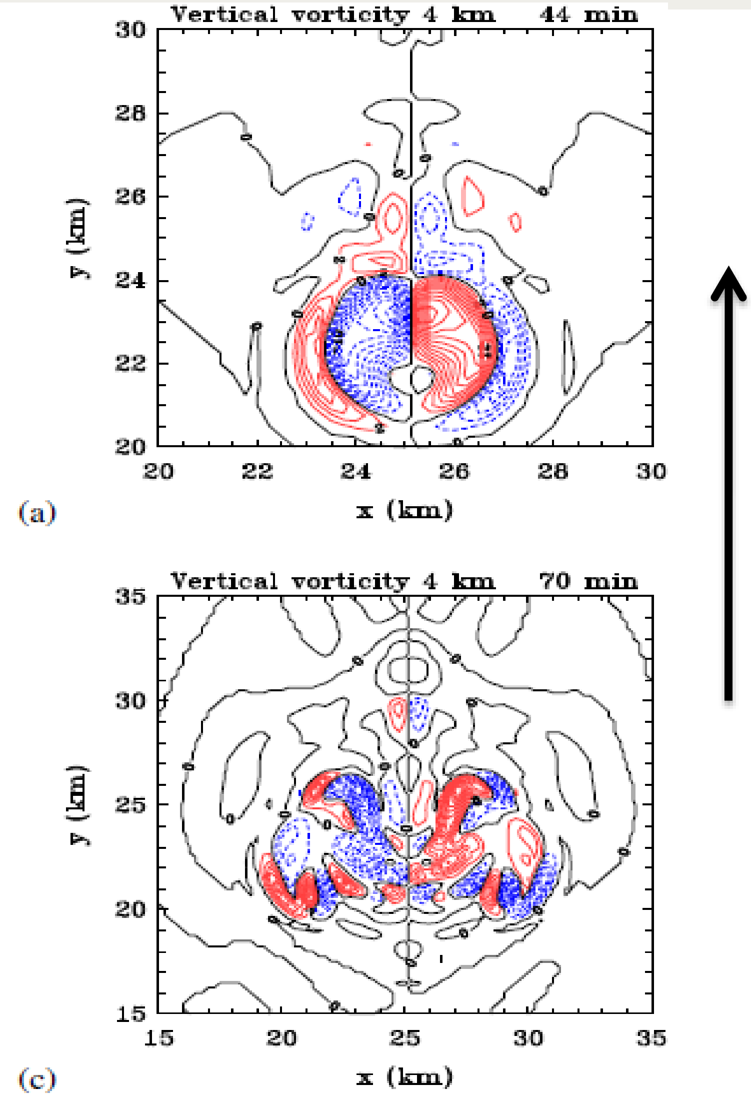


Splitting is messy!

Requirements:

- Large low level vertical shear
- High CAPE
- Low CIN
- Strong initial thermal

Here is a split with a unidirectional shear profile



- Inner core region is typically one of reduced CAPE, and increased CIN, at least in the later stages of storm development, can splitting occur?
- Boundary layer leads to a vorticity dipole reversal in sign with height, so that the left mover has an anticyclonic anomaly at low levels
- HS14 did not consider the boundary layer, but said it was a key issue they would study later
- How can a vorticity structure like this lead to an inward contraction of eyewall convection?

Development of convection in a vortex

- Go one step further in realism and investigate vertical vorticity generation in an initially balanced warm-cored vortex.
- In such a vortex there are spatial variations of temperature, vorticity and tangential wind, all of which could influence the development of vertical vorticity within deep clouds growing in such an environment.
- Vertical vorticity developing within the radius of maximum tangential winds (where the flow regime is vorticity dominated) is expected to be longer lived, whereas that developing outside the radius of maximum winds (where the flow is strain dominated) is expected to be weaker and more filamented (Rozoff *et al.* 2009).
- Clouds developing outside the radius of maximum winds are more susceptible also to the entrainment of ambient air (Rozoff *et al.* 2006), which will affect the maximum updraught

Six experiments – convection located at different radii

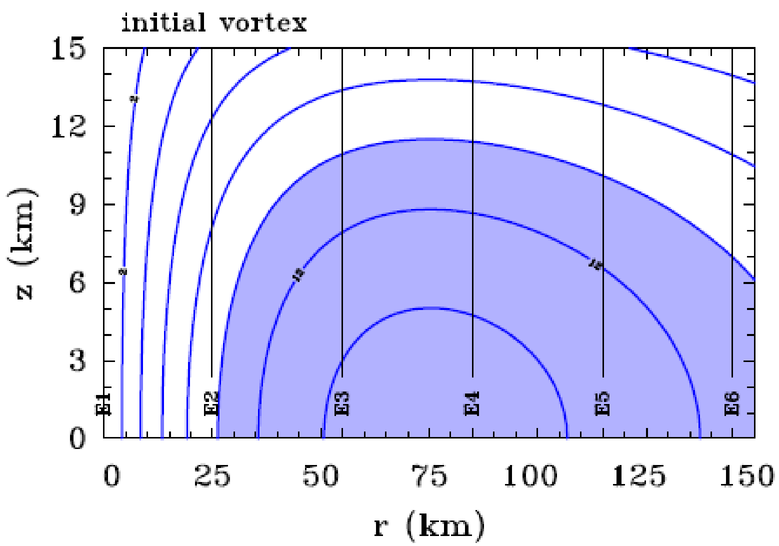


Figure 2. Vertical cross section of the initial vortex structure and the locations of the initial thermals. Contours: V wind component contoured every 2 m s^{-1}, with blue shading for regions stronger than 10 m s^{-1}. The locations of the initial thermal perturbation for the various experiments are indicated by vertical lines labelled E1, E2, etc.

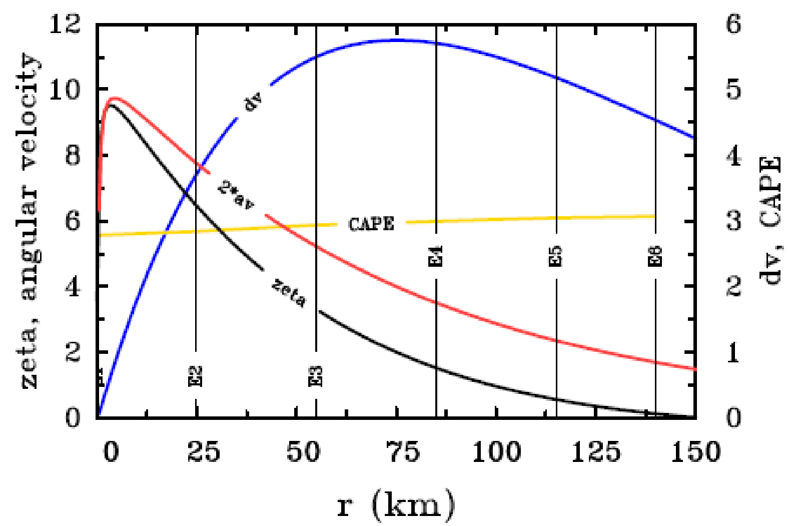
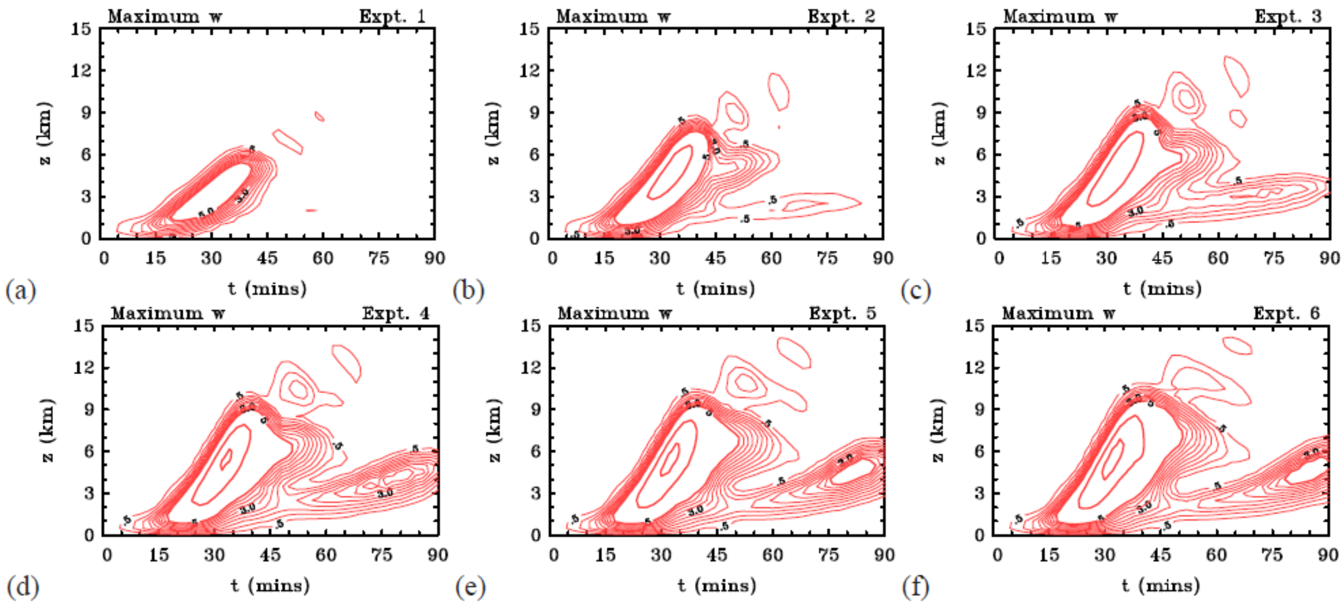
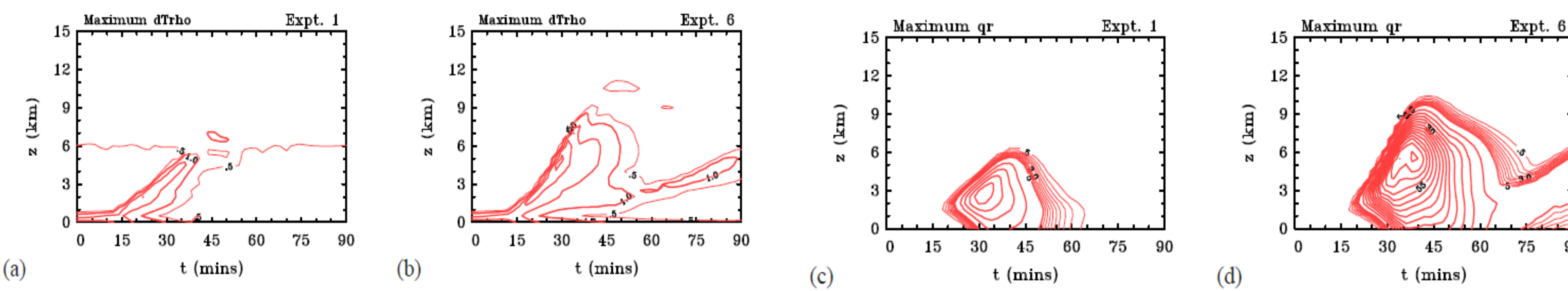
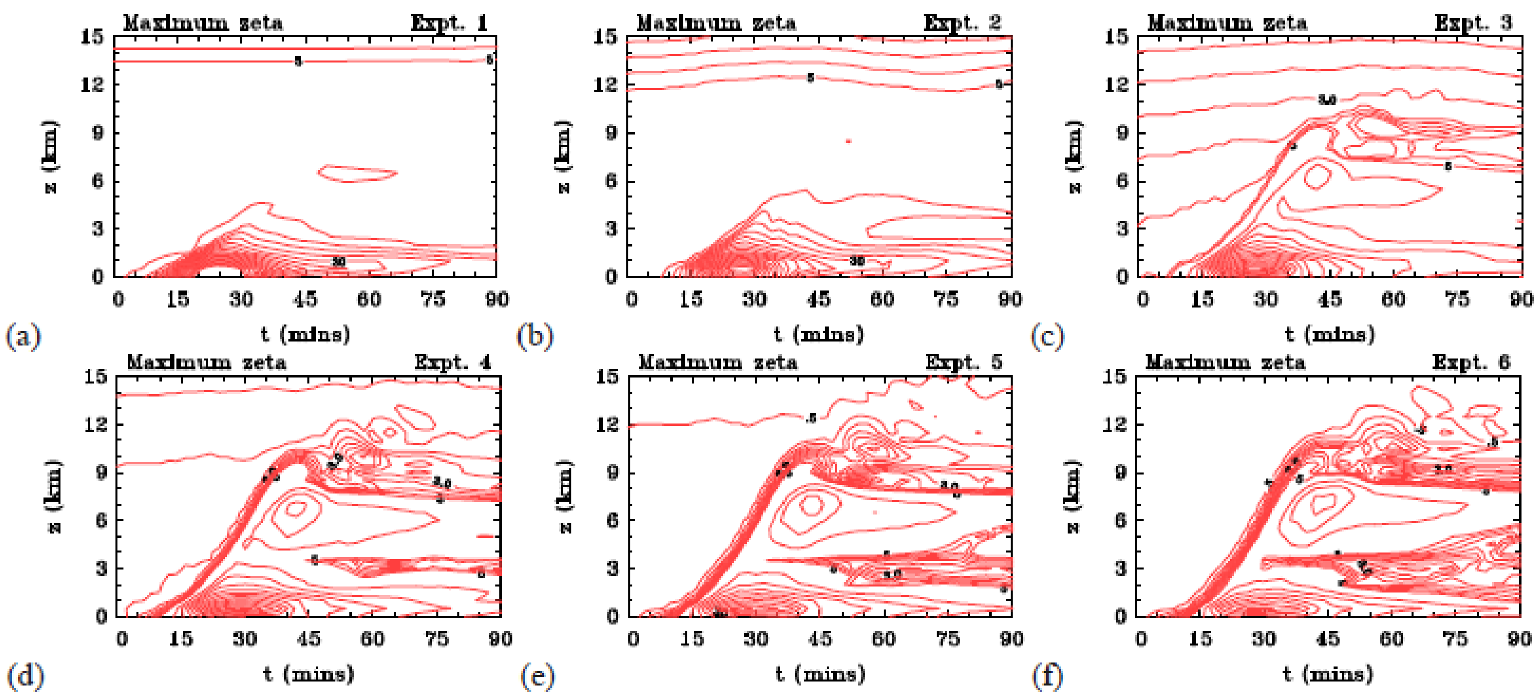
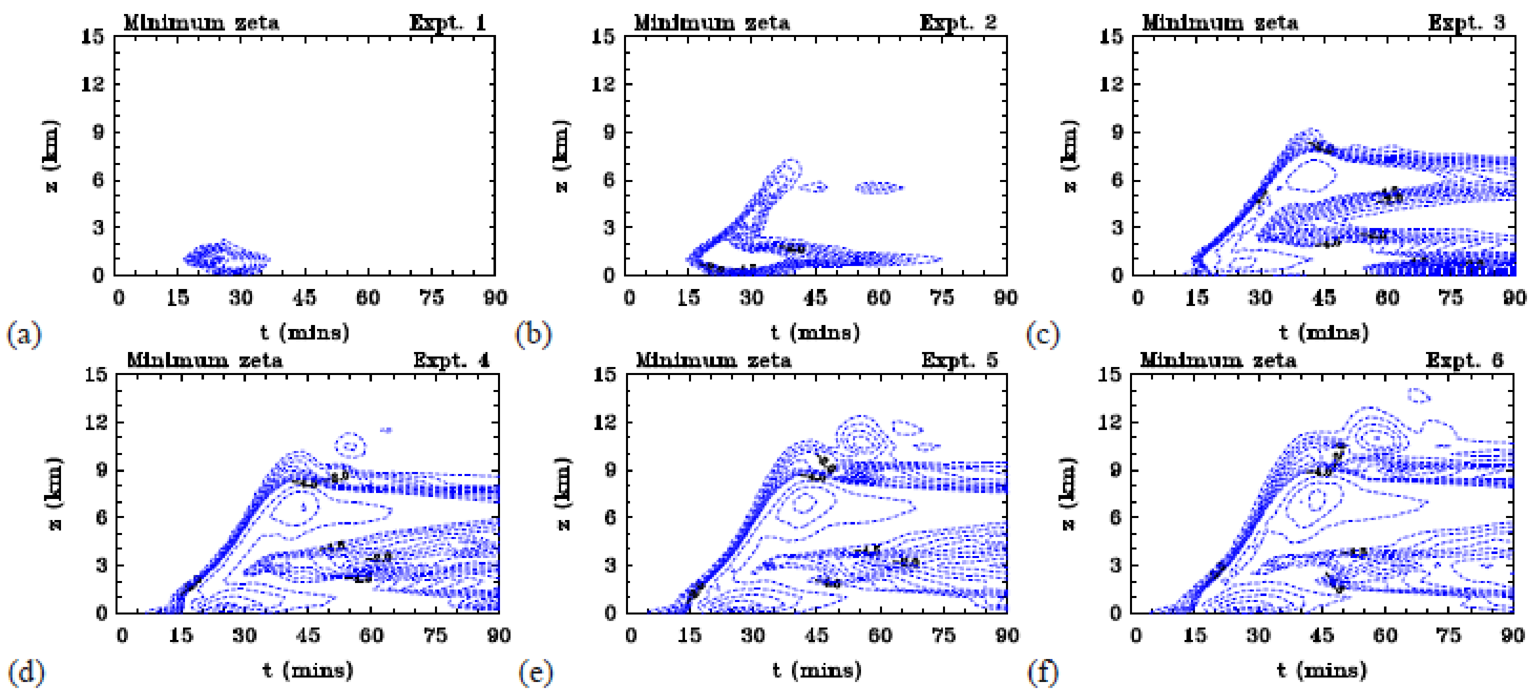


Figure 3. Radial variation of relative vorticity (zeta) and twice the relative angular velocity (2*av) at the surface, vertical shear (dv), and CAPE for the initial vortex shown in Figure 2. The locations of the initial thermal perturbation for the various experiments are indicated by vertical lines labelled E1, E2, etc.


Maximum w
Expt. 1
Expt. 2
Expt. 3
Expt. 4
Expt. 5
Expt. 6
z (km)
t (mins)
(a)
(b)
(c)
(d)
(e)
(f)


Maximum dTrho
Expt. 1
Expt. 6
Maximum qr
Expt. 1
Expt. 6
z (km)
t (mins)
(a)
(b)
(c)
(d)


Maximum zeta
Expt. 1
Expt. 2
Expt. 3
Expt. 4
Expt. 5
Expt. 6
z (km)
t (mins)
0 15 30 45 60 75 90
0 3 6 9 12 15
(a)
(b)
(c)
(d)
(e)
(f)


Minimum zeta
Expt. 1
Expt. 2
Expt. 3
Expt. 4
Expt. 5
Expt. 6
z (km)
t (mins)
(a)
(b)
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(d)
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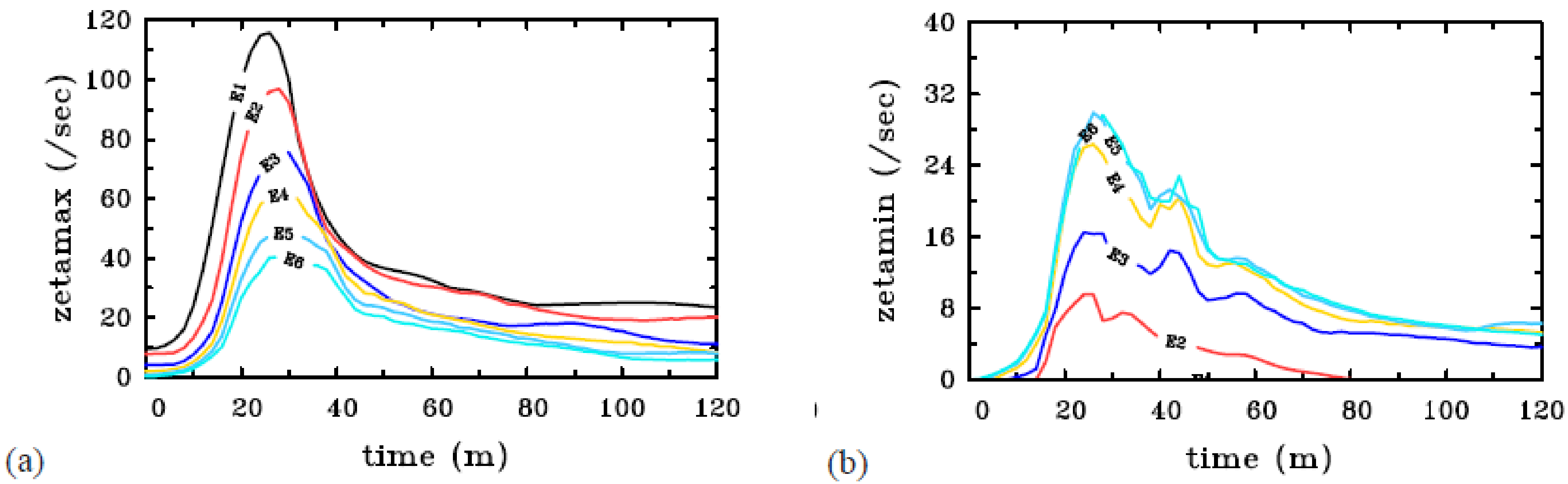
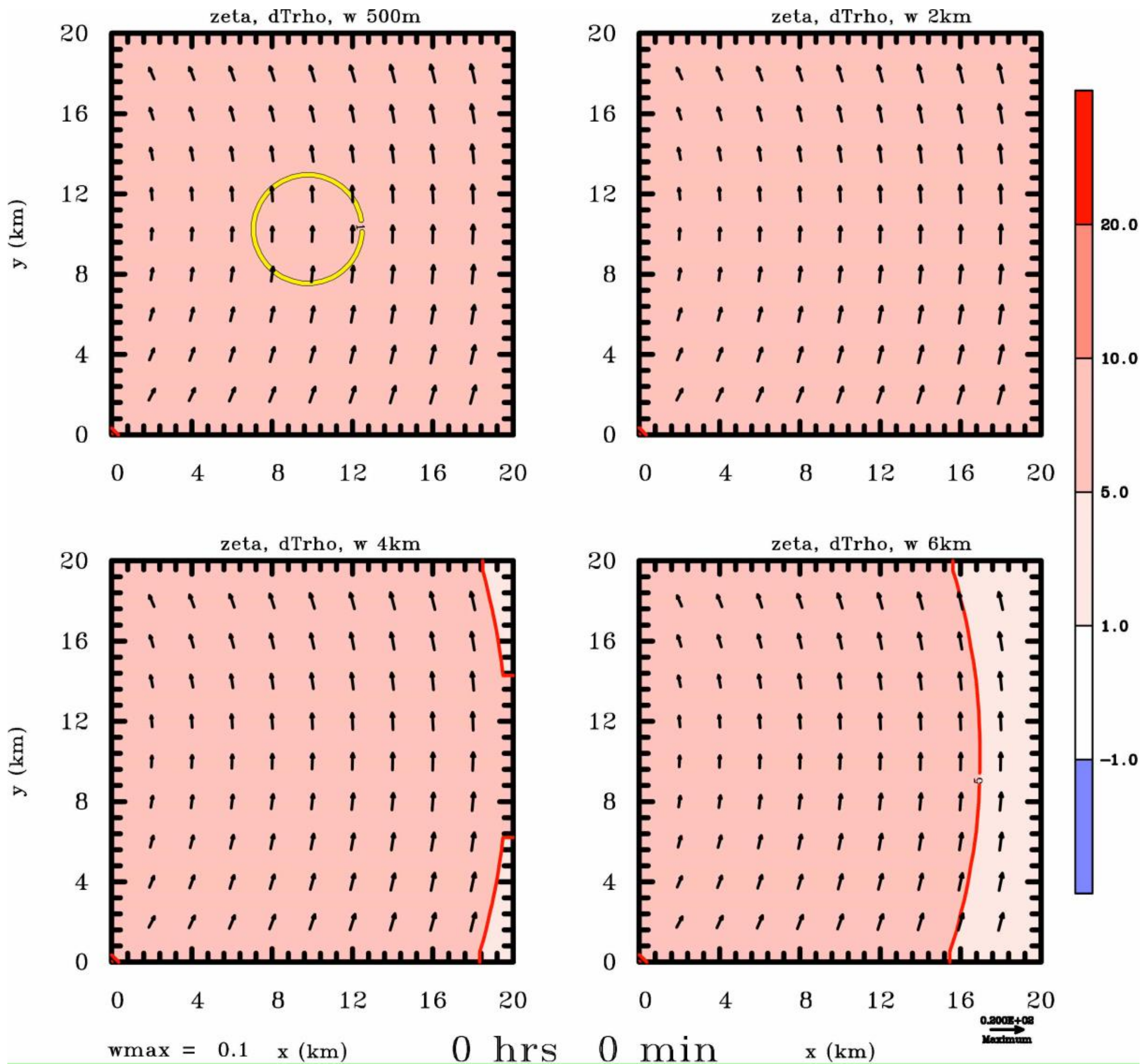
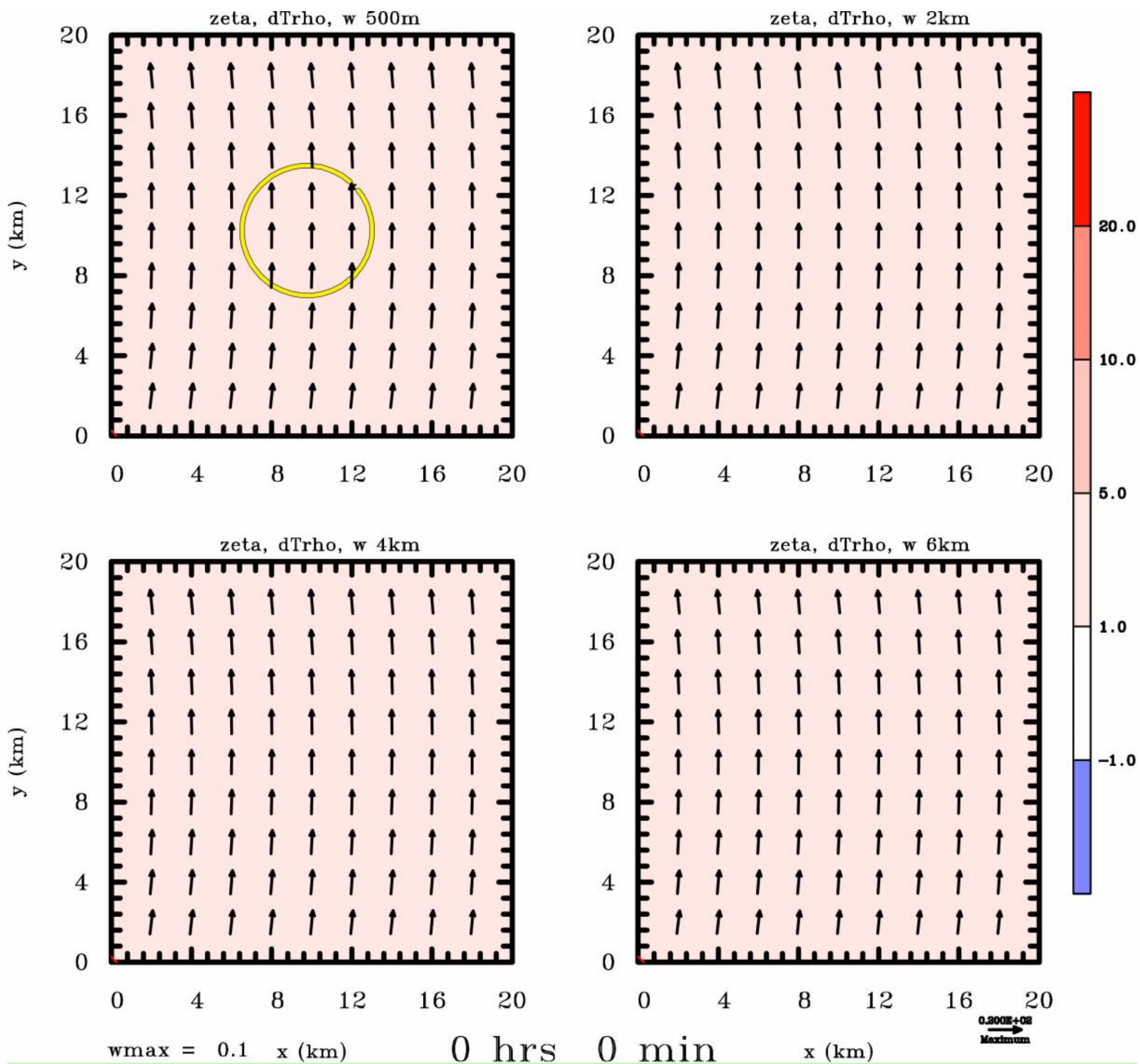
(f)


zetamax (/sec)
time (m)
E1
E2
E3
E4
E5
E6
(a)
zetamin (/sec)
time (m)
E6
E5
E4
E3
E2
(b)


zeta, dTrho, w 500m
zeta, dTrho, w 2km
zeta, dTrho, w 4km
zeta, dTrho, w 6km
y (km)
x (km)
20.0
10.0
5.0
1.0
−1.0
wmax = 0.1
0 hrs 0 min


zeta, dTrho, w 500m
zeta, dTrho, w 2km
zeta, dTrho, w 4km
zeta, dTrho, w 6km
y (km)
x (km)
0 4 8 12 16 20
20.0
10.0
5.0
1.0
−1.0
wmax = 0.1
0 hrs 0 min
0.200E+02
Maximum

As the location of the thermal perturbation is moved away from the axis of rotation:

(1) the updraught that develops becomes stronger
(2) the cyclonic vorticity anomaly generated by the updraught becomes weaker
(3) the structure of the vorticity anomaly changes
(4) the depth of the vorticity anomaly increases.

For an updraught along or near the vortex axis, the vorticity anomaly has the structure of a monopole and little or no anticyclonic vorticity is generated in the core.

This finding obviates the need in previous simulations of tropical cyclone intensification to explain how anticyclonic vorticity would be expelled from the core of an intensifying vortex when there are multiple clouds contributing to drive the intensification.

Vorticity dipoles are generated in updraughts near or beyond the radius of maximum tangential wind speed and this structure reverses in sign with height.

In all cases these anomalies persists long after the initial updraught has decayed.

Basis for unified view of tropical cyclogenesis and intensification:

- Deep convection developing in the presence of vertical vorticity amplifies the vorticity locally by vortex tube stretching, irrespective of the strength of the updraught and the depth of convection

- The vortical remnants outlive the convection that produced them in the first place.

- The vortical remnants tend to aggregate in a quasi two dimensional manner with a corresponding upscale energy cascade and some of these remnants will be intensified further by subsequent convective episodes.

Continued…

- The amplification and aggregation of vorticity represents an increase in the circulation within a fixed circuit encompassing the convective area – provided vorticity enters the circuit.
- As the circulation progressively increases in strength, there is some increase in the surface moisture fluxes.
- It is not necessary that the moisture fluxes continue to increase with surface wind speed. (WISHE discussion)

The Finite-Amplitude Nature of Tropical Cyclogenesis

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(Manuscript received 2 February 1989, in final form 30 May 1989)

ABSTRACT

We have constructed a simple, balanced, axisymmetric model as a means of understanding the existence of the threshold amplitude for tropical cyclogenesis discovered by Rotunno and Emanuel. The model is similar to Ooyama's but is phrased in Schubert and Hack's potential radius coordinates.

The essential difference between this and other balanced models lies in the representation of convective clouds. In the present model the cumulus updraft mass flux depends simply and directly on the buoyancy (on angular momentum surfaces) of lifted subcloud-layer air and is not explicitly constrained by moisture convergence. The downdraft mass flux is equal to the updraft flux multiplied by $(1 - \epsilon)$, where ϵ is the precipitation efficiency. The complete spectrum of convective clouds in nature is here represented by two extremes: deep clouds with a precipitation efficiency of one, and shallow, nonprecipitating clouds. The former stabilize the atmosphere both by heating the free atmosphere and drying out the subcloud layer, whereas the shallow clouds stabilize only through drying of the subcloud layer. The two cloud types may coexist. In the crude vertical structure of

Some Aspects of Hurricane Inner-Core Dynamics and Energetics

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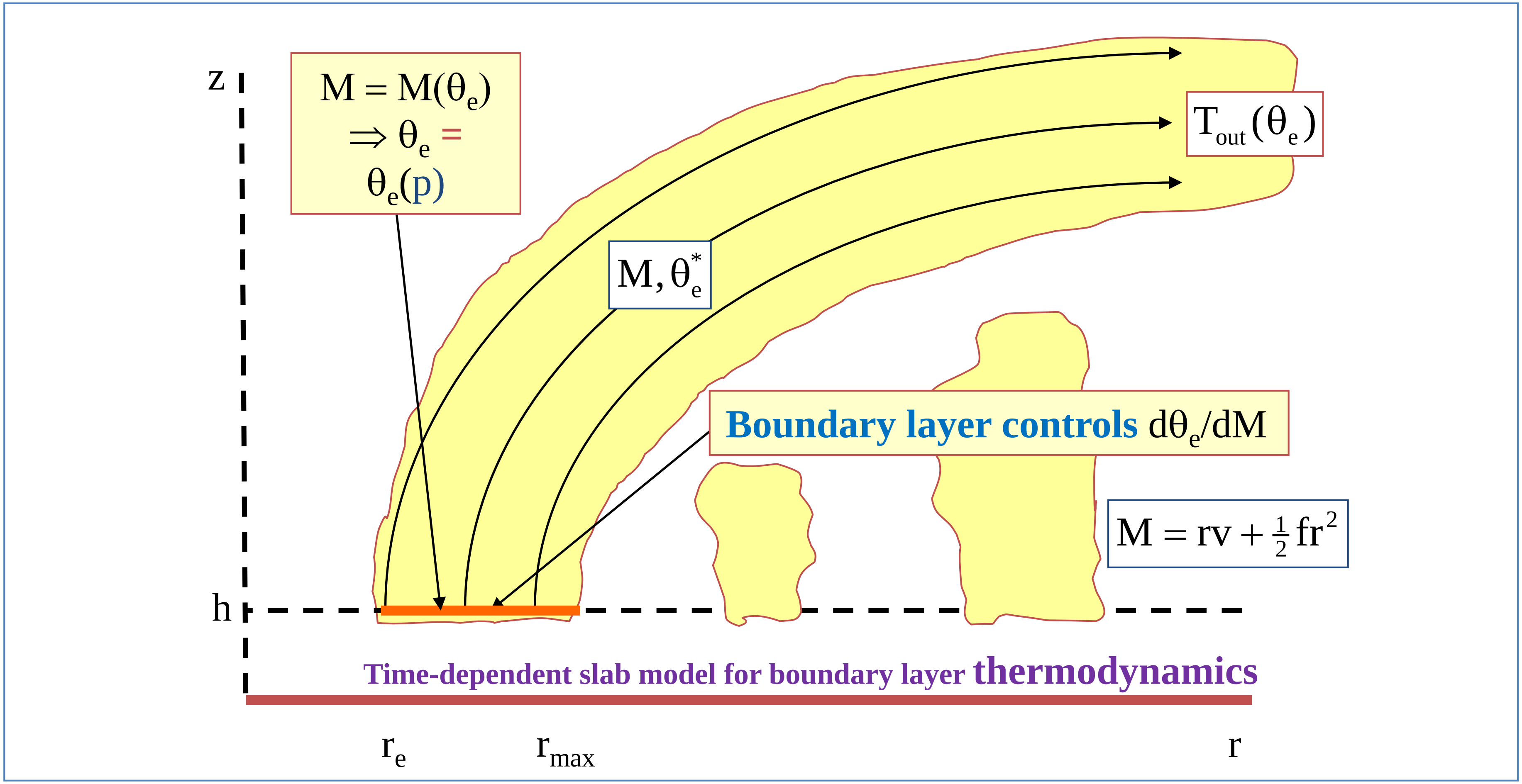
(Manuscript received 26 December 1995, in final form 18 October 1996)

WISHE

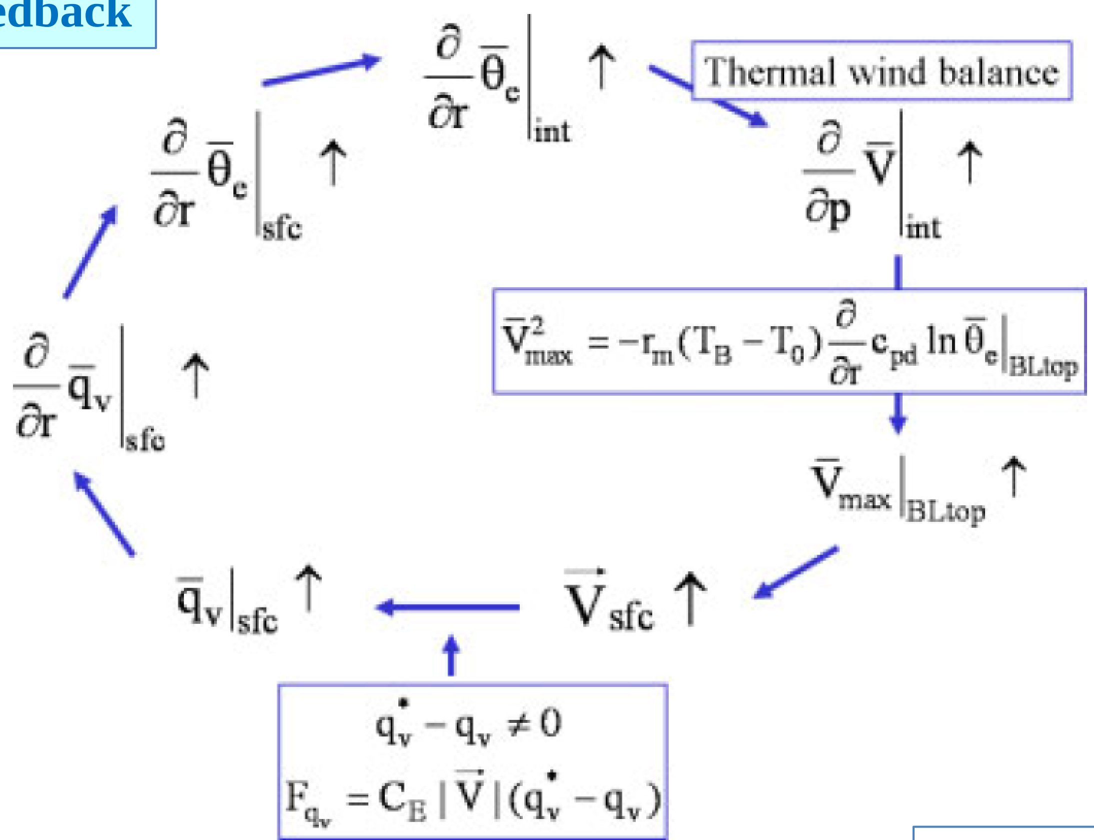
- **The air-sea interaction instability model for vortex intensification comprises a postulated multi-step feedback loop involving, in part, the near-surface wind speed and the evaporation of water from the underlying ocean, with the evaporation rate being a function of wind speed and thermodynamic disequilibrium.**
- **The evaporative-wind feedback mechanism is now commonly known as the Wind-Induced Surface Heat Exchange (WISHE) mechanism.**
- **Until very recently, the WISHE mechanism has been presented as a finite-amplitude instability that requires a *finite-amplitude precursor disturbance* generated by some independent means (such as an easterly wave) to "kick start the heat engine" (cf. Hakim 2011, Holton and Hakim 2012).**

Emanuel's 1997 model for hurricane intensification



TROPICAL CYCLONES AND WISHE

The WISHE feedback



From Montgomery et al. QJ (2009)

WISHE

- WISHE has been presented as a finite amplitude instability of an incipient tropical depression vortex and has achieved widespread acceptance in meteorology textbooks and other didactic material (e.g., Rauber et al. 2008, Holton 2004, Ahrens 2008, Holton and Hakim 2012, The COMET Program course in Tropical Meteorology), tropical weather briefings and the current literature (Lighthill 1998, Smith 2003, Molinari et al. 2004, Nong and Emanuel 2004, Montgomery et al. 2006, Terwey and Montgomery 2008, Braun et al. 2010, Fang and Zhang 2010).
- The last five citations and others have talked about "**igniting the WISHE mechanism**" after the vortex (or secondary maximum in the tangential wind) has reached some threshold intensity.
- Kepert (2011, p13) presents WISHE in the **context of a steady-state vortex** as "The role of the surface enthalpy fluxes in making the expansion of the inflowing boundary layer air isothermal rather than adiabatic ... ." He **does not mention the wind-speed dependence of the fluxes** and does not make a distinction between dry and moist enthalpy.

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(www.interscience.wiley.com) DOI: 10.1002/qj.459

RMetS
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Do tropical cyclones intensify by WISHE?

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RESEARCH ARTICLE
10.1002/2014MS000362

Putting to rest WISHE-ful misconceptions for tropical cyclone intensification

2015

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Key Points:
- WISHE is not the essential intensification mechanism for TCs
- WISHE is inconsistently represented

http://www.meteo.physik.uni-muenchen.de/~roger/Publications/M19

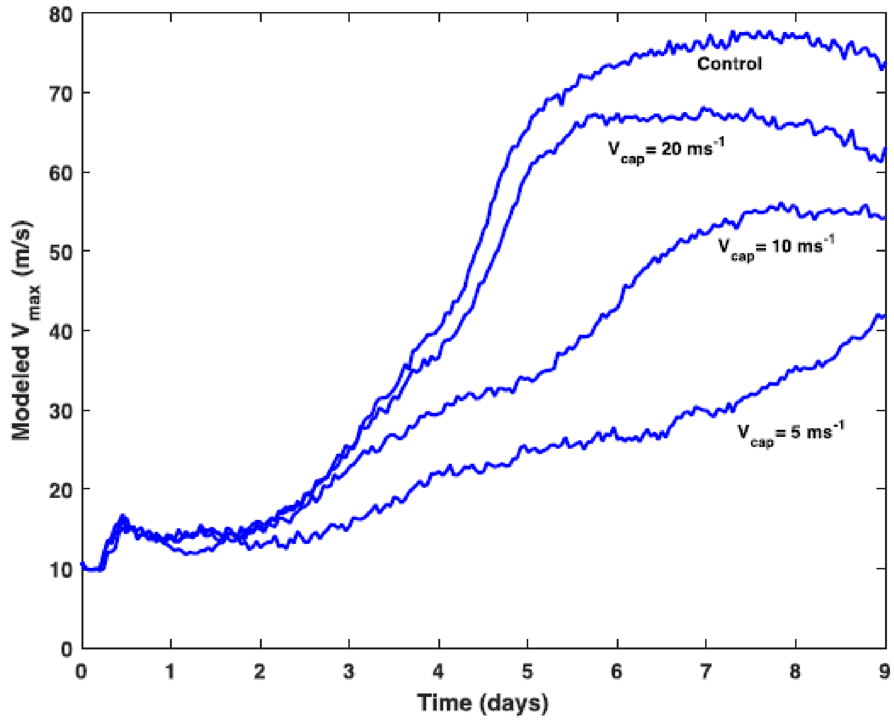

Figure 4: Evolution over time of the ensemble mean of the four experiments for which $C_k = 1 \times 10^{-3}$. The capping wind speeds label the curves.

- Intensification from a finite-amplitude initial vortex is shown to not require this evaporation–wind feedback process.

- Indeed, when the surface wind speed in the sea-to-air vapour fluxes is capped at a nominal (trade-wind) value, the vortex still intensifies by the same pathway identified in the main experiments via the generation of locally buoyant VHTs and the near-surface convergence that the VHTs induce within the boundary layer.

Zhang and Emanuel (2016)

On the Role of Surface Fluxes and WISHE in Tropical Cyclone Intensification

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"We show that the feedback between surface wind and surface enthalpy flux **is an important influence** on tropical cyclone evolution, even though, as with any classical instability mechanism, such a feedback is **not strictly necessary**."

END

A numerical study of rotating convection during tropical cyclogenesis

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Tropical convection: the effects of ambient vertical and horizontal vorticity

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Tropical-cyclone convection: the effects of a vortex boundary layer wind profile on deep convection

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A numerical study of deep convection in tropical cyclones

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