Interaction of Tropical Cyclones with the Upper Ocean

- **Resonance with near-inertial oscillations**
- **Mixed layer cooling by entrainment**
- **Coupled models**

Change on SST needed to cancel increase in enthalpy in core:

$$L_v q^*(T_a)H + c_p T_a = L_v q^*(T_a - \Delta T) + c_p (T_a - \Delta T)$$

$$L_v q^*(T_a - \Delta T) \cong L_v q^*(T_a) - L_v \frac{\partial q^*}{\partial T} \Delta T$$

$$= L_v q^*(T_a) - L_v \frac{L_v q^*}{R_v T_a^2} \Delta T$$

$$\rightarrow \Delta T \cong \frac{L_v q^* \left(1 - H\right)}{c_p + \dfrac{L_v{}^2 q^*}{R_v T_a^2}} \cong 2.5^o C$$

Physics of near-inertial oscillations:

PEs linearized about a rotating stratified fluid at rest:

$$\frac{\partial u}{\partial t} = -\alpha_0 \frac{\partial p}{\partial x} + fv$$

$$\frac{\partial v}{\partial t} = -\alpha_0 \frac{\partial p}{\partial y} - fu$$

$$\frac{\partial w}{\partial t} = -\alpha_0 \frac{\partial p}{\partial z} + B$$

$$\frac{\partial B}{\partial t} = -N^2 w$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0$$

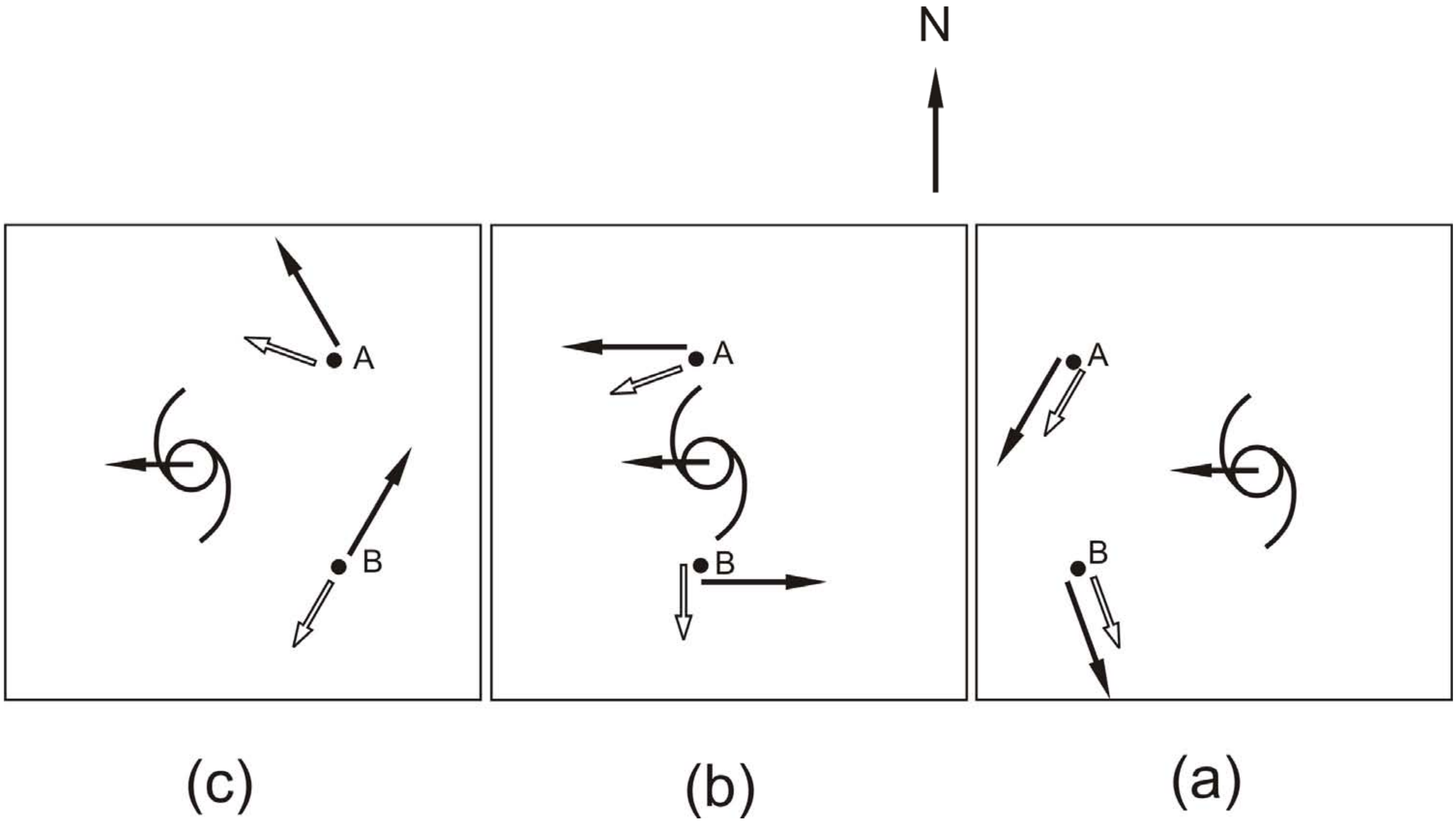
$$\rightarrow \frac{\partial^2}{\partial t^2} \nabla_3^2 w + N^2 \nabla_2^2 w + f^2 \frac{\partial^2 w}{\partial z^2} = 0$$

$$w = w_0 e^{i(kx+ly+rz-\omega t)}$$

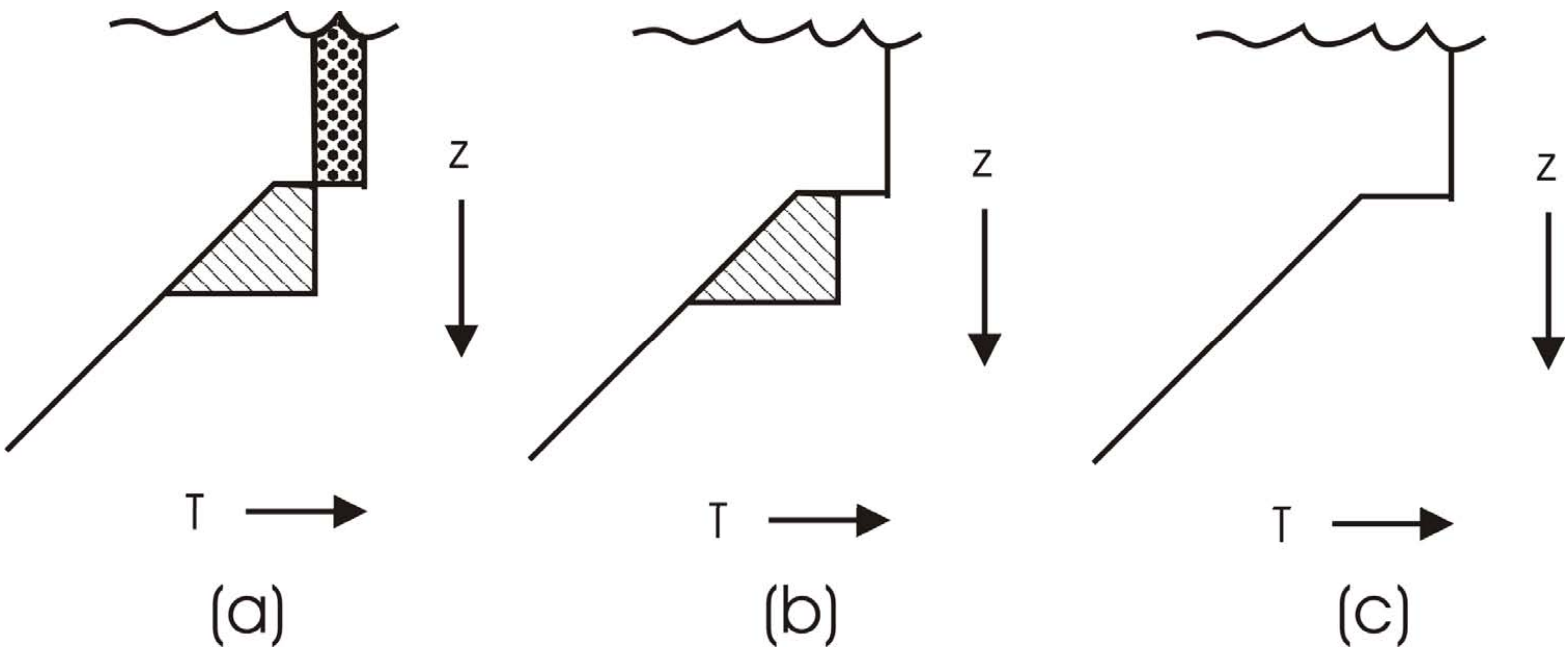
$$\omega^2 = N^2 \frac{\lambda^2}{\lambda^2 + r^2} + f^2 \frac{r^2}{\lambda^2 + r^2},$$

$$\lambda^2 \equiv k^2 + l^2$$

$$\omega^2 \cong f^2 \quad for \quad r^2 >> \frac{N^2}{f^2} \lambda^2$$


N
A
B
A
B
A
B
(c)
(b)
(a)

Mixing and Entrainment:



Entrainment Formulation:

Criticality of a Bulk Richardson Number:

$$Ri \equiv \frac{gh\Delta\rho}{\rho u^2}$$

Assume that density jump is what would result from eroding a constant background stratification down to depth *h* :

$$Ri \equiv \frac{1}{2}\frac{h^2 N^2}{u^2}$$

Equivalently,

$$\boxed{Nh = R'u} \qquad (1)$$

$$R' \equiv \sqrt{2Ri}$$

Criticality assumption: *R' = constant.*

Mixed layer momentum conservation (neglecting Coriolis turning) :

$$\frac{\partial\left(\rho_w uh\right)}{\partial t} = \rho_a u_*^2 . \qquad (2)$$

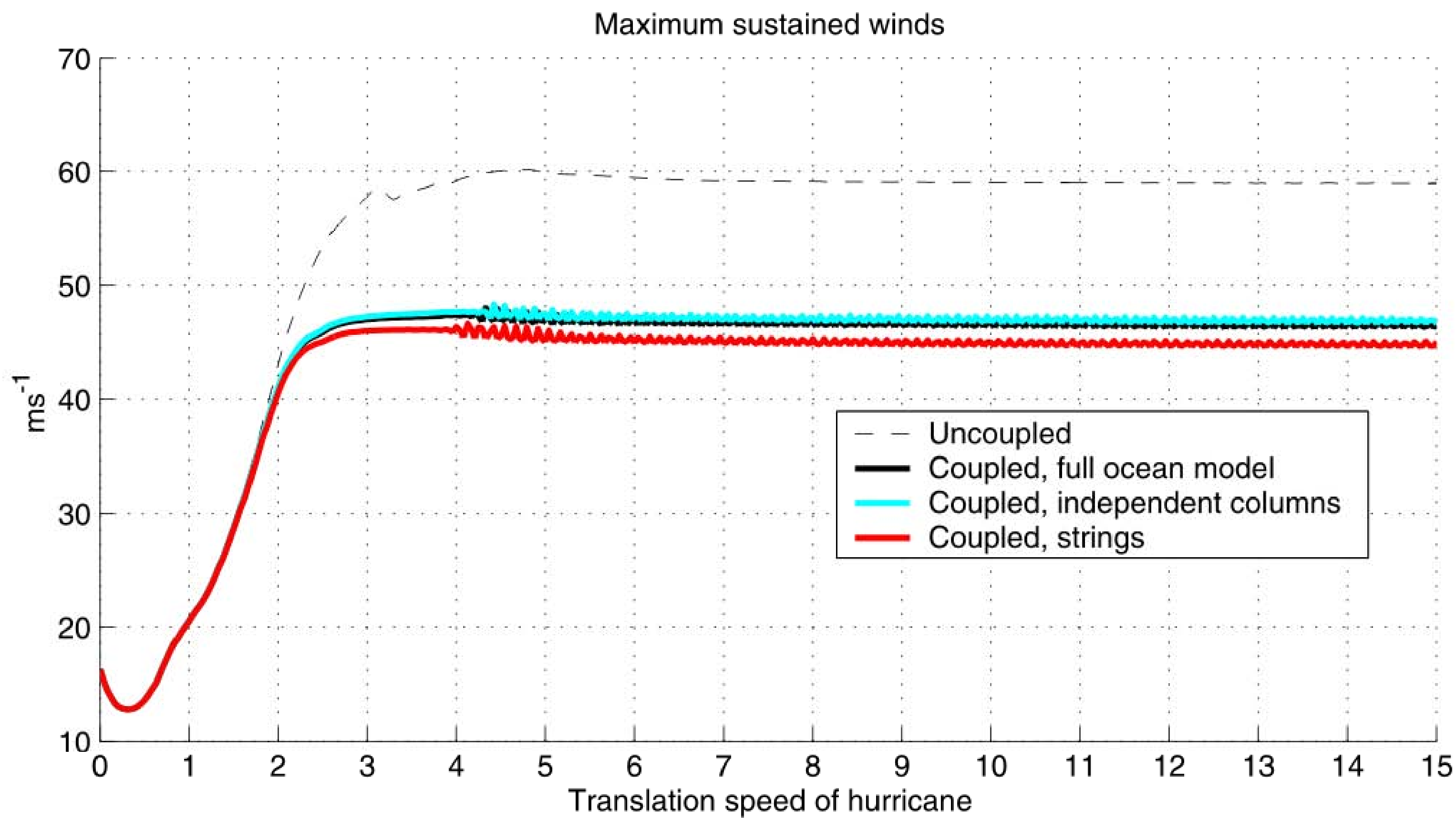
$$u_*^2 \equiv C_D \mid \mathbf{V} \mid^2$$

Combine (2) with (1):

$$\frac{\partial h^2}{\partial t} = R' \frac{\rho_a}{\rho_w} \frac{u_*^2}{N}$$

Note: units of diffusivity

Comparison with same atmospheric model coupled to 3-D ocean model; idealized runs:
Full model (black), string model (red)



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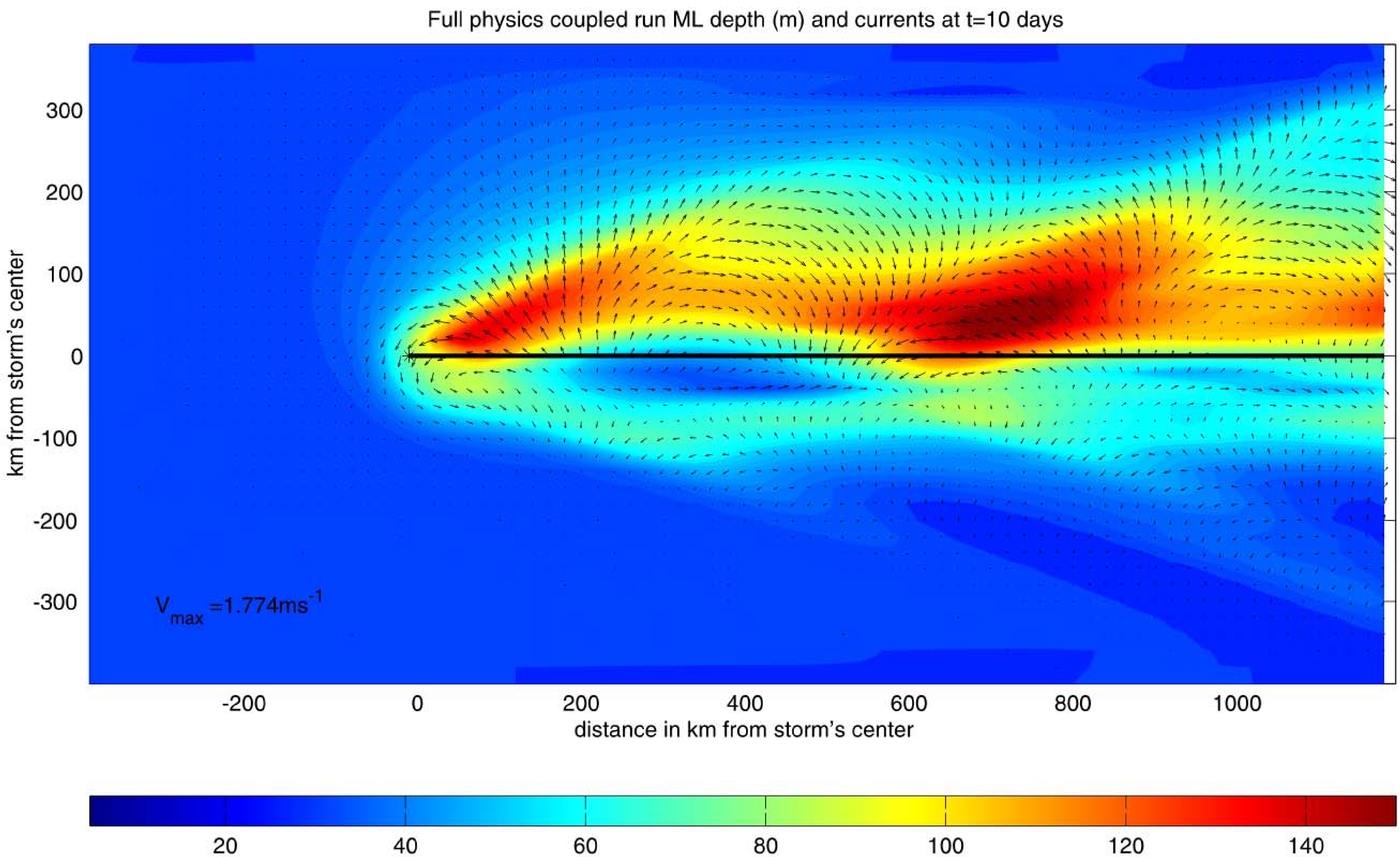
Mixed layer depth and currents



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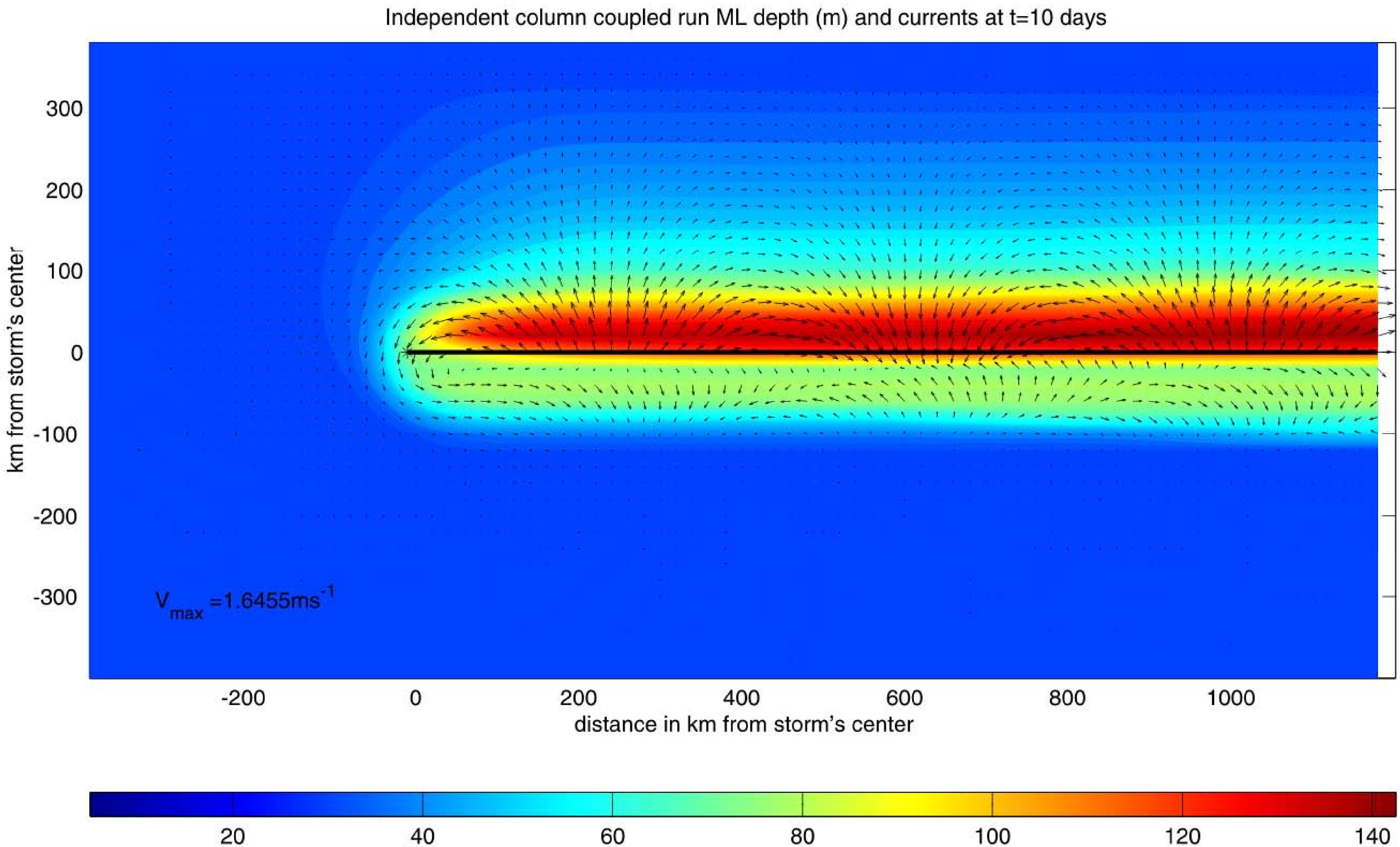

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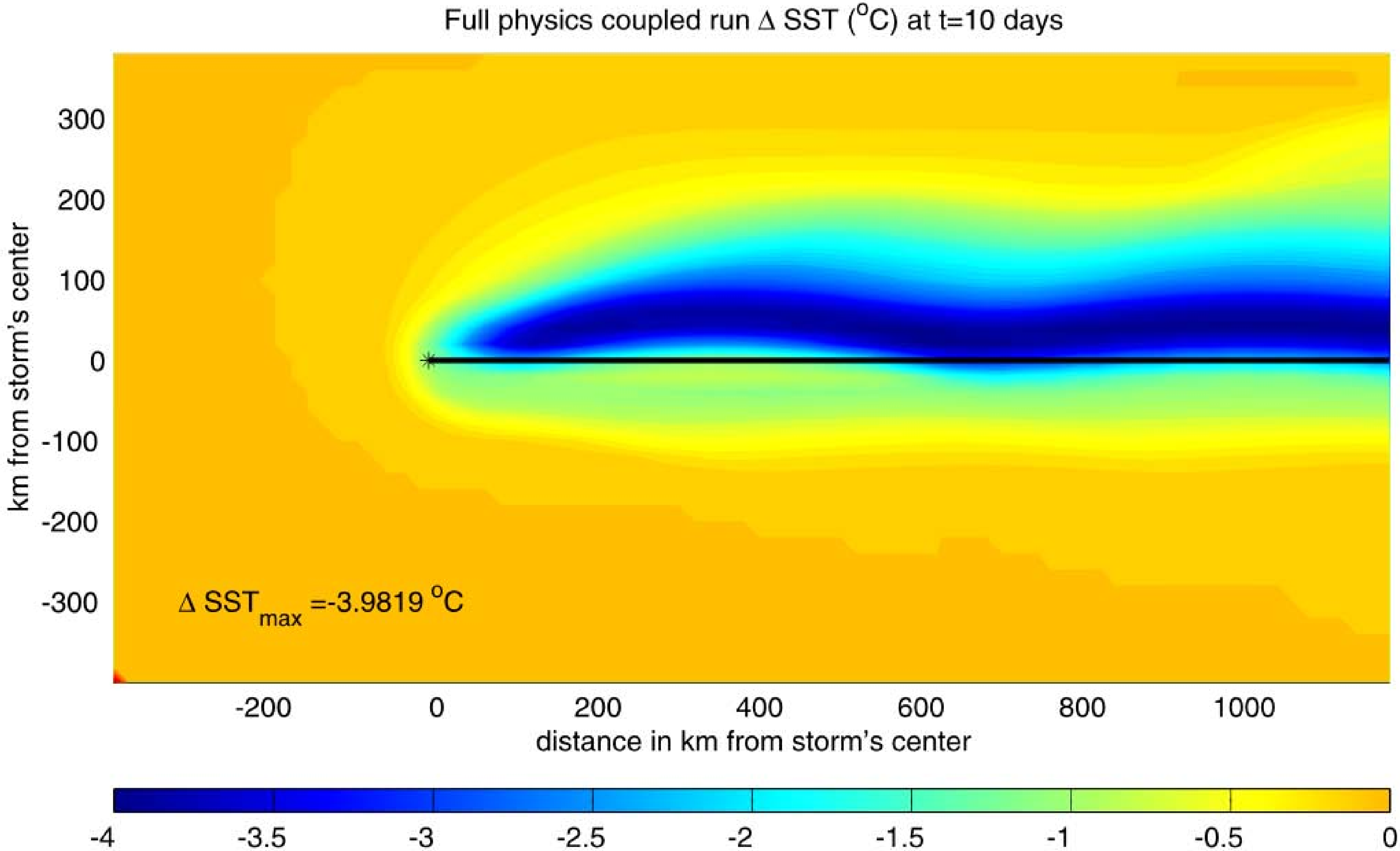
SST Change



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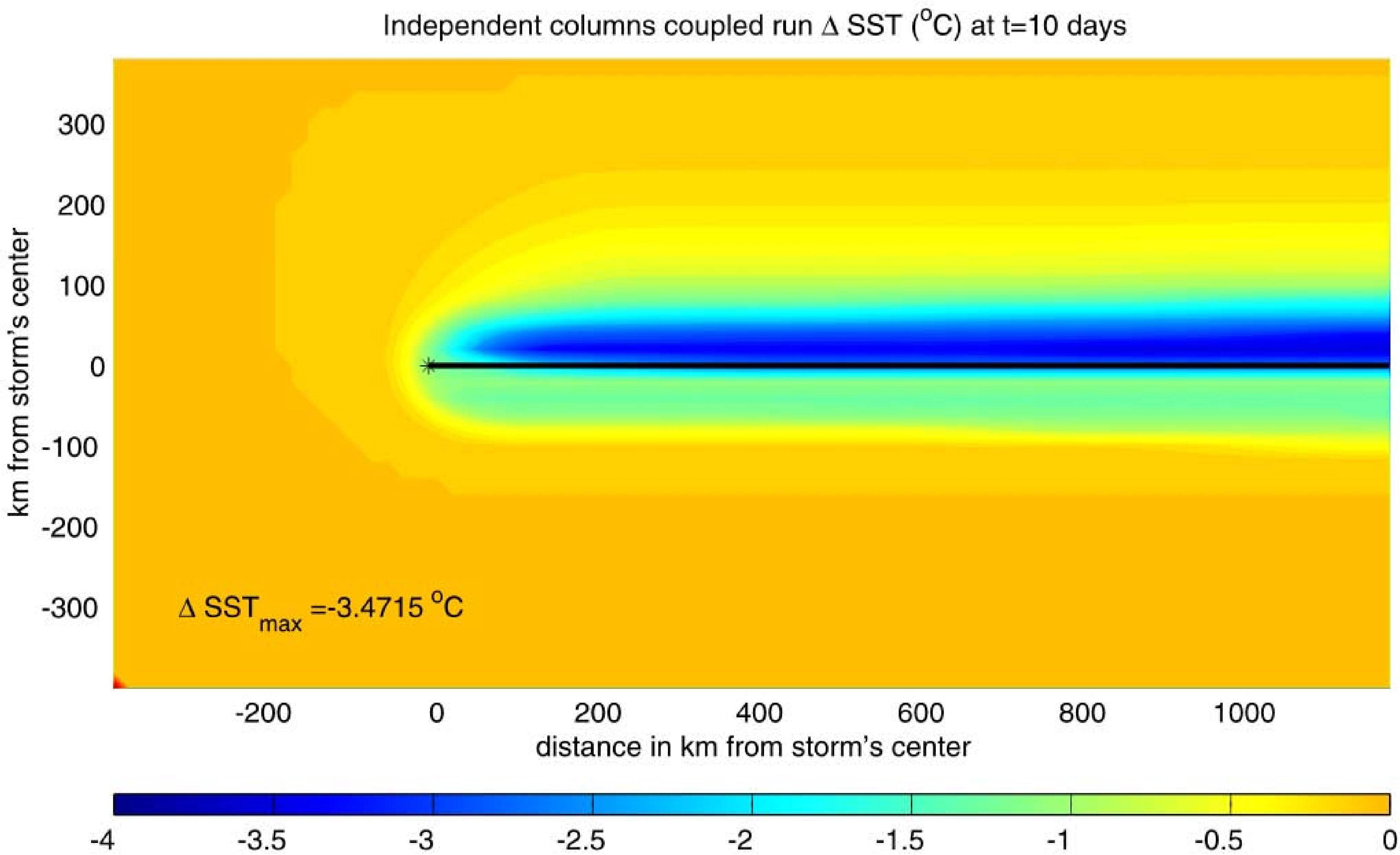

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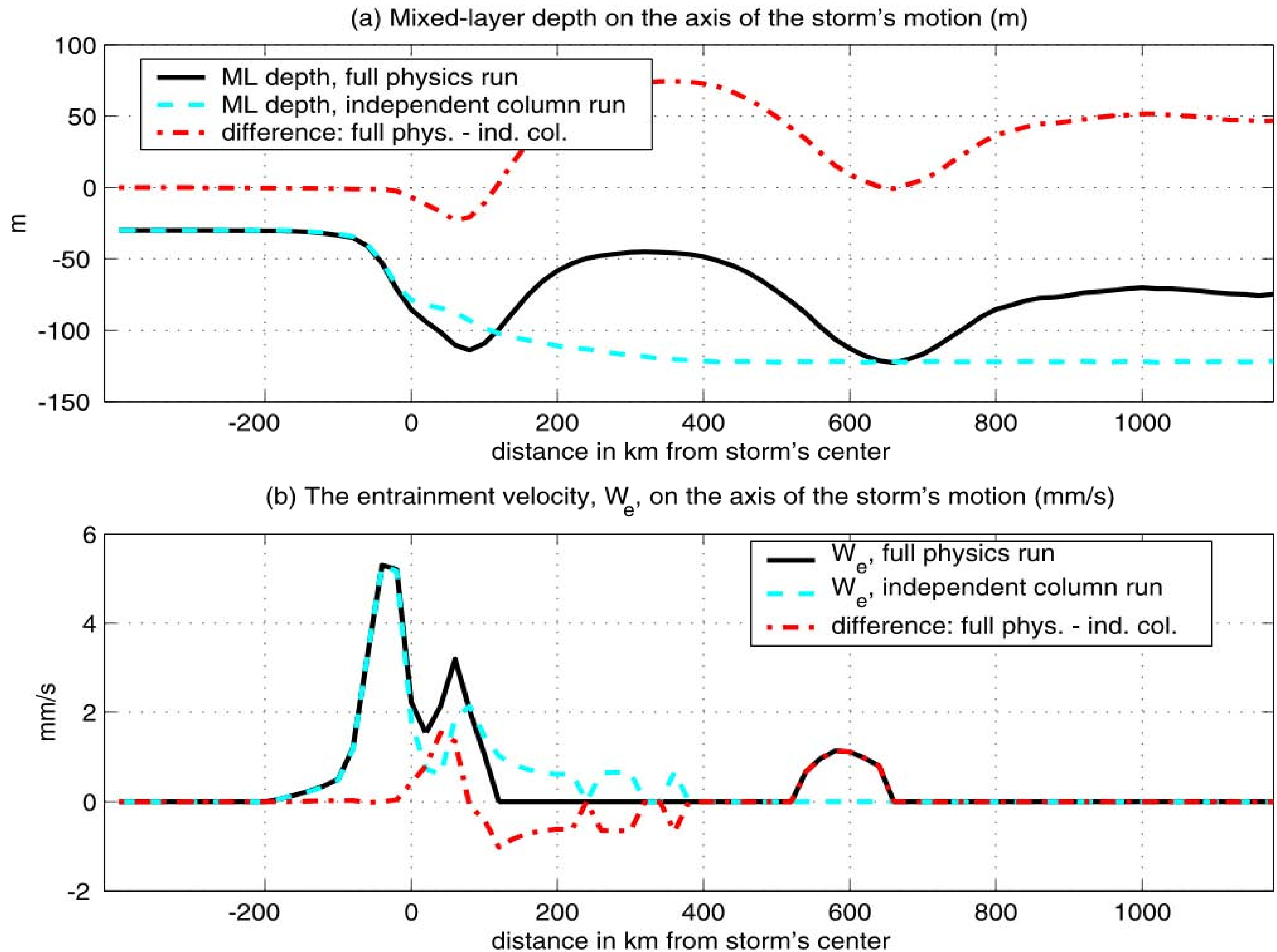

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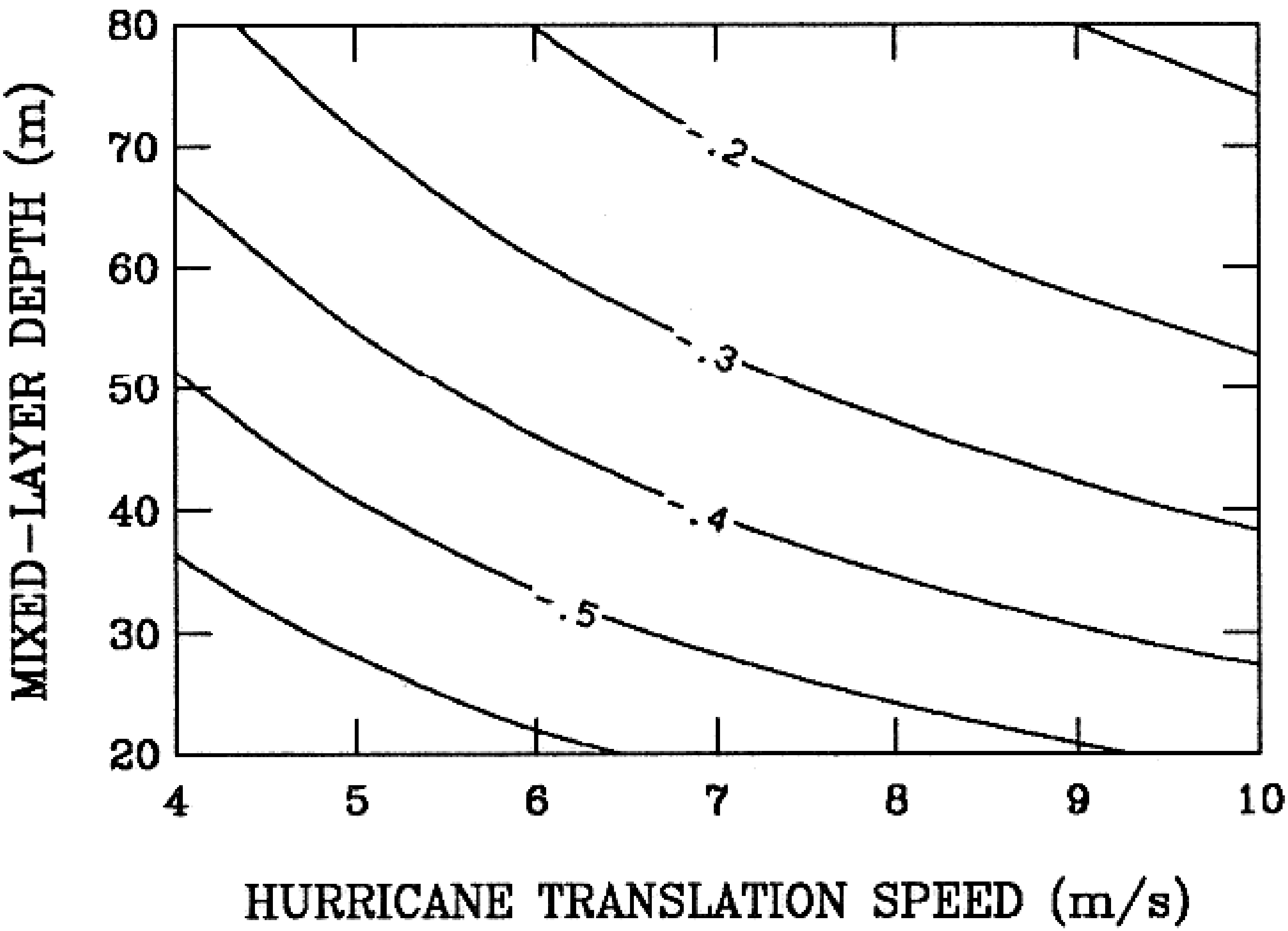
Define feedback factor:

$$F_{SST} = \frac{\Delta p}{\Delta p\,|_{SST}} - 1,$$

where $\Delta p\,|_{SST}$ is the central pressure drop at fixed SST. Do many, many numerical expreiments, varying SST, Coriolis parameter, traslantion speed, etc. Curve fit dependence of F_{SST} on these parameters. Result:



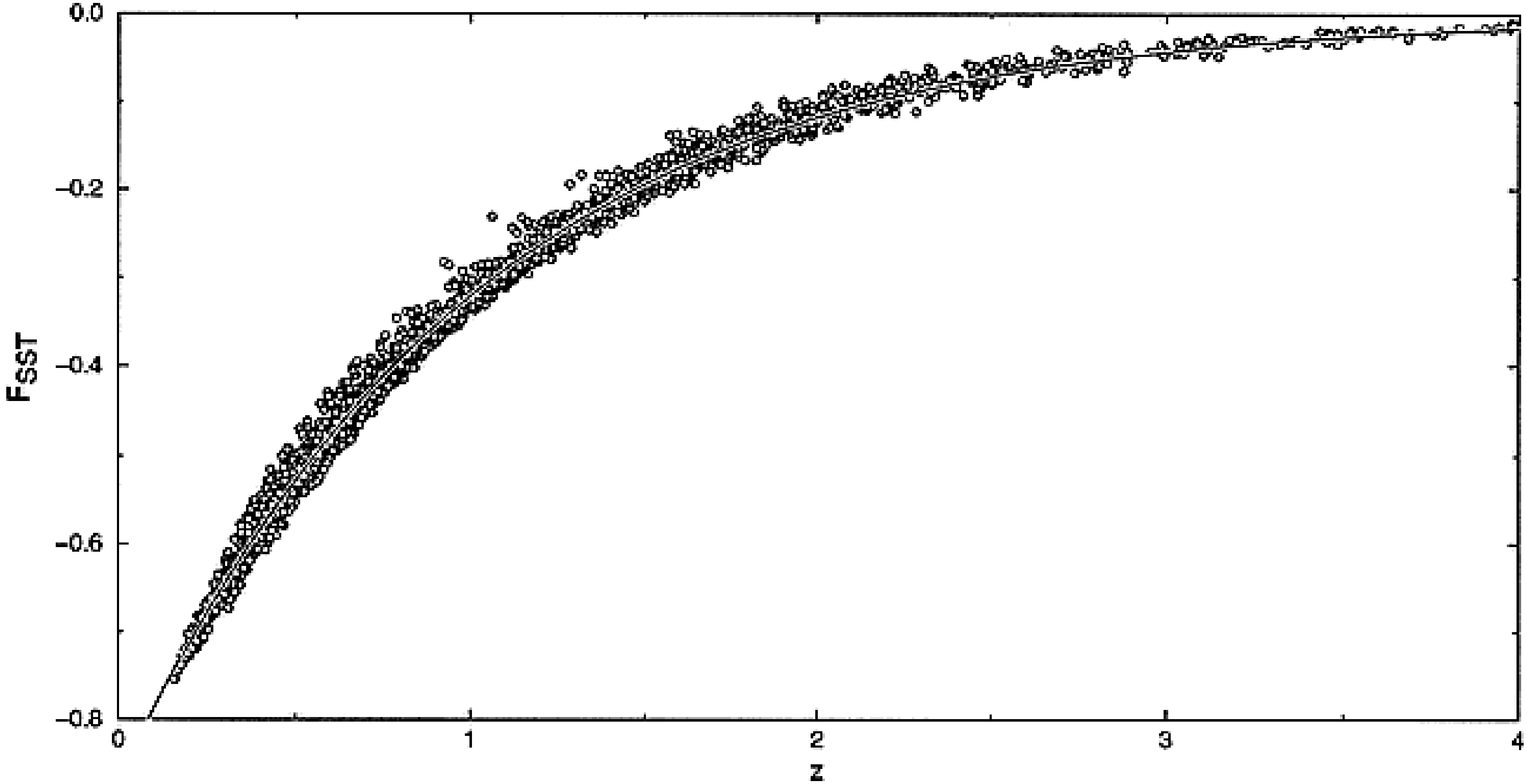
80
70
60
50
40
30
20
MIXED-LAYER DEPTH (m)
.2
.3
.4
.5
4
5
6
7
8
9
10
HURRICANE TRANSLATION SPEED (m/s)

$$F_{SST} = -0.87 e^{-z}$$

$$z \equiv 0.55 \left(\frac{h_0}{30\,m} \right)^{1.04} \left(\frac{u_T}{6\,m\,s^{-1}} \right)^{0.97} \left(\frac{\Delta p\,|_{SST}}{50\,hPa} \right)^{-0.78} \times$$

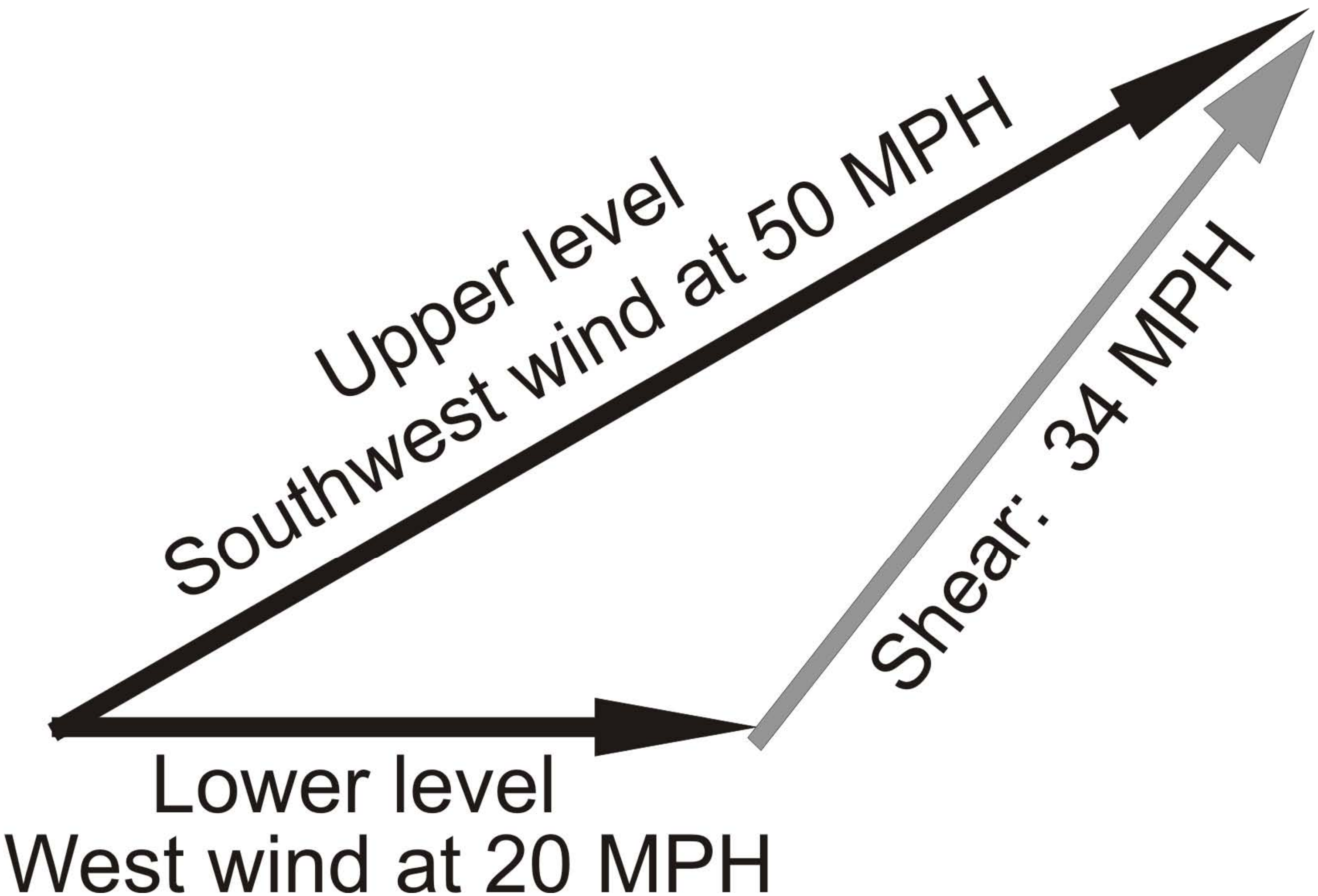
$$\eta^{-0.85} \left(\frac{f}{5 \times 10^{-5}\, s^{-1}} \right)^{0.59} \left(\frac{\Gamma}{8 \times 10^{-2}\, K\, m^{-1}} \right)^{-0.40} \left(\frac{1 - \mathcal{H}}{0.2} \right)^{0.46}$$

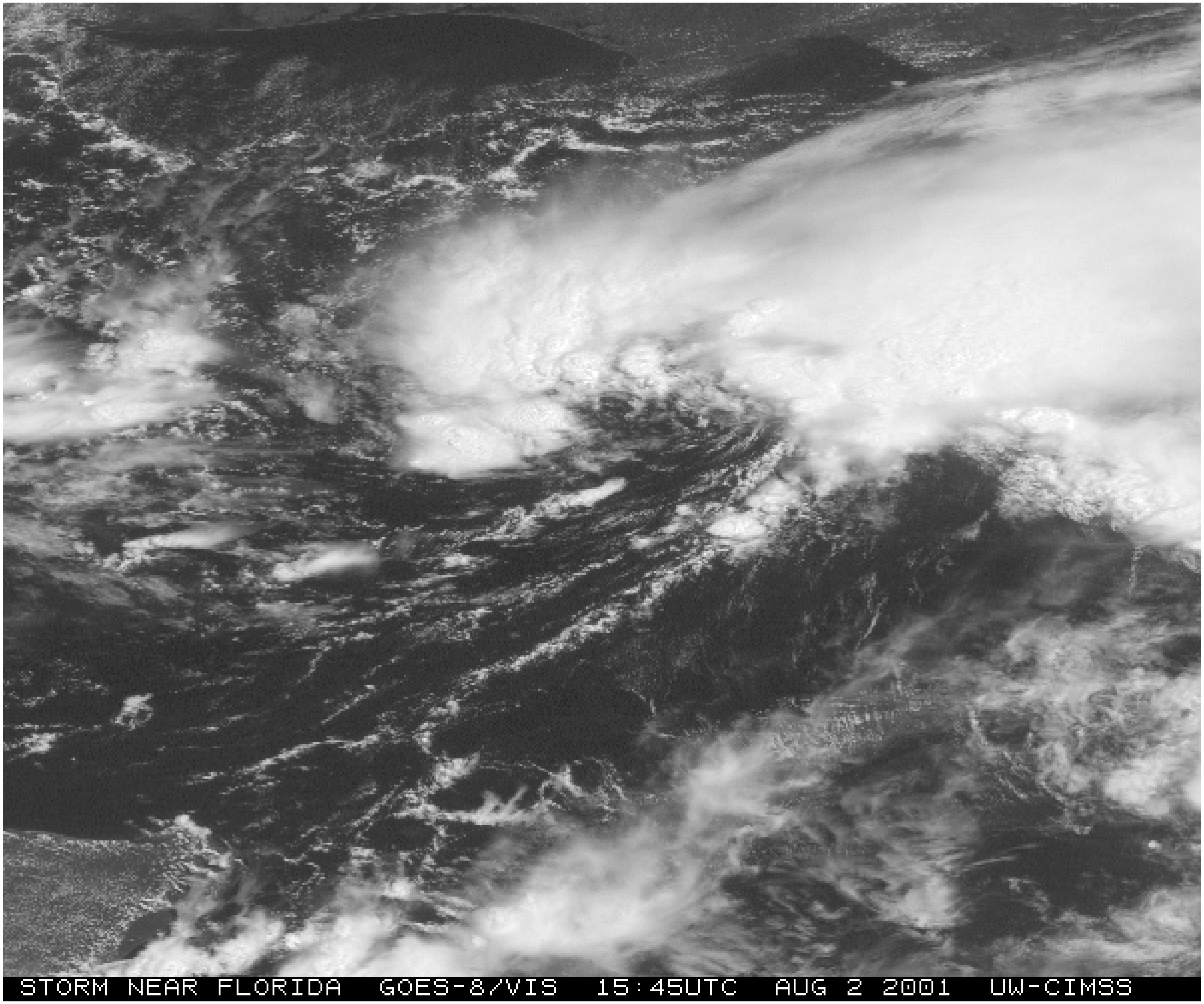
η= storm size scaling factor


0.0
-0.2
-0.4
-0.6
-0.8
F_SST
0
1
2
3
4
z

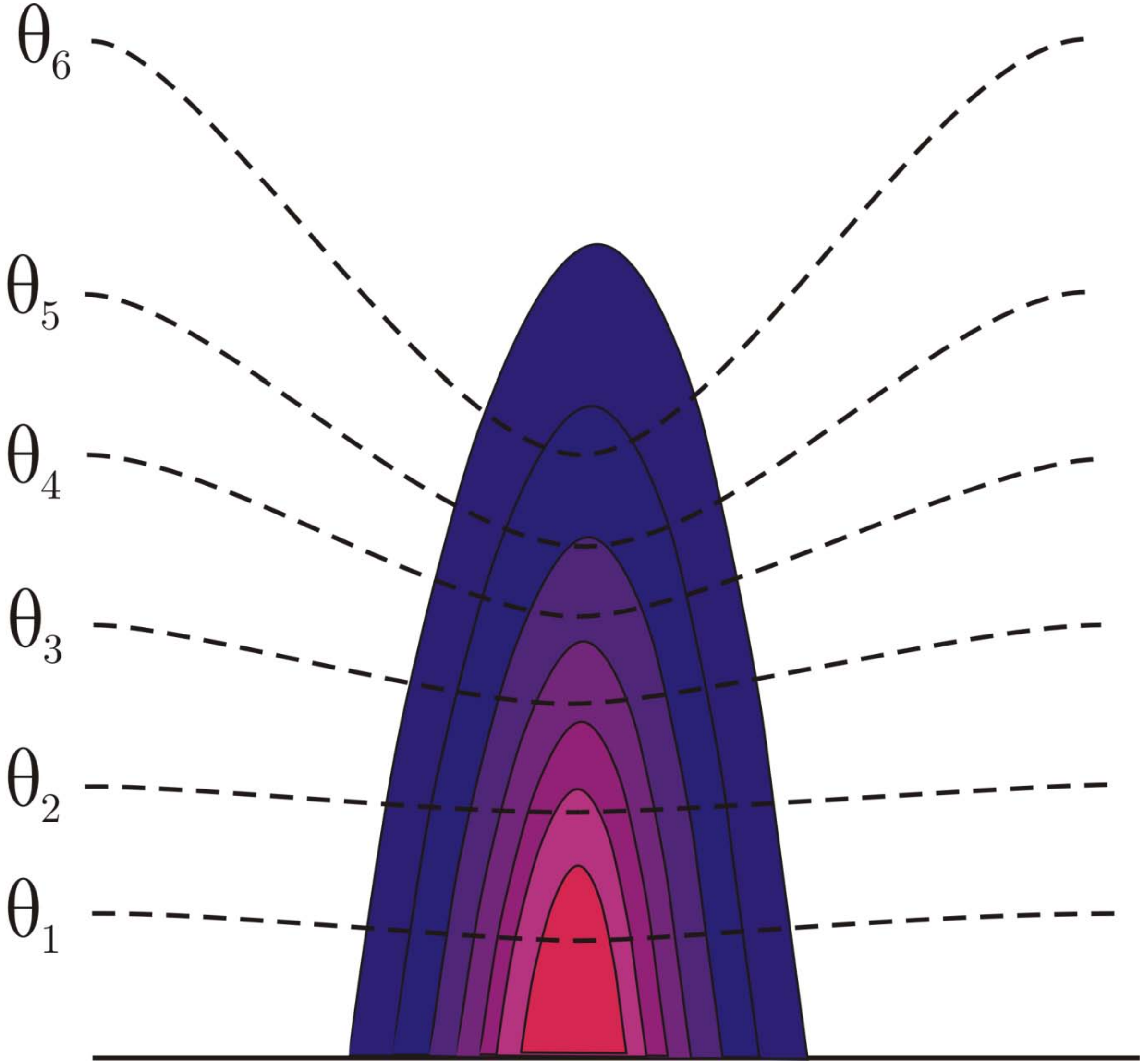
Effects of Environmental Wind Shear

- **Dynamical effects**
- **Thermodynamic effects**
- **Net effect on intensity**

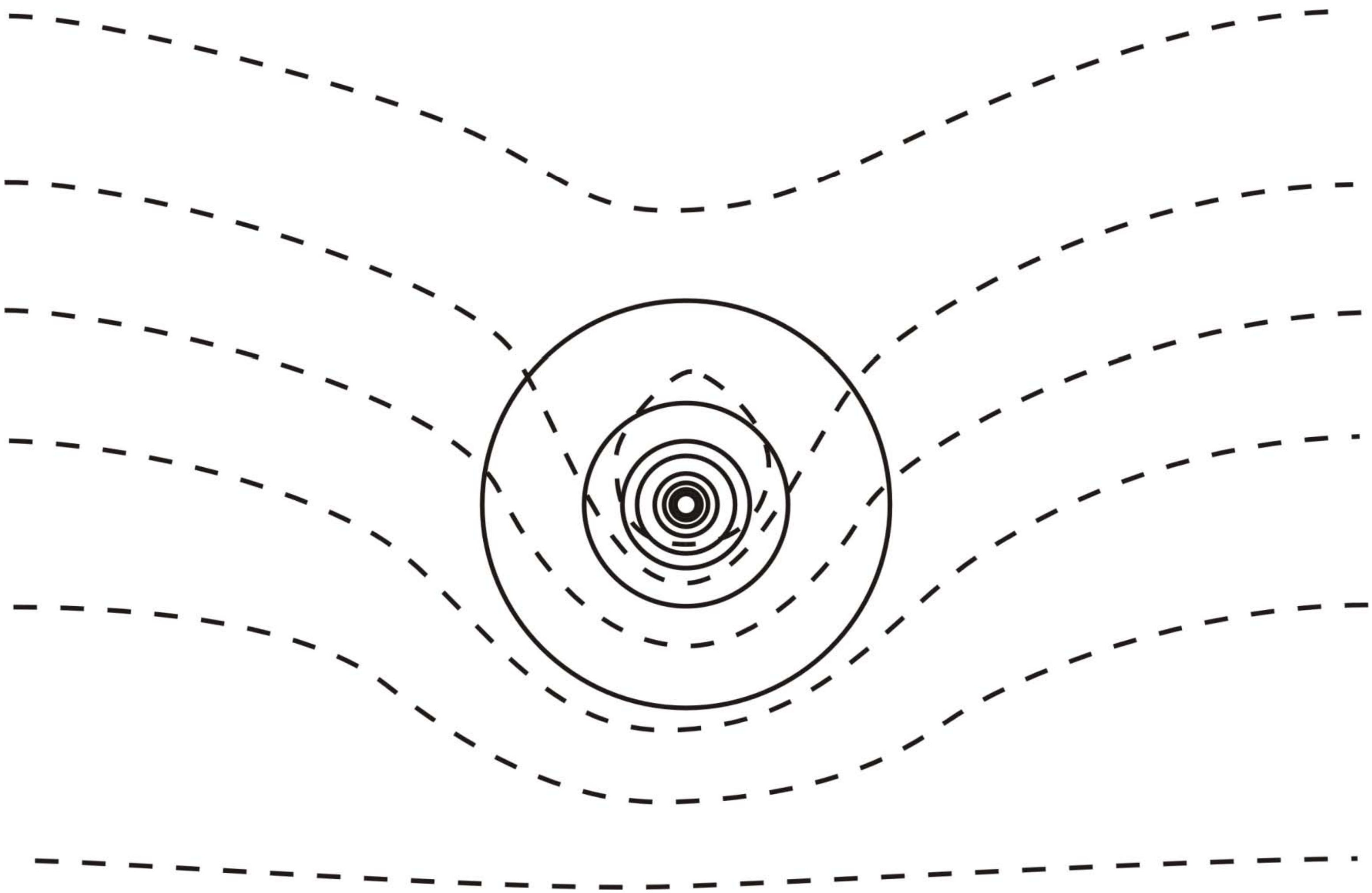

Upper level
Southwest wind at 50 MPH
Shear: 34 MPH
Lower level
West wind at 20 MPH

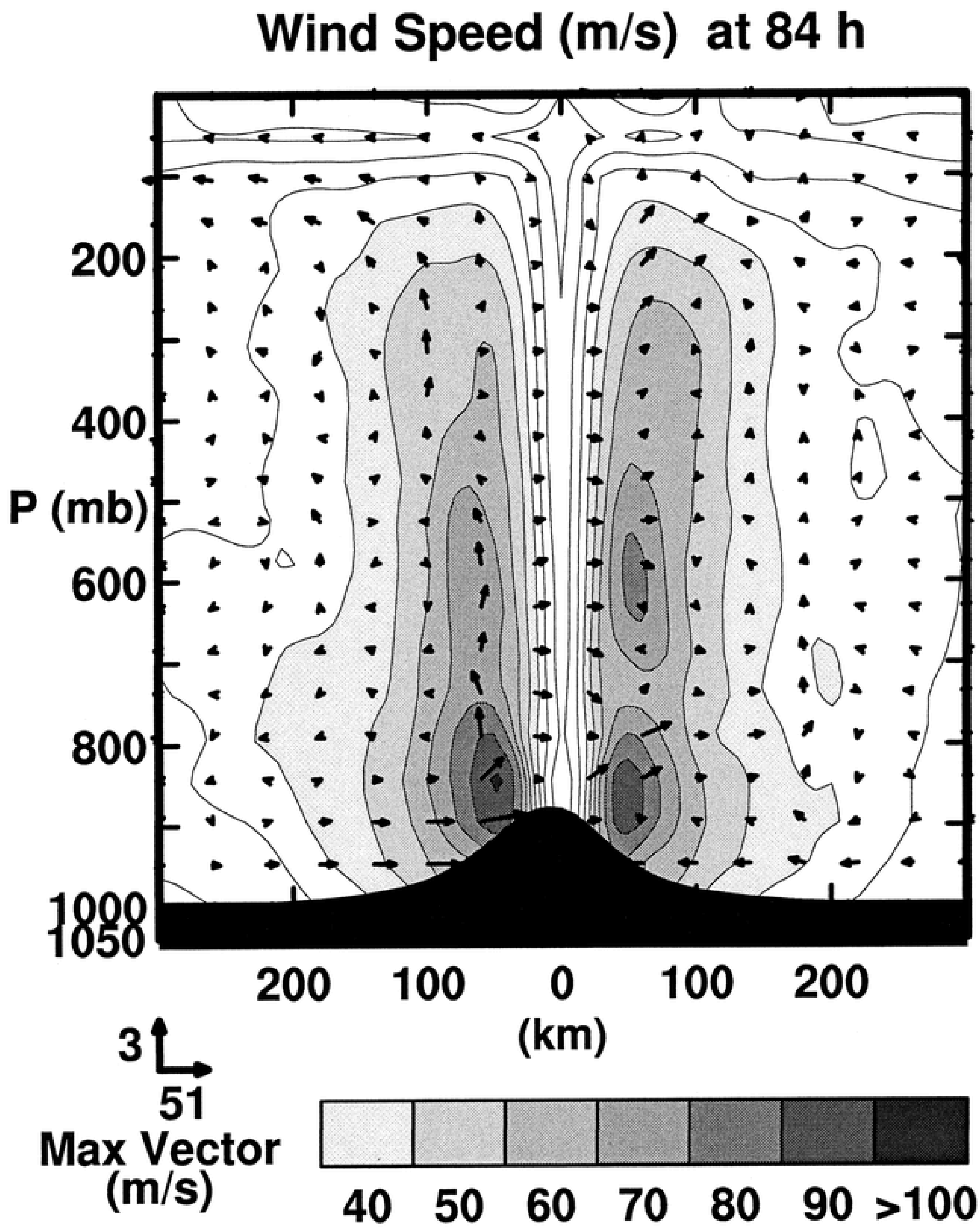

STORM NEAR FLORIDA GOES-8/VIS 15:45UTC AUG 2 2001 UW-CIMSS

PV dynamics

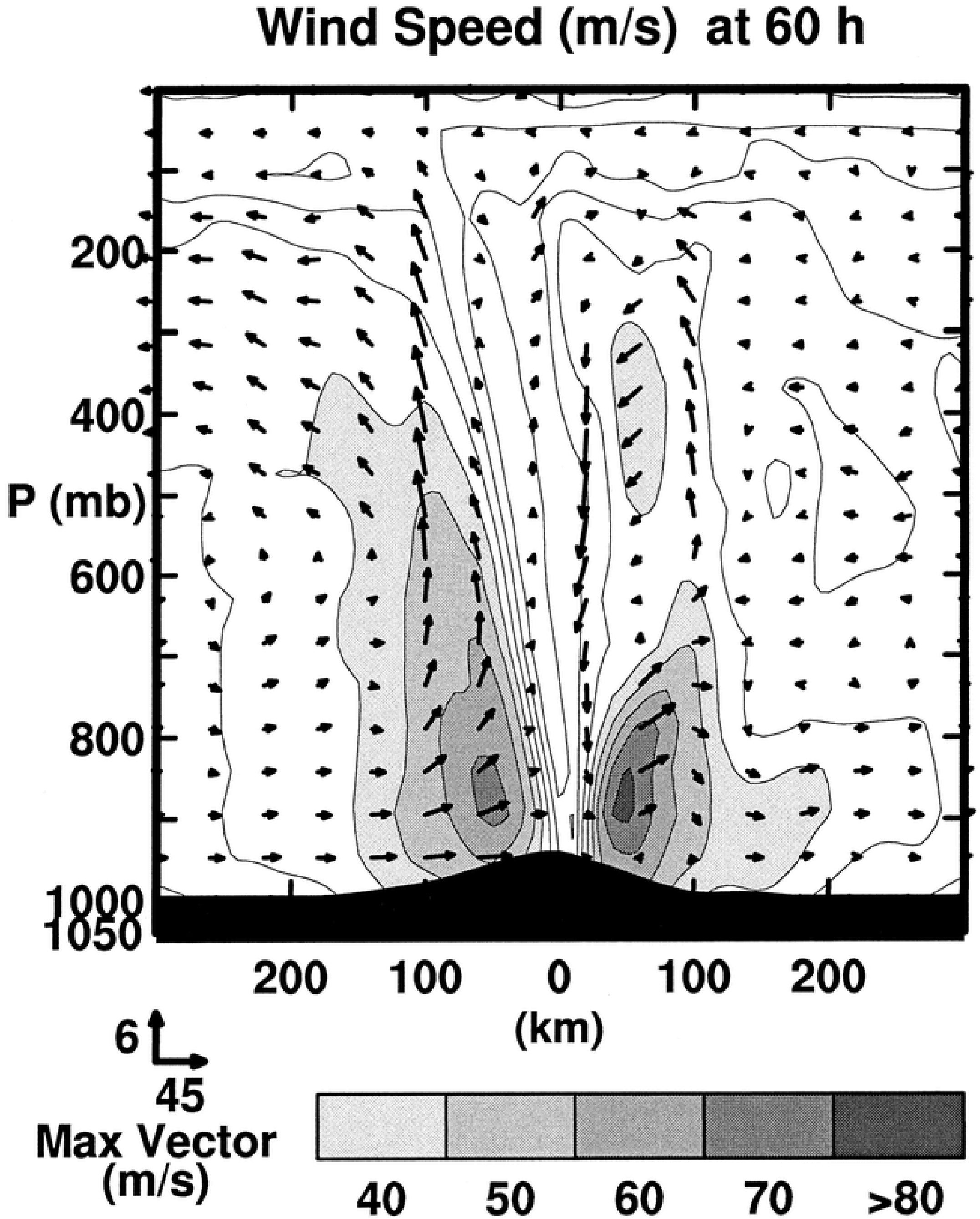


Streamlines (dashed) and θ surfaces (solid)





Wind Speed (m/s) at 84 h
P (mb)
200
400
600
800
1000
1050
200
100
0
100
200
(km)
3
51
Max Vector
(m/s)
40
50
60
70
80
90
>100


Wind Speed (m/s) at 60 h
200
400
P (mb)
600
800
1000
1050
200 100 0 100 200
(km)
6
45
Max Vector
(m/s)
40
50
60
70
>80

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Secondary circulation of a idealized TC (a), along with the legs of the secondary circulation represented on a entropy-temperature diagram (b). TC without ventilation travels along A-B-C-D, while a ventilated TC travels along A-B′-C′-D. Hatched region in (b) denotes the work lost due to ventilation. From Tang(2010)

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Normalized equilibrium solutions (solid and dashed lines) for the steady state intensity of a ventilated TC. Arrows denote intensifying and weakening TCs for off-equilibrium values of intensity and ventilation. From Tang (2010).

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(a) Jul.-Oct. ventilation index for the Northern Hemisphere and (b) Dec.-Mar. ventilation index for the Southern Hemisphere averaged over 1981-2000. Results are shown as the $\log_{10}$(VI). From Tang (2010).

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Percentage of days with a VI below 0.1 for (a) the Northern Hemisphere TC season and (b) the Southern Hemisphere TC season during 1981-2000 (shaded with contours every 10%). TC genesis points for the same period are denoted by black dots. From Tang (2010).

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Grayscale shading indicates the number of daily TC observations in the MGR as a function of the VI and normalized intensity, i.e. the intensity divided by the potential intensity. Arrows signify the mean 24 hour normalized intensity change for TCs in each bin with green and red arrows indicating normalized strengthening and weakening, respectively. The maximum arrow length corresponds to a normalized intensity change of 0.4 over 24 hours. From Tang (2010).

Model
(Rappin, 2010)

WRF V2.2.1

- Doubly periodic domain
- Fixed SST
- 3 km horizontal resolution
- 40 vertical levels (stretched in height)
- YSU PBL scheme
  - Improved drag formulation (Donelan et al. 2004; Davis et al. 2008)
- WRF 6-species microphysics scheme
- RRTM longwave scheme
- Goddard shortwave with perpetual equinox

Methodology

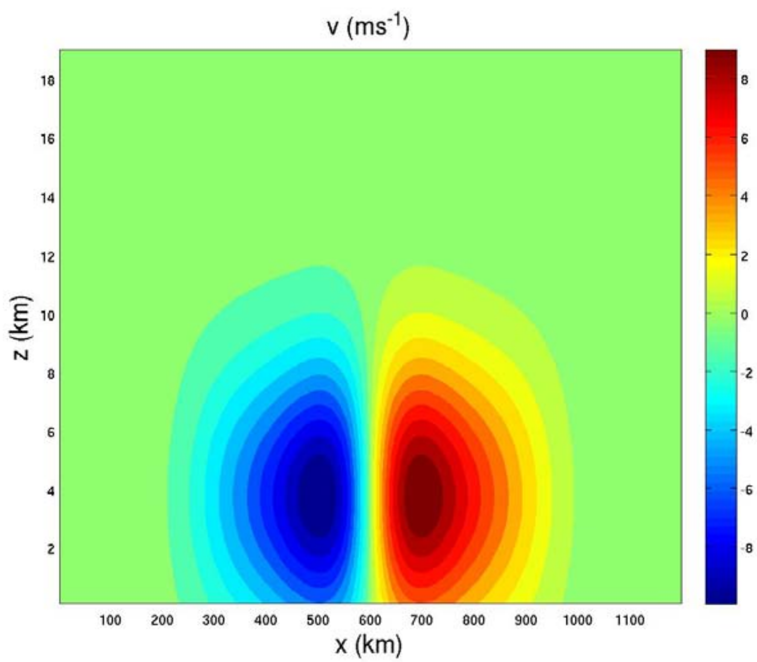
1. Fix SST, vertical shear, Coriolis parameter, and mean surface wind.

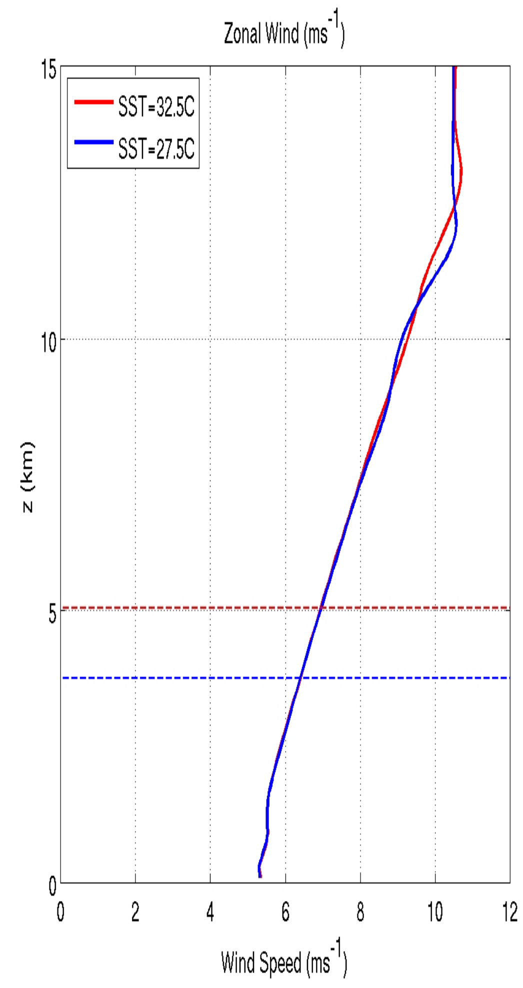
2. Run a small domain (150 km x 150 km) simulation with random low-level thermal perturbations to radiative-convective equilibrium (90 days).

3. Calculate RCE thermodynamic soundings and wind profiles. Calculate large scale quantities (i.e. MPI, CAPE, χ…).

4. Use RCE data to initialize large domain (1200 km x 1200 km) simulation of tropical cyclogenesis from a mid-level vortex modeled after easterly waves.

Initial Condition



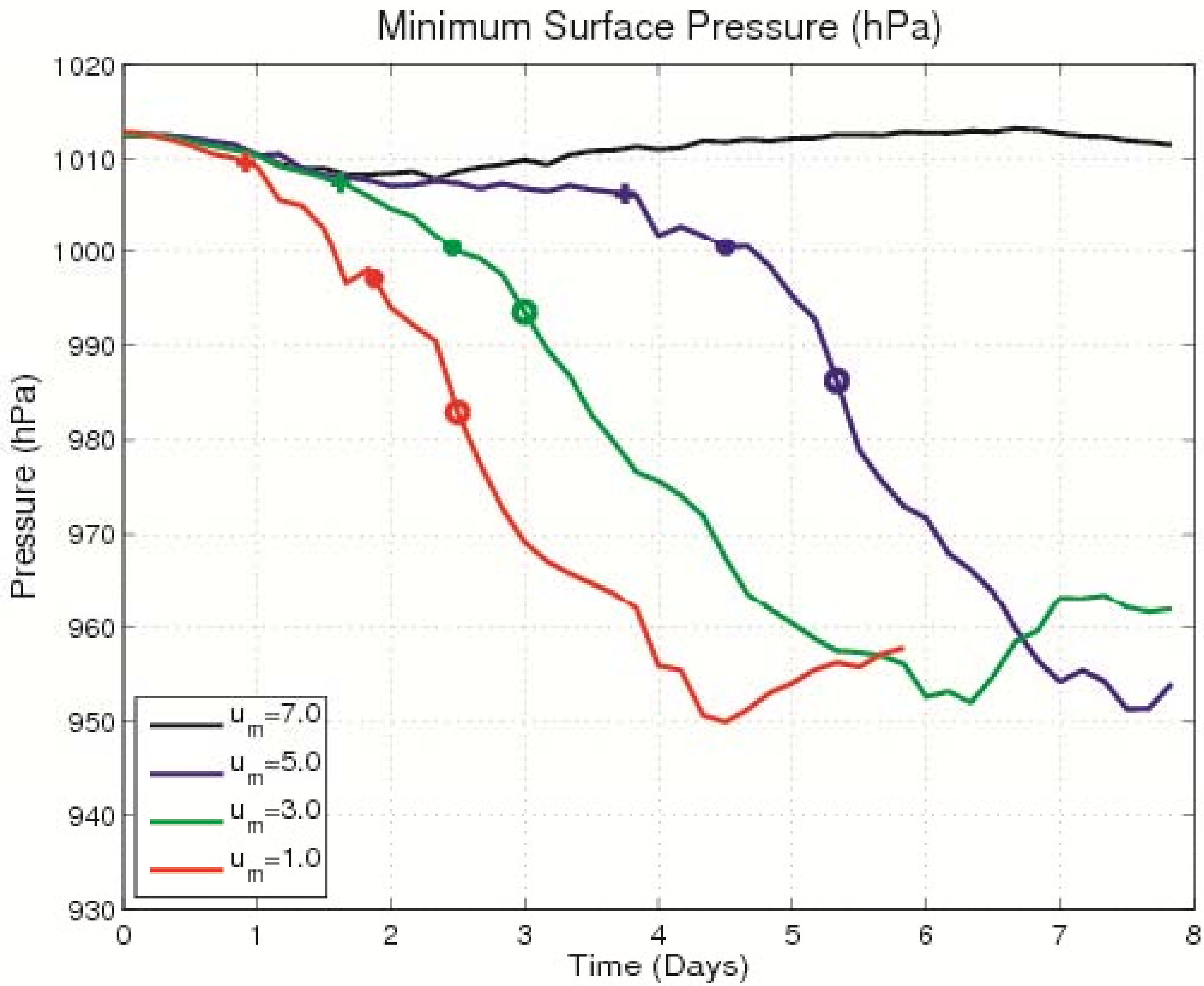


Vertical Shear

How do we incorporate shear into a doubly periodic domain?

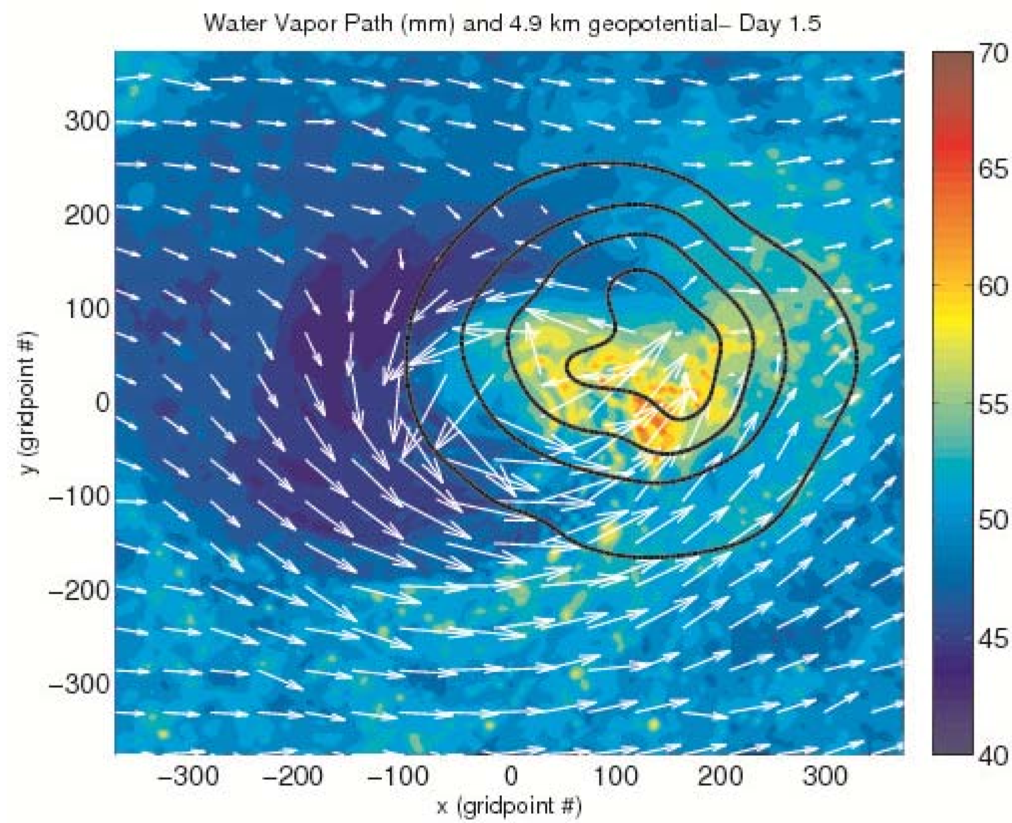
1. Initialize domain with RCE wind profile U(z).
2. Add a term to the pressure gradient that balances U(z):

$$\frac{Dv}{Dt} + fu = -\frac{\partial \Phi}{\partial y} + fU(z)$$



Minimum Surface Pressure (hPa)
Pressure (hPa)
1020
1010
1000
990
980
970
960
950
940
930
0
1
2
3
4
5
6
7
8
Time (Days)
u m=7.0
u m=5.0
u m=3.0
u m=1.0

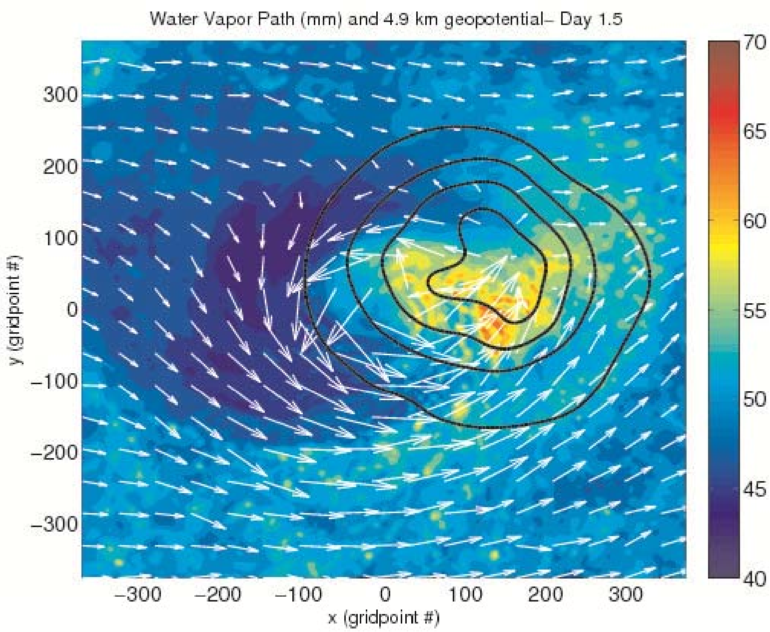
Experiment Control (Developing): WVP (colorfill), surface wind vectors, and 550 hPa geopotential height

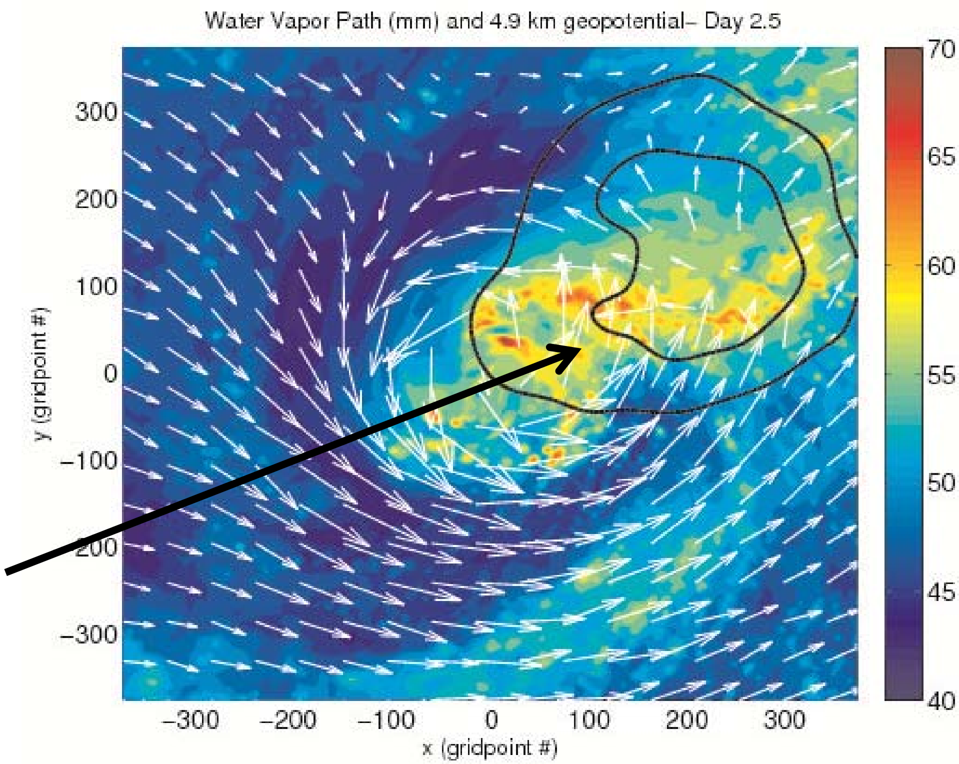


Wavenumber 1 asymmetry

Surface divergence beneath mid-level vortex

Experiment Control (Developing): WVP (colorfill), surface wind vectors, and 550 hPa geopotential height

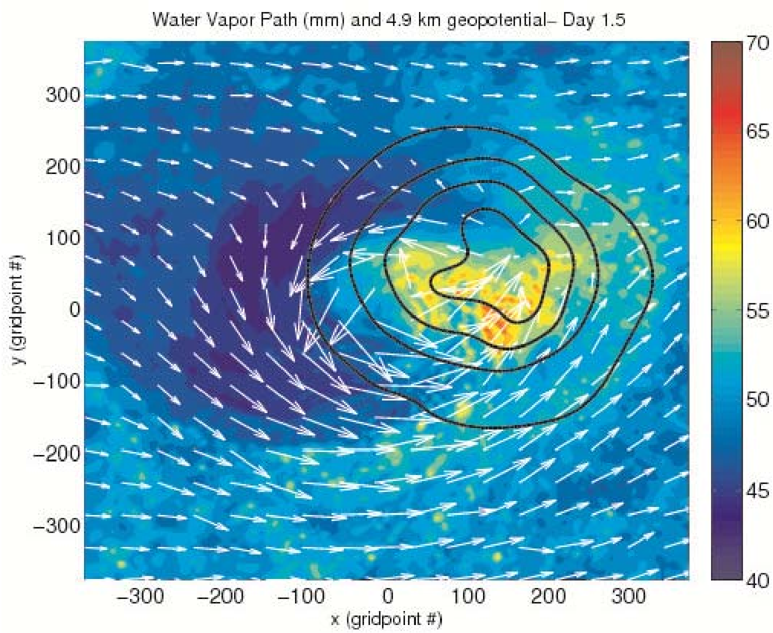


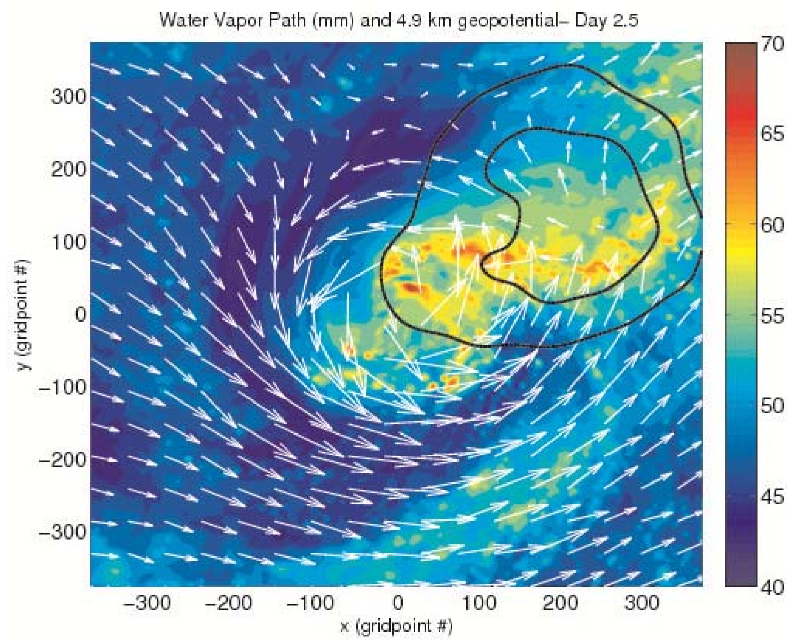


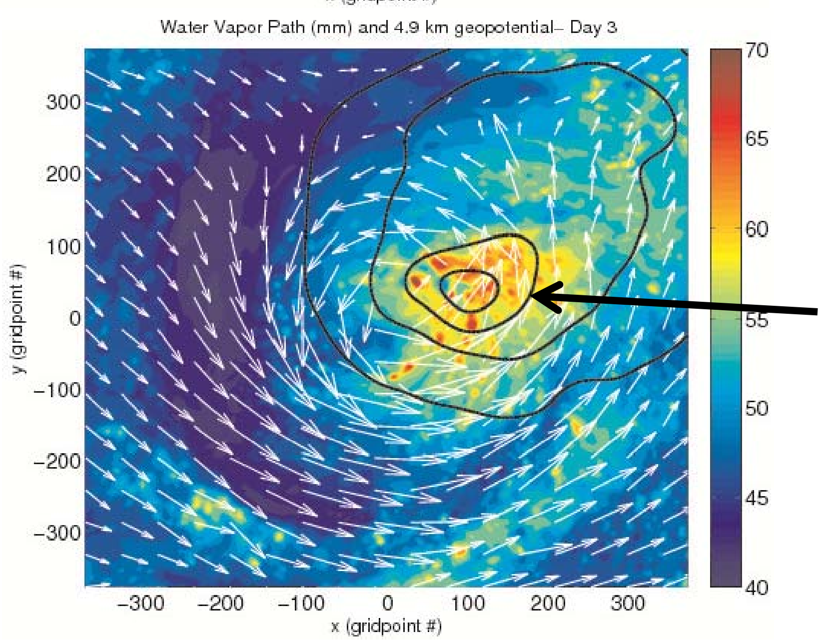
Zonal line of convection develops where the primary flow impinges upon the surface divergence field.

An MCS!

Experiment Control (Developing):
WVP (colorfill), surface wind vectors, and 550 hPa geopotential height

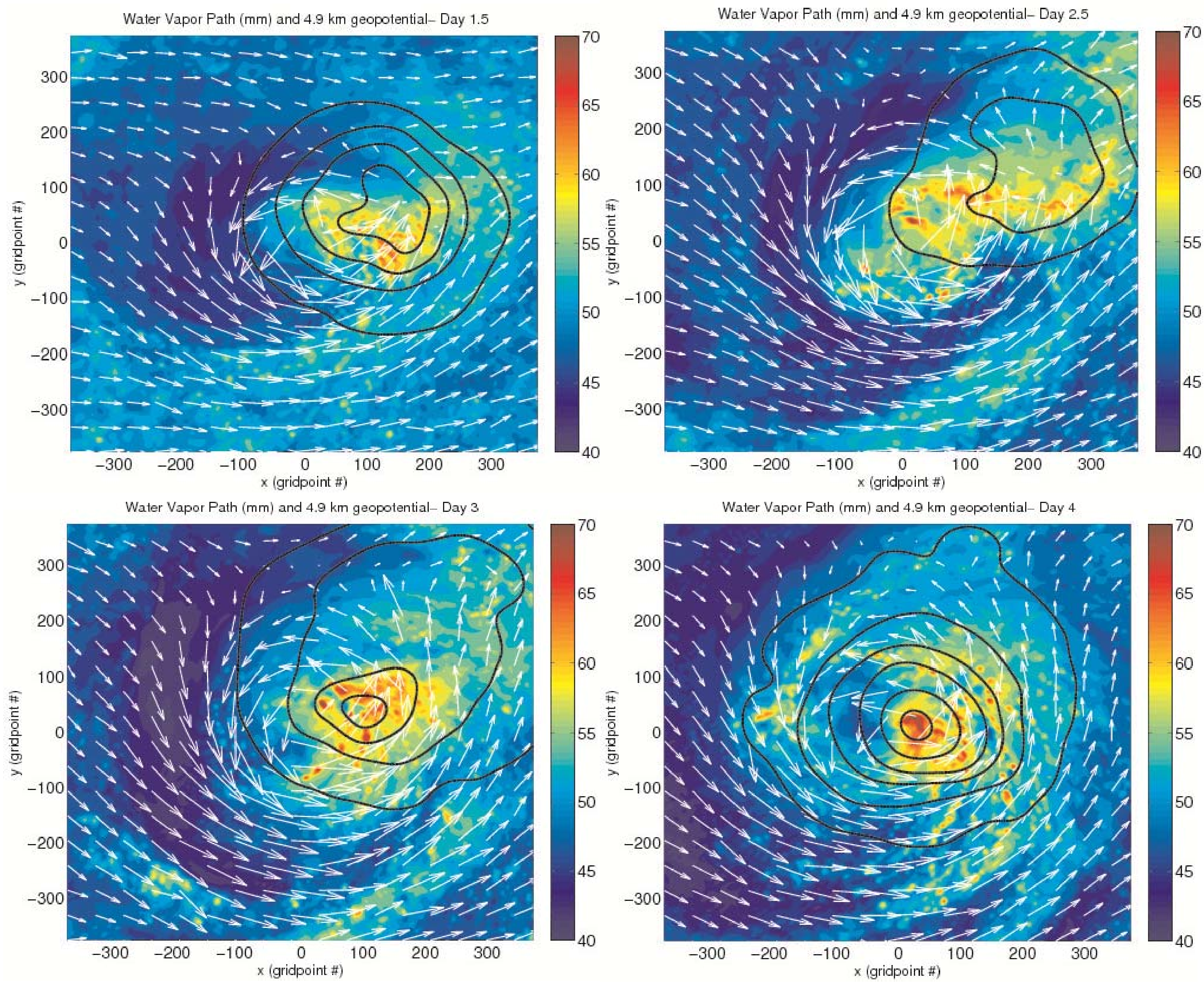




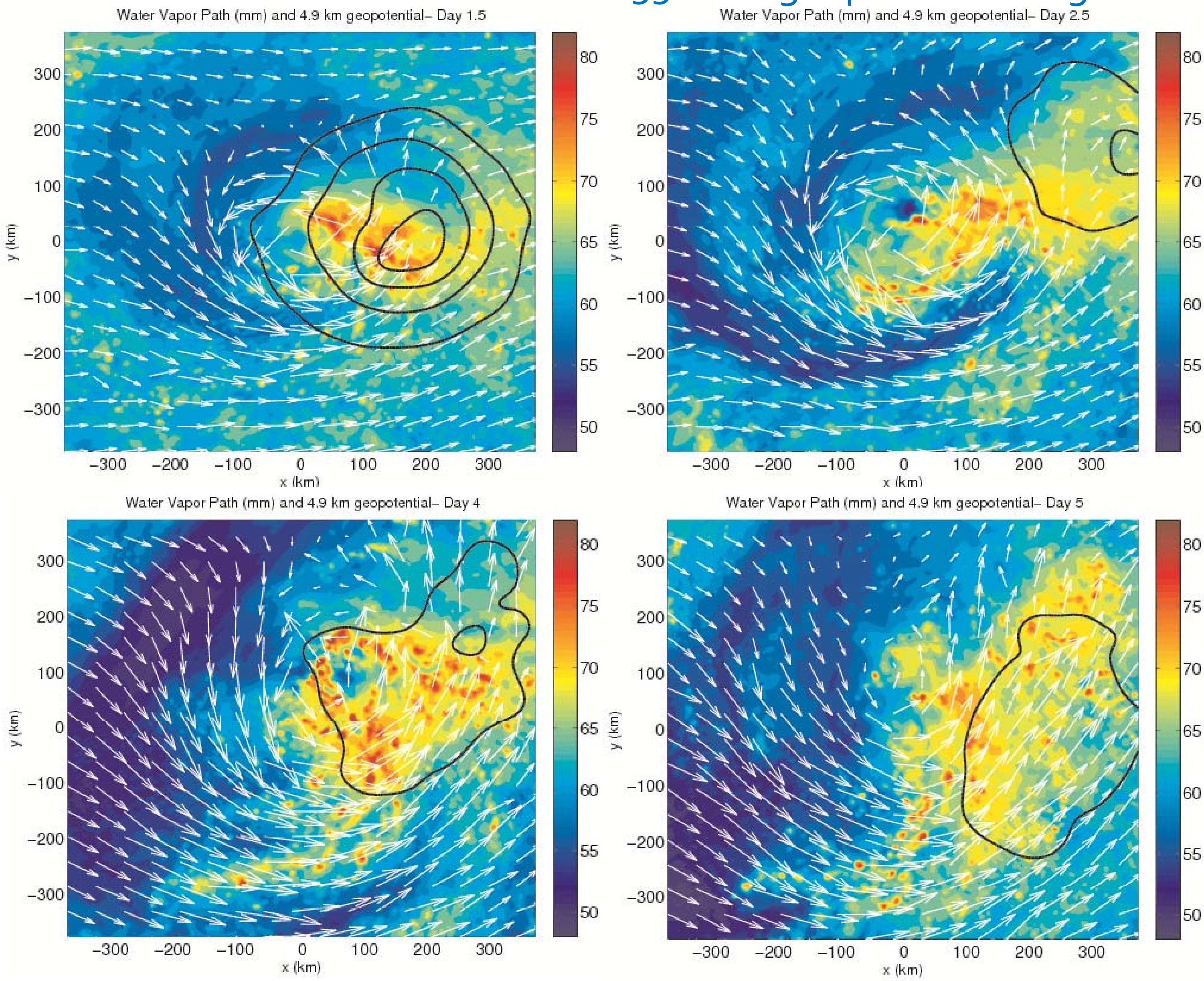


Sustained convergence on down-shear left flank results in height falls due to vortex tube stretching.

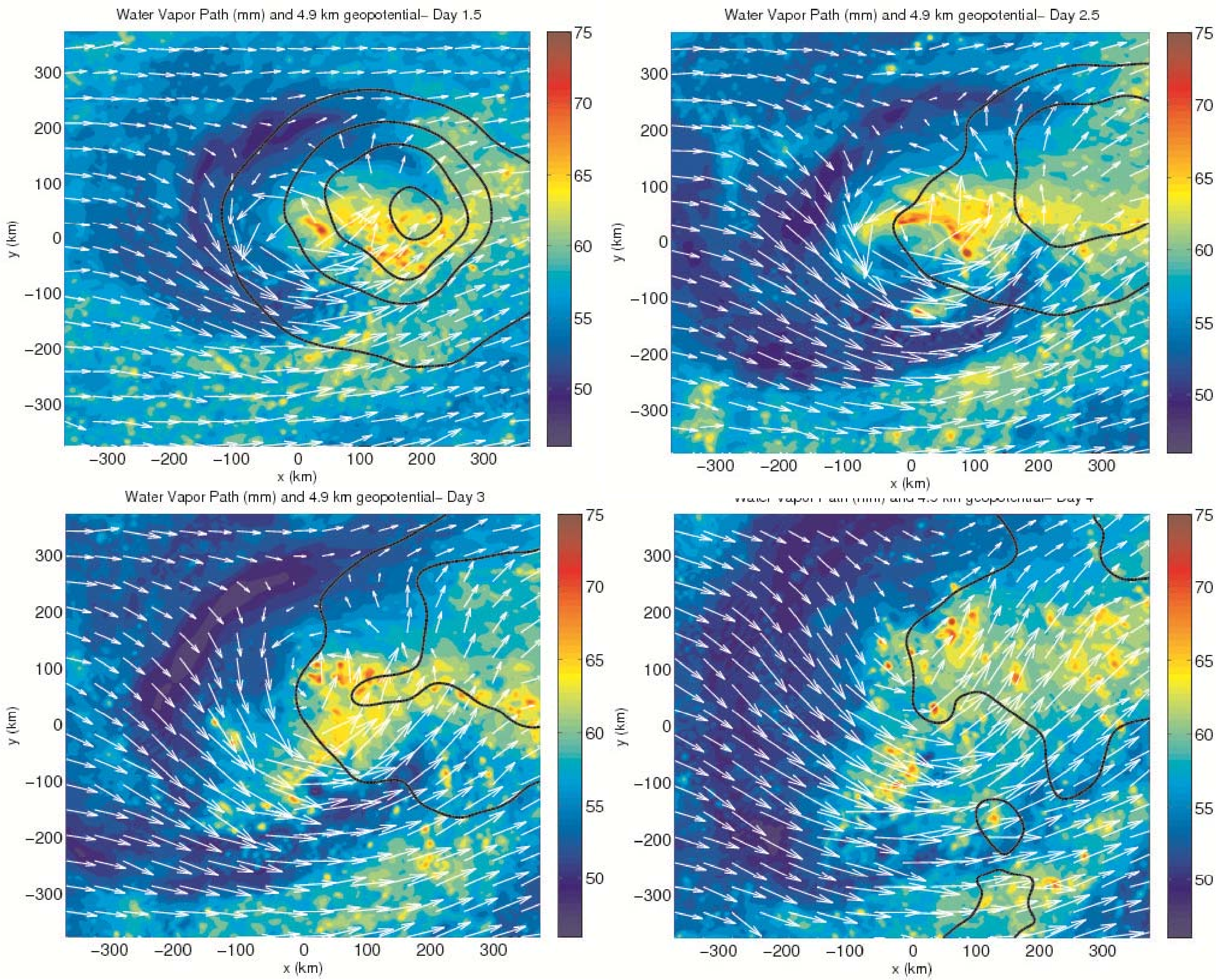
Experiment Control (Developing):
WVP (colorfill), surface wind vectors, and 550 hPa geopotential height



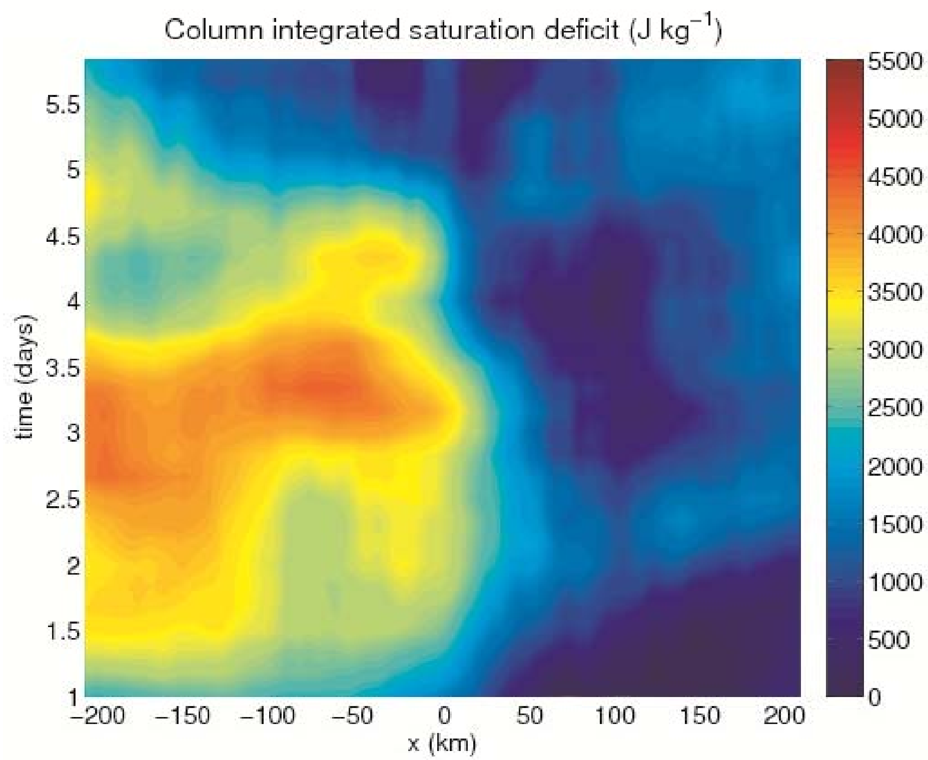
Experiment WARM (non-developing) : WVP, surface wind vectors, and 550 hPa geopotential height.



Experiment WIND (non-developing):
WVP(colorfill), surface wind vectors, and 550 hPa geopotential height.



Control: Meridionally averaged column integrated saturation deficit



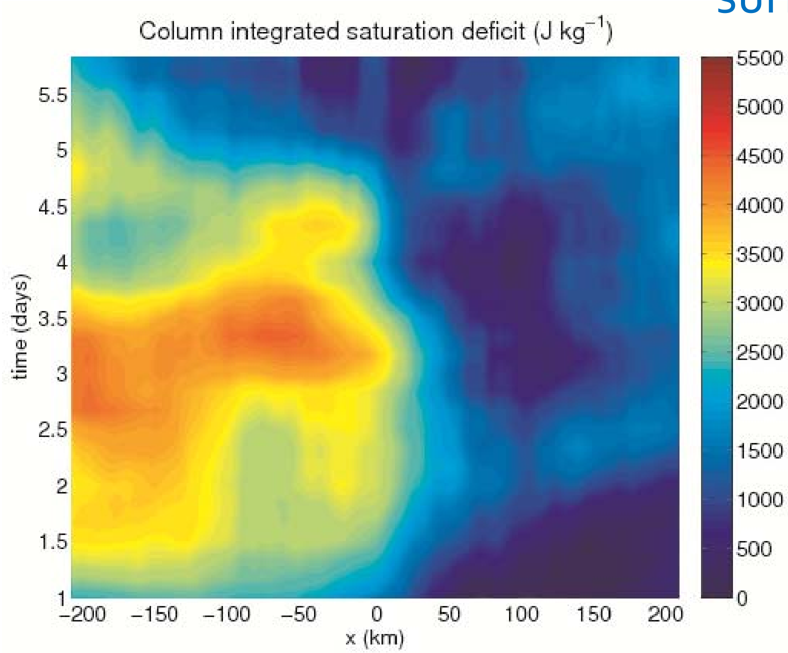
Rapid Intensification (mesoscale saturation)

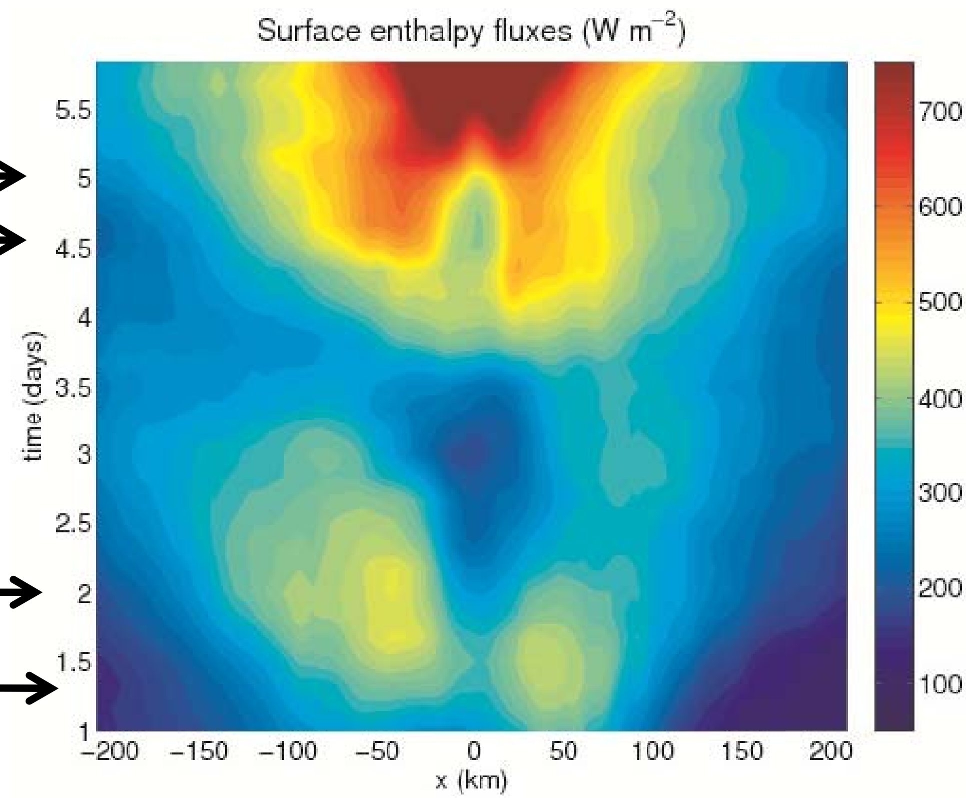
Genesis Proxy

Development and inward precession of new mid-level vortex

Propagation of the original mid-level vortex

Control: Meridionally averaged column integrated saturation deficit and surface fluxes





Rapid intensification →

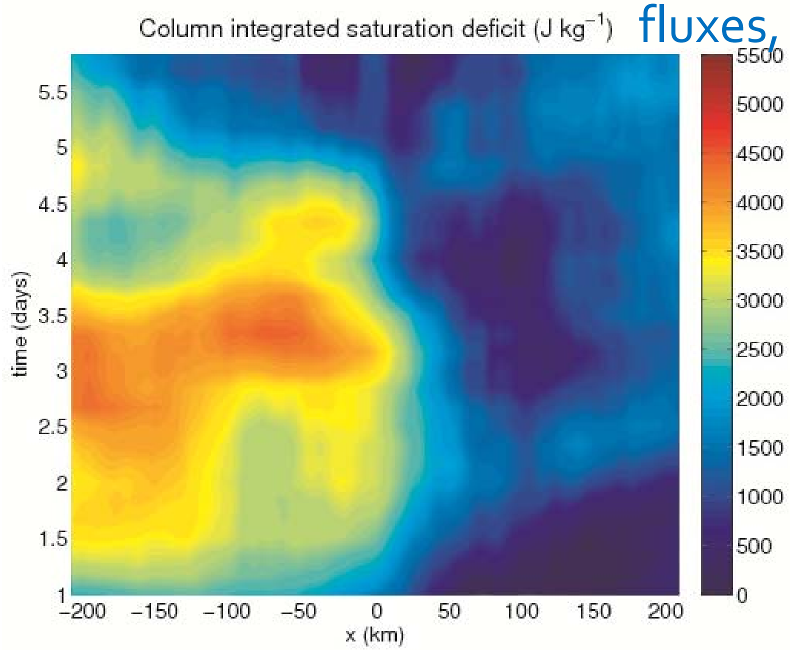
Genesis proxy →

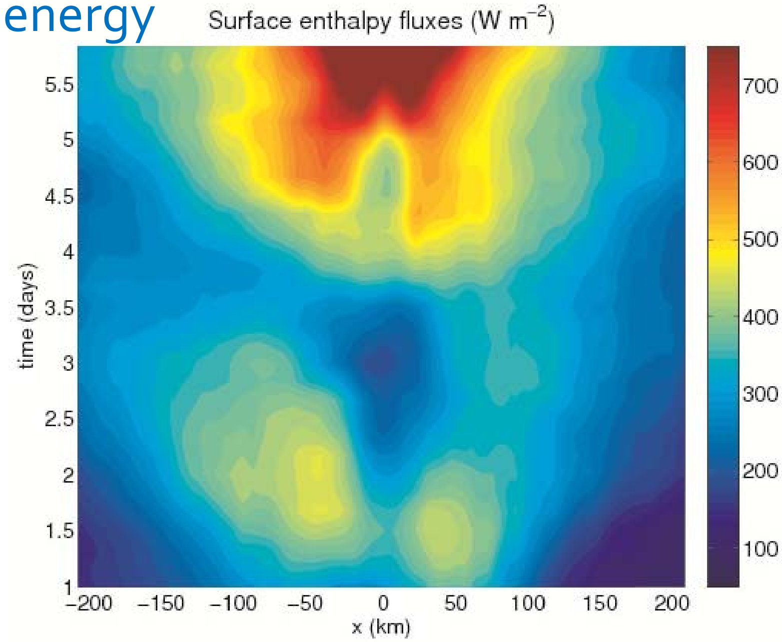
Mid-level vortex formation →

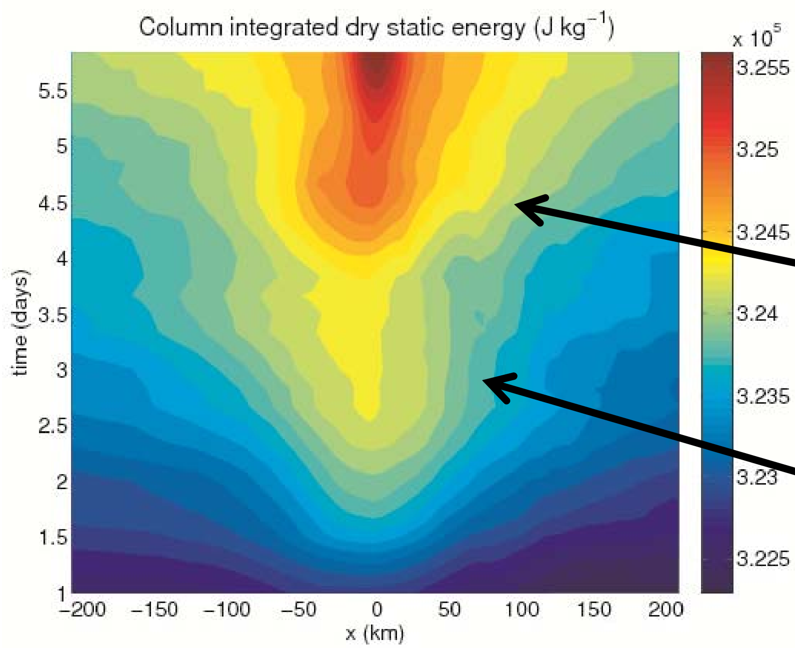
Surface fluxes due to thermodynamic disequilibrium →

Surface fluxes due to convective gustiness →

Control: Meridionally averaged column integrated saturation deficit, surface fluxes, and dry static energy





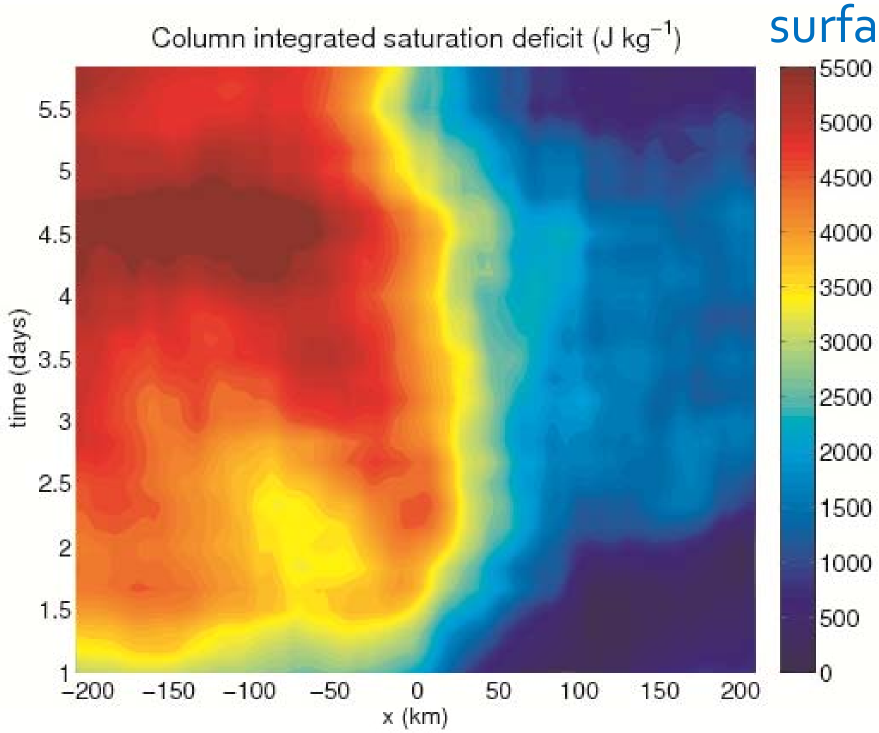


Column integrated dry static energy controlled by column integrated temperature

Post-saturation – Boundary layer warms by surface fluxes. Convection carries moist static energy aloft. Column warming results in hydrostatic pressure falls.

Pre-saturation – Diabatic heating offset by adiabatic cooling. Warming by surface fluxes offset by convective downdrafts.

WARM: Meridionally averaged column integrated saturation deficit and surface fluxes

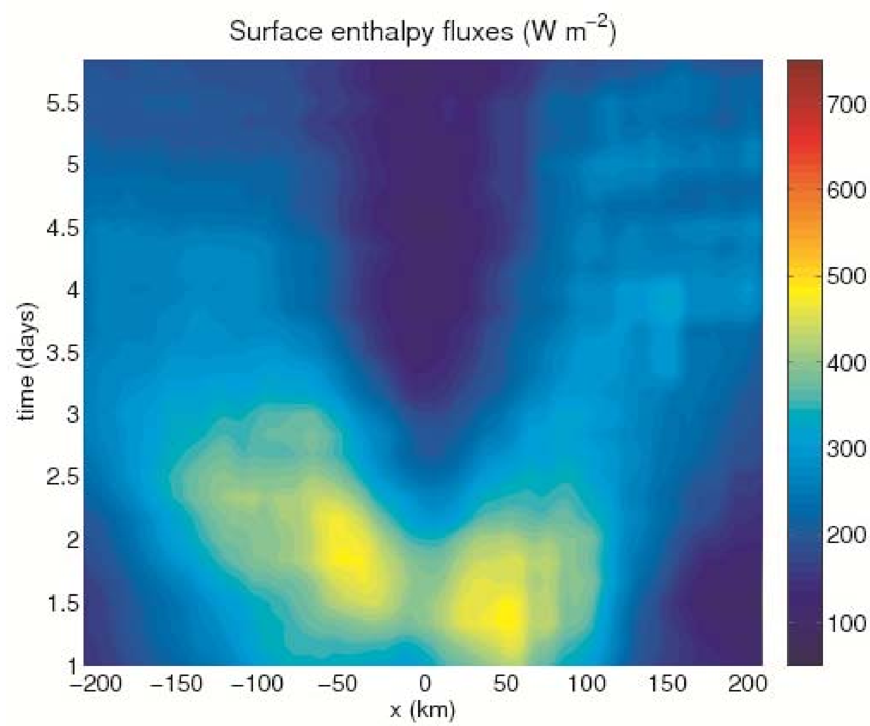


Very large column integrated saturation deficit in the core

Downdrafts!

No sustained localized convection due to convective downdrafts

No mid-level vortex!



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