Some problems related to sub-grid closures

Isaac Held
DCMIP, Boulder, 2016

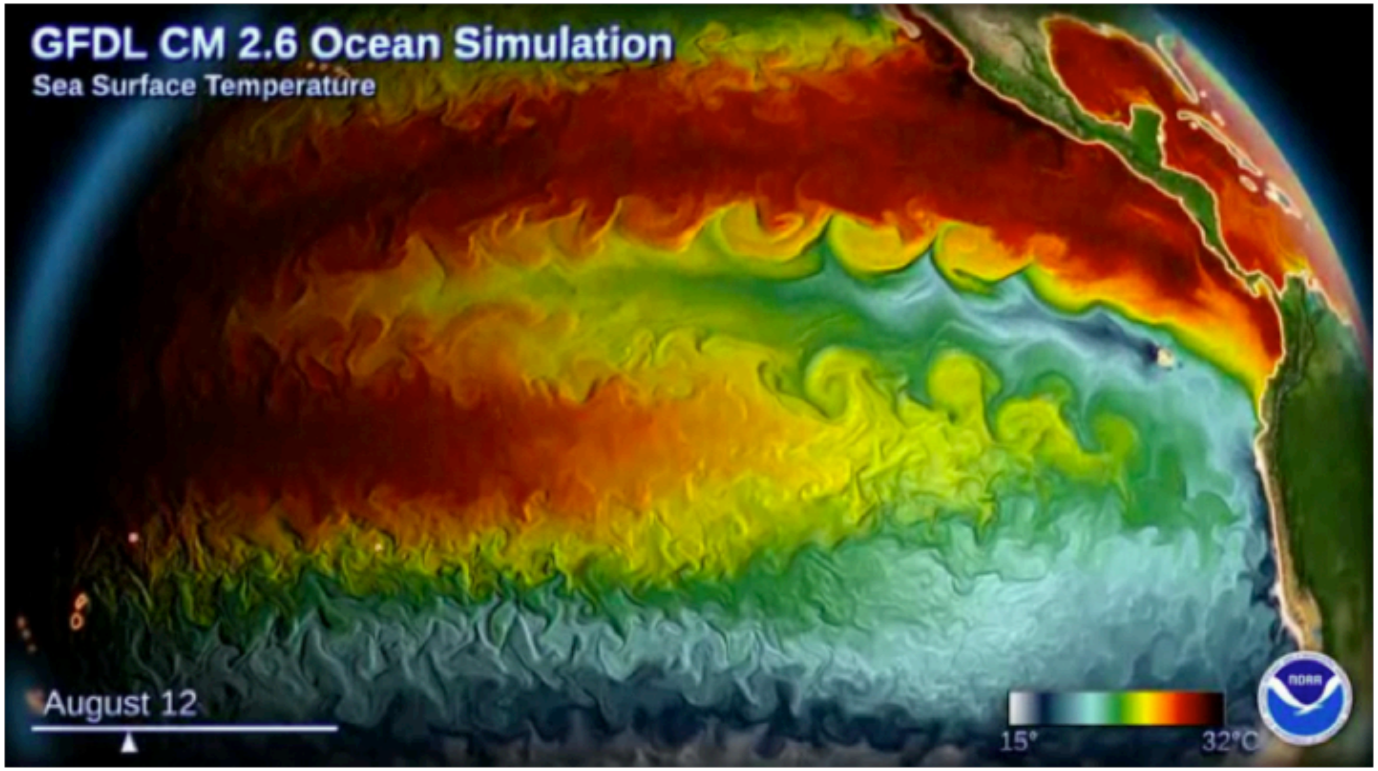
oceanic mesoscale eddies
(geostrophic turbulence)

deep convection
(moist convective turbulence)

2 types of sub-grid closures:

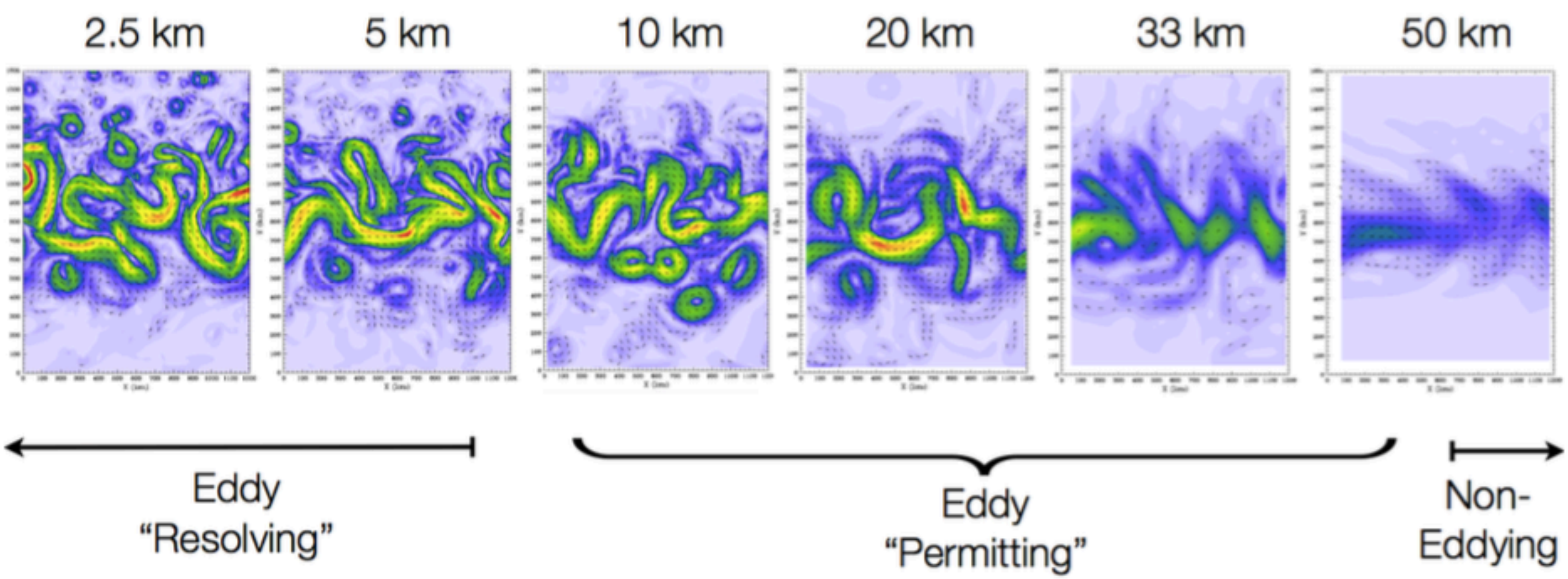
Scale separation
i.e. atmospheric/oceanic surface boundary layers

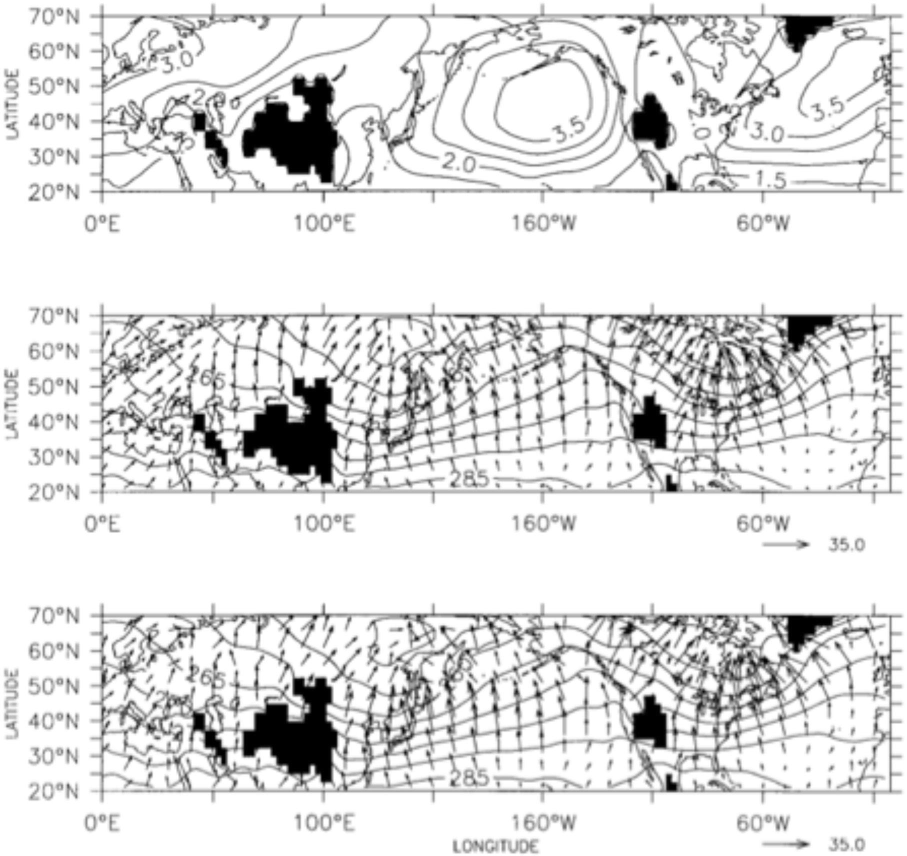
No scale separation
i.e. Smagorinsky viscosity closures when
grid scale is in middle of 3D direct cascades


GFDL CM 2.6 Ocean Simulation
Sea Surface Temperature
August 12
15°
32°C
NOAA

Series of idealized simulations of zonal jet in 1200 km x 1600 km zonally re-entrant domain (from Hallberg 2013)

Snapshots of upper ocean velocity at various resolutions ($k_d^{-1} \approx 30$km):





Diffusive approximation

Observed

Local diffusion with a diffusivity proportional to rms eddy streamfunction can reproduce observed (irrotational) eddy heat fluxes in lower troposphere – streamfunction has same units as diffusivity (L^2/T)

Kushner and Held, GRL 1998

$$\frac{\partial q_i}{\partial t} = -J(\psi_i, q_i) + Heat + Friction \ (i = 1, 2)$$

Conservation of QG PV

$$J(\psi, B) \equiv \frac{\partial \psi}{\partial x}\frac{\partial B}{\partial y} - \frac{\partial \psi}{\partial y}\frac{\partial B}{\partial x}$$

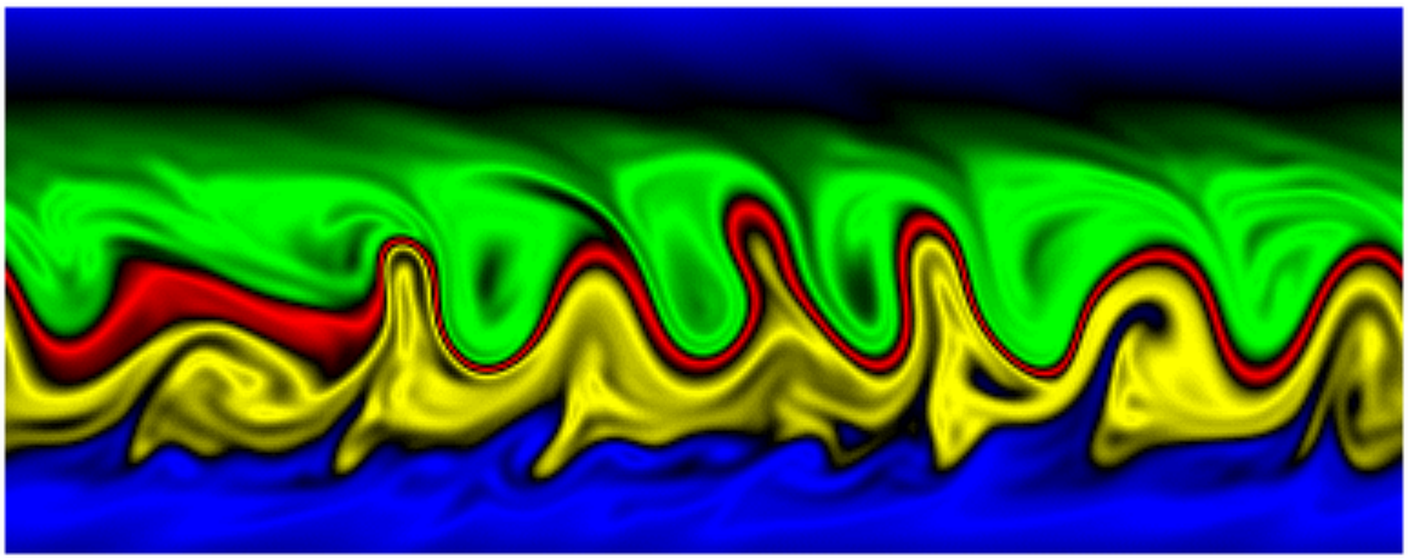
Geostrophic advection

$$q_1 = \beta y + \nabla^2 \psi_1 - \frac{\psi_1 - \psi_2}{\lambda^2}$$

PV in upper layer

$$q_2 = \beta y + \nabla^2 \psi_2 + \frac{\psi_1 - \psi_2}{\lambda^2}$$

PV in lower layer



Homogeneous theory for 2-layer QG model

$$\Psi_i = - U_i\, y + \psi$$

doubly periodic

External parameters:

β = Planetary vorticity gradient

λ = radius of deformation

U = imposed vertical shear (U_1 -U_2)

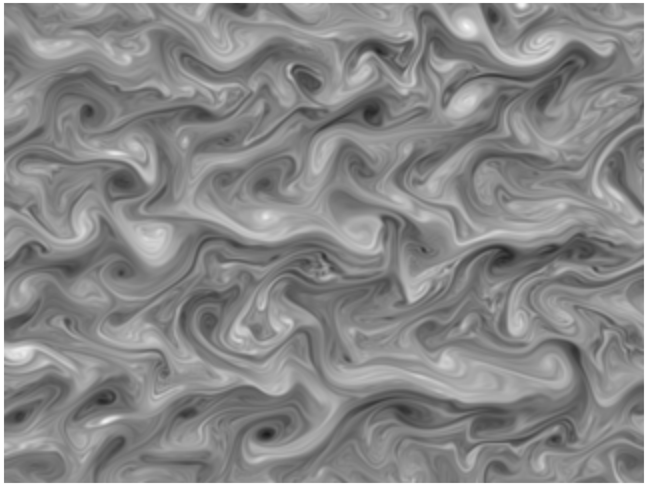
κ = surface frictional damping rate

Predicting

ε = rate of (inverse) cascade of energy =
rate of dissipation of kinetic energy

$\mathcal{D}$ => PV fluxes, heat fluxes

5



Would like a theory for

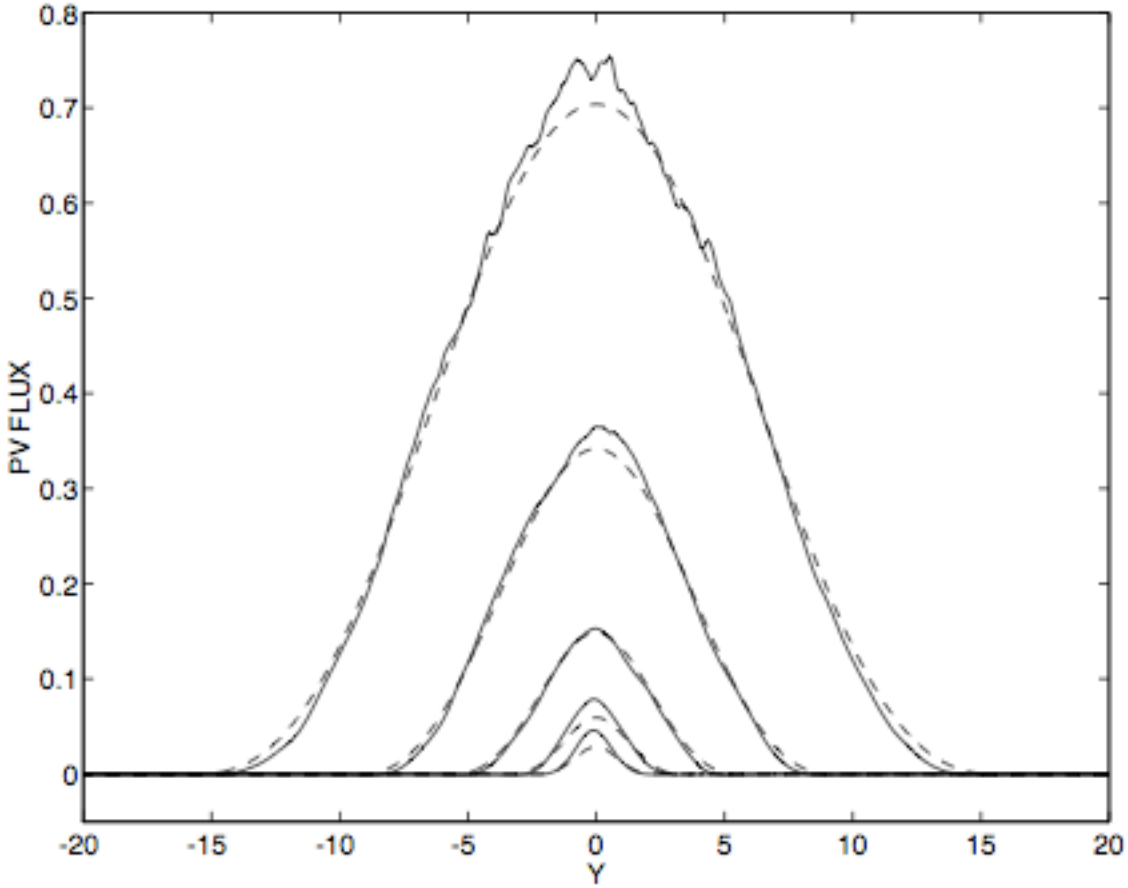
$$\mathcal{D}/(U\lambda) \sim F(\xi, \kappa_F, \kappa_T)$$

$$\xi = U/(\beta\lambda^2)$$

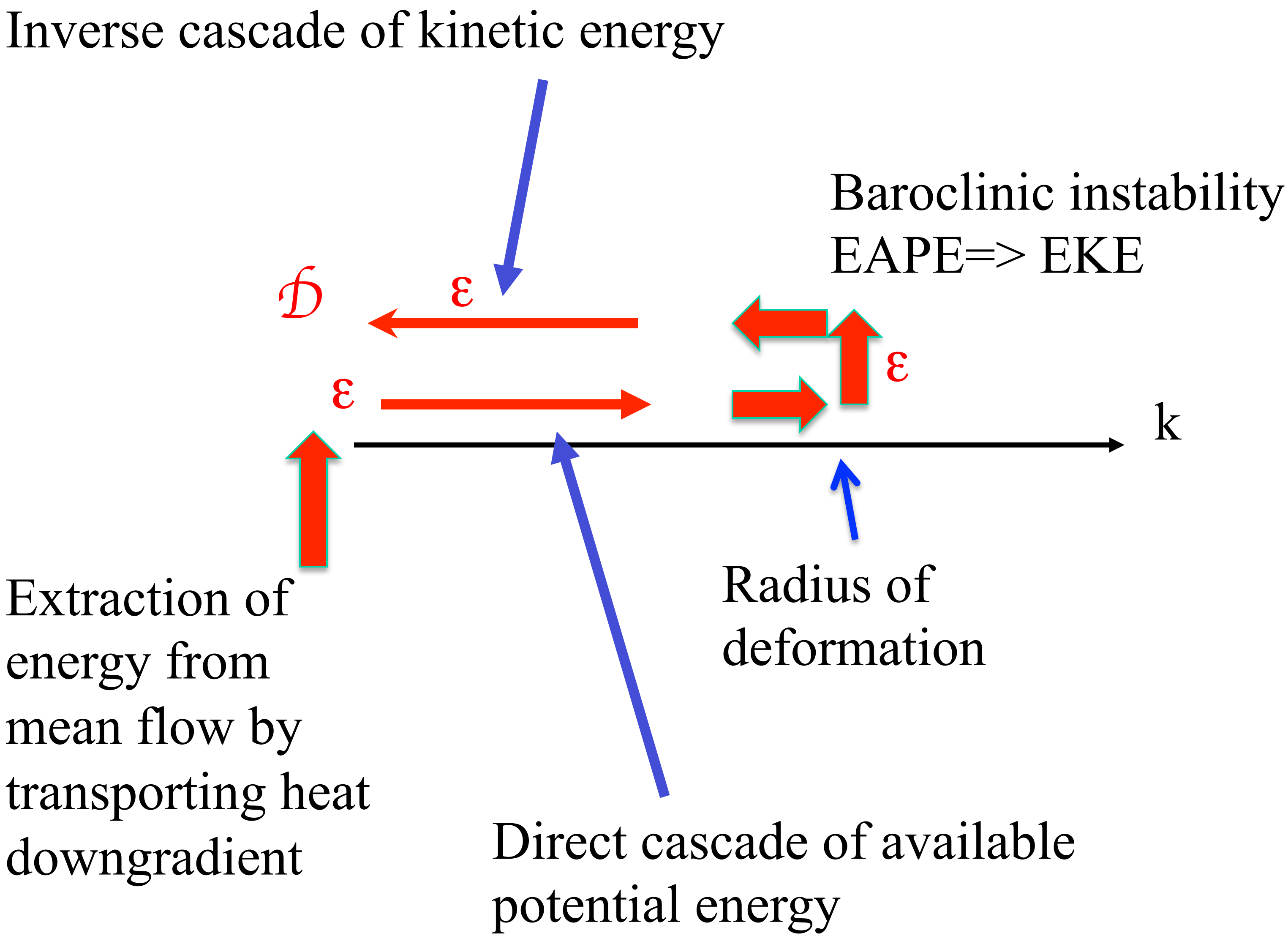
$\kappa_F = k_F\,(\lambda/U)$; damping of lower layer wind

In lieu of a complete theory, just use the double-periodic model to **measure** the model's diffusivity

Comparing numerical simulation for unstable jets of various widths with diffusive theory



Pavan and Held, JAS



Inverse cascade of kinetic energy
Baroclinic instability
EAPE=> EKE
𝒟
ε
ε
ε
k
Radius of
deformation
Extraction of
energy from
mean flow by
transporting heat
downgradient
Direct cascade of available
potential energy

Assume energy cascade stopped at the "Rhines scale" where flow becomes linear

How do you create something with units of Diffusivity (L^2T^{-1}) from ε (L^2T^{-3}) and β ($L^{-1}T^{-1}$)

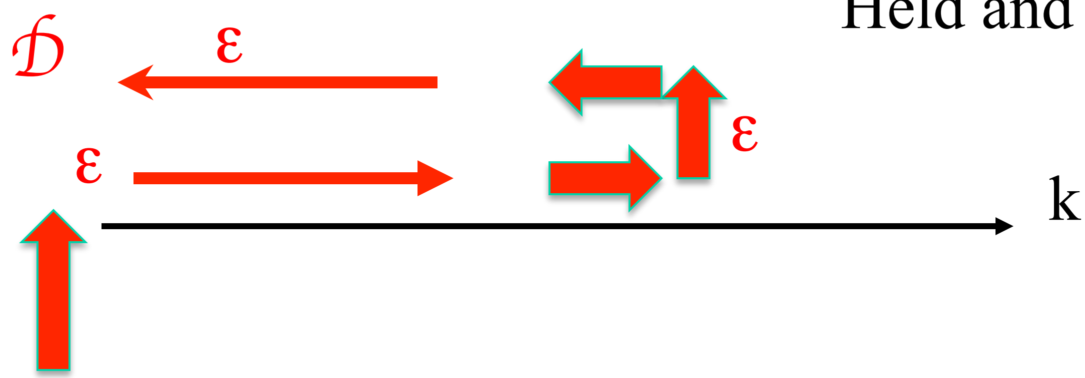
$$\mathcal{D} \sim \varepsilon^{3/5}\, \beta^{-4/5}$$

$$\varepsilon = \mathcal{D}\, U^2\, \lambda^{-2}$$

$$\Rightarrow \mathcal{D} \sim \beta^{-2}\, U^3 \lambda^{-3}$$

(Held and Larichev, JAS, 1995)

Held and Larichev, 1995



ε= production of EAPE ~ flux x gradient

$$\varepsilon \sim \text{diffusivity x (gradient)}^2$$

$$\varepsilon \sim \mathcal{D}\, U^2 \lambda^{-2} = \mathcal{D}\, T^{-2} \quad (T = \lambda/U)$$

Combined with $\mathcal{D} \sim \varepsilon^{3/5} \beta^{-4/5}$

$\Rightarrow \mathcal{D} \sim \beta^{-2}\, T^{-3}$ proportional to cube of temperature gradient

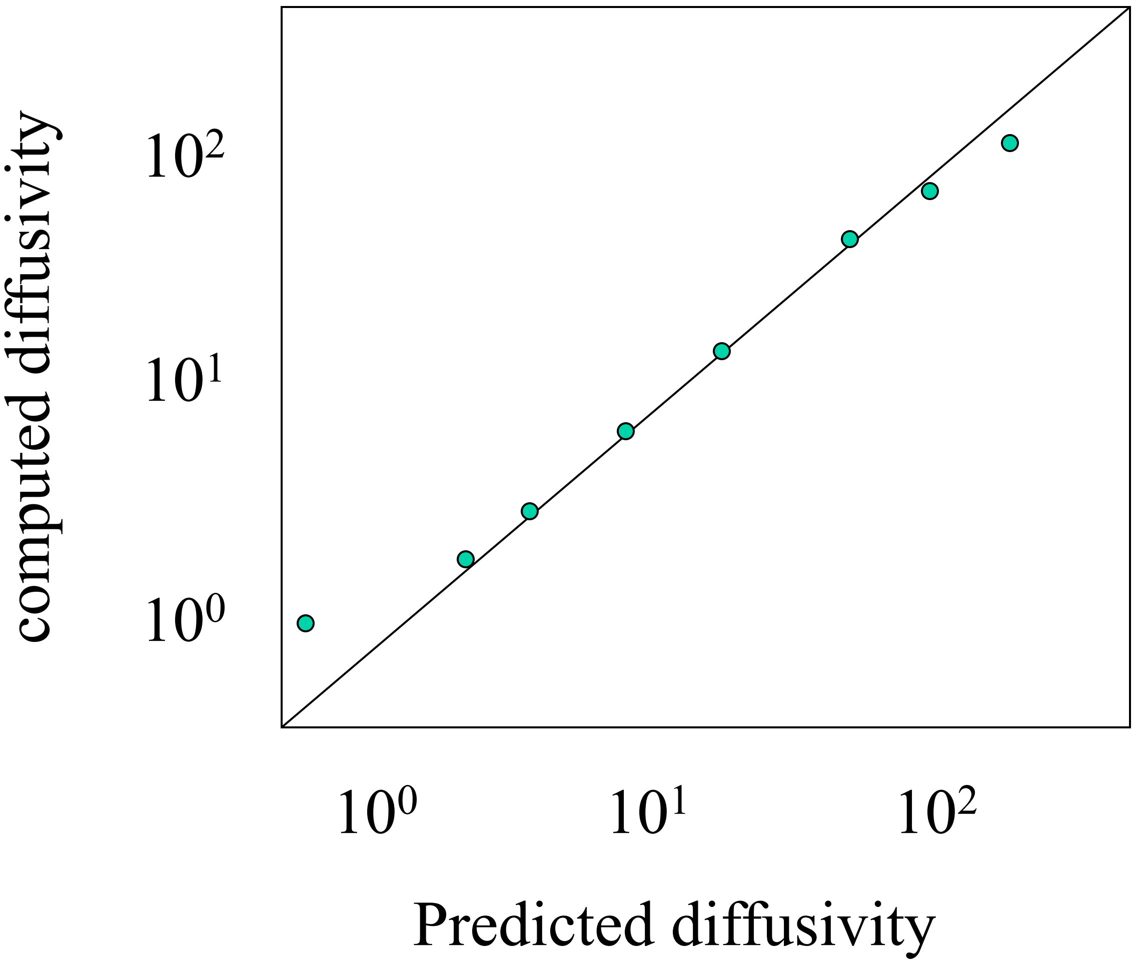
Lapeyre and Held (2003):

for arbitrary β, diffuse lower layer PV - why?

$$\mathcal{D}_1 = \varepsilon^{3/5}\,\beta^{-4/5} \text{ then implies}$$

$$\mathcal{D}_1 = (U\lambda)\,\xi^2\,(1 - \xi^{-1})^{3/2}$$

$$\xi = U/(\beta\lambda^2)$$



Lapeyre and Held, JAS, 2003

Would like a theory for

$$\mathcal{D}/(U\lambda) \sim F(\xi, \kappa_F, \kappa_T)$$

$\xi = U/(\beta\lambda^2)$

$\kappa_F = k_F\ (\lambda/U)$; damping of lower layer wind

$\kappa_T = k_T\ (\lambda/U)$; damping of temperatures

Dependence on damping is not understood

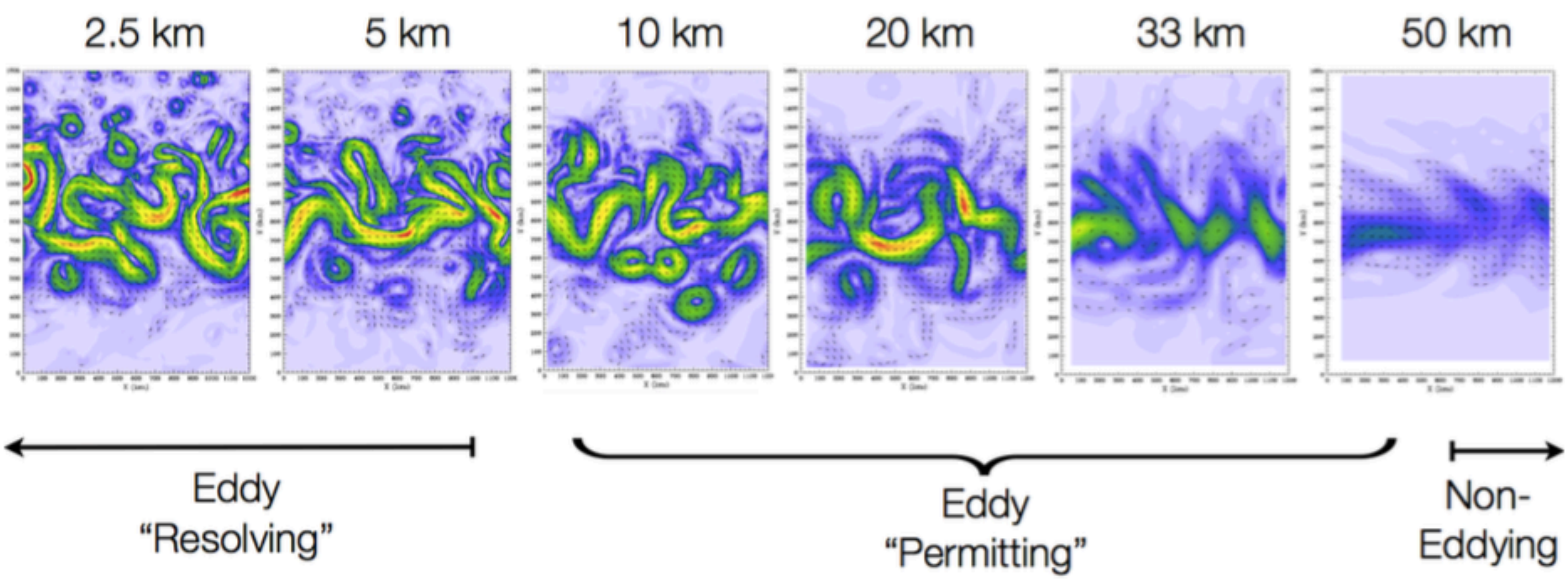
HL scaling suggests $\mathcal{D}$ independent of κ_F

and $\mathcal{D}$ => infinity as β => 0

(neither is correct, ie, Thompson+Young, JAS 2006

Series of idealized simulations of zonal jet in 1200 km x 1600 km zonally re-entrant domain (from Hallberg 2013)

Snapshots of upper ocean velocity at various resolutions ($k_d^{-1} \approx 30$km):

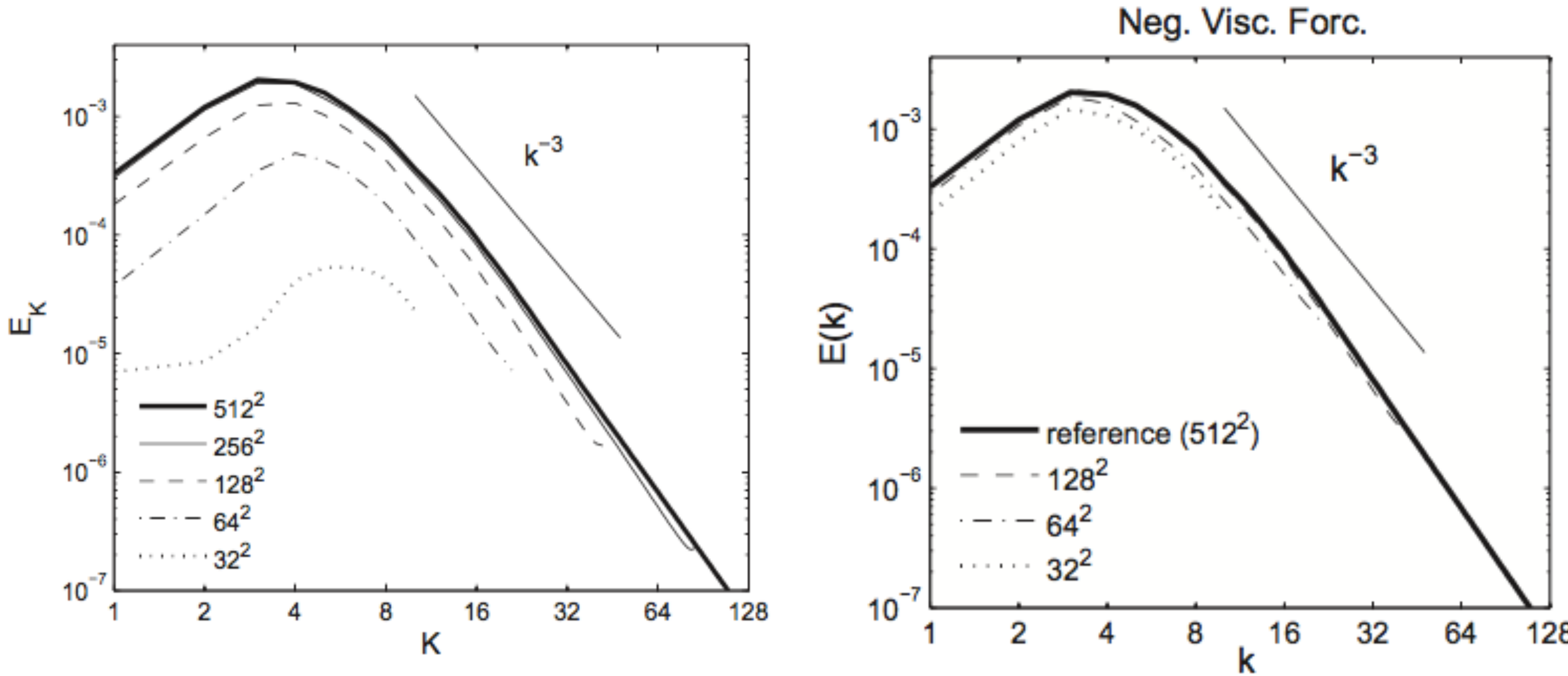


$$\frac{\partial \zeta}{\partial t} = \ldots\ldots\ldots + \nu \nabla^4 \zeta$$

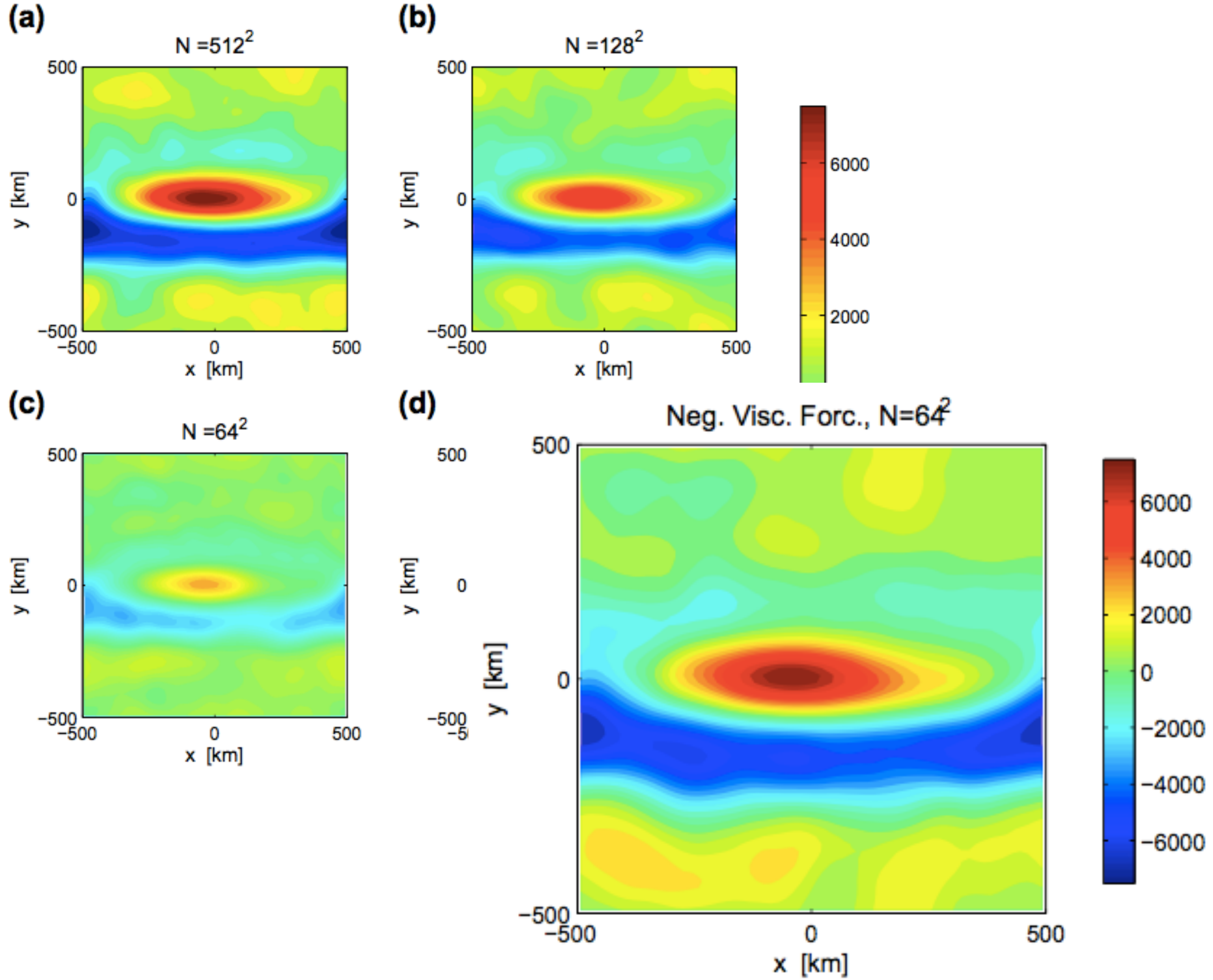
$$\frac{\partial \zeta}{\partial t} = \ldots\ldots\ldots + \nu \nabla^4 \zeta - \mu \nabla^2 \zeta$$

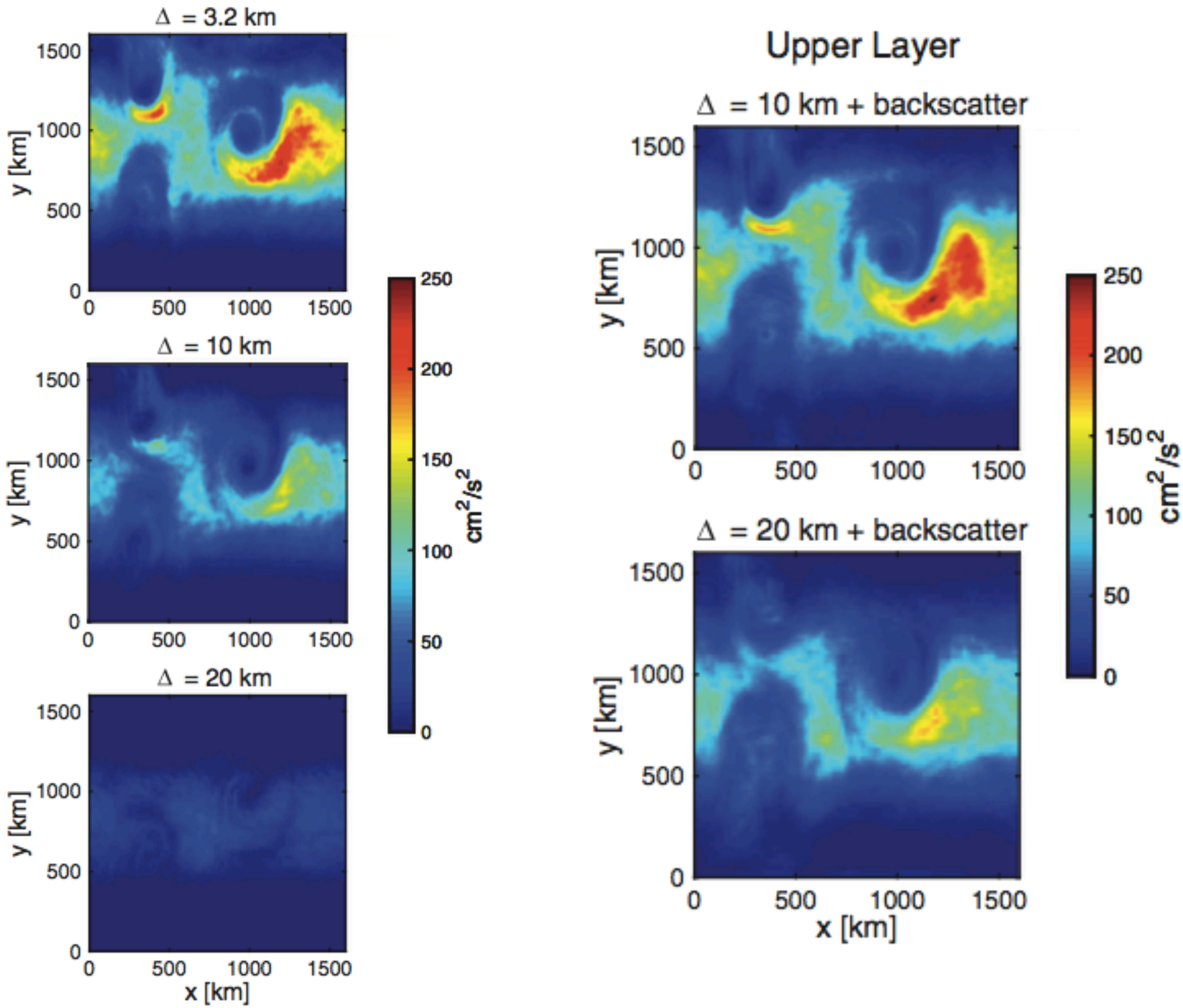
$$\frac{\partial e}{\partial t} = S_{diss} - S_{back} + \ldots\ldots \quad ; \quad \mu \propto \sqrt{e}$$

Jansen and Held, Ocean Modeling, 2012
Jansen et al , Ocean Modeling, 2014
Jansen et al, Ocean Modeling, 2015
Zurita-Gator et al, JAS, 2015



E_K
K
k^-3
512^2
256^2
128^2
64^2
32^2
Neg. Visc. Forc.
E(k)
k
k^-3
reference (512^2)
128^2
64^2
32^2


(a)
N =512^2
(b)
N =128^2
(c)
N =64^2
(d)
Neg. Visc. Forc., N=64^2
y [km]
x [km]
500
0
−500
6000
4000
2000
0
−2000
−4000
−6000



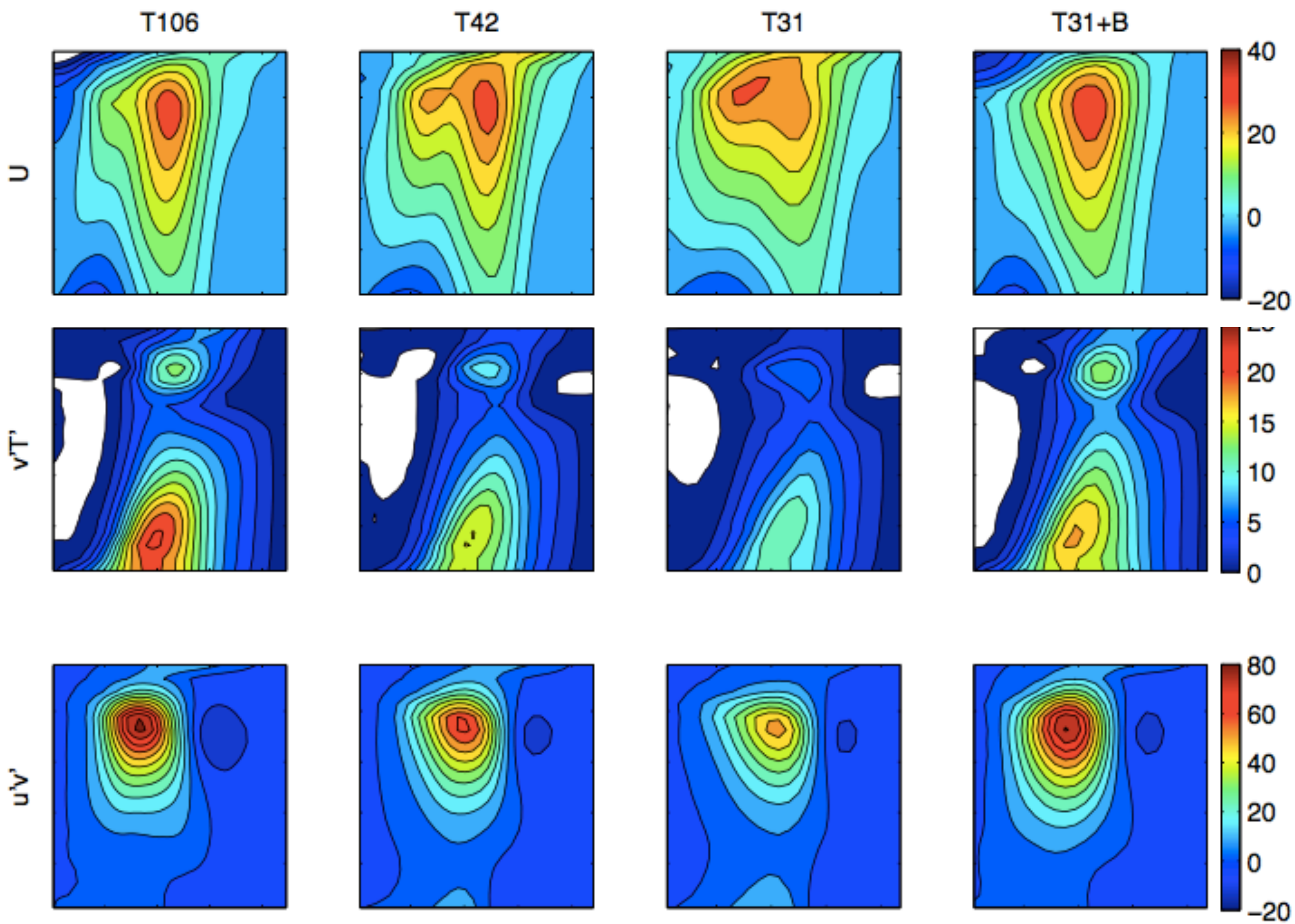
Upper Layer
Δ = 3.2 km
Δ = 10 km
Δ = 20 km
Δ = 10 km + backscatter
Δ = 20 km + backscatter
y [km]
x [km]
cm²/s²
0
500
1000
1500
50
100
150
200
250

Spectral dry dynamical core

U

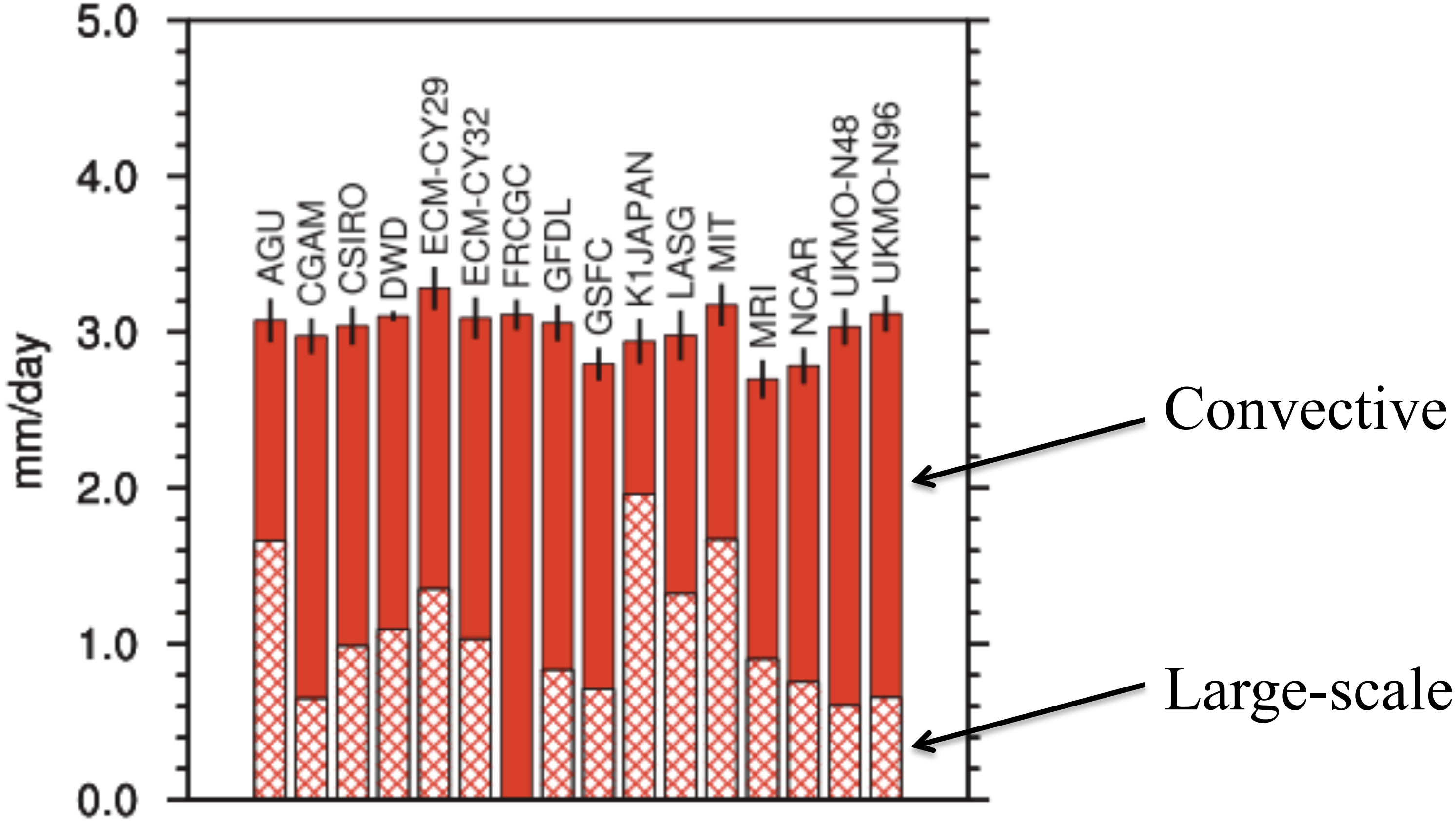
v’T’

u’v’



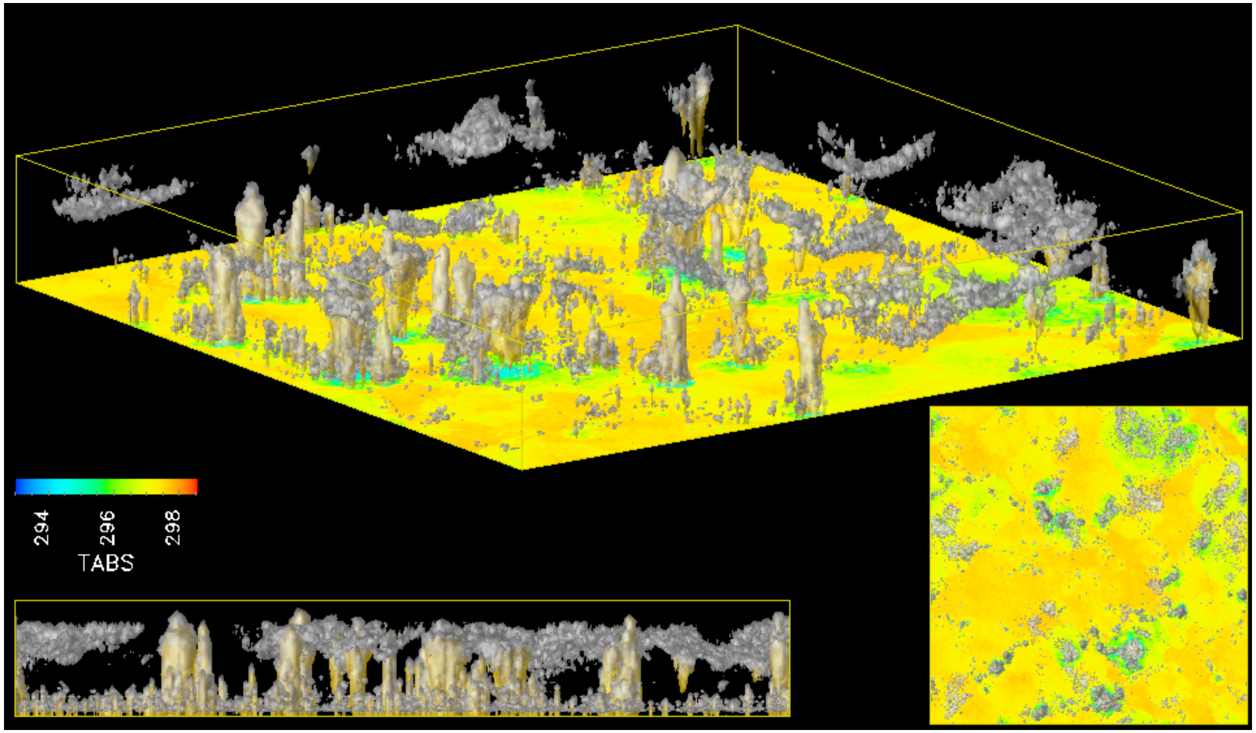
Zurita-Gator et al, JAS 2015

“convective” vs “large-scale” tropical precipitation – (Aquaplanet Model Intercomparison Project)



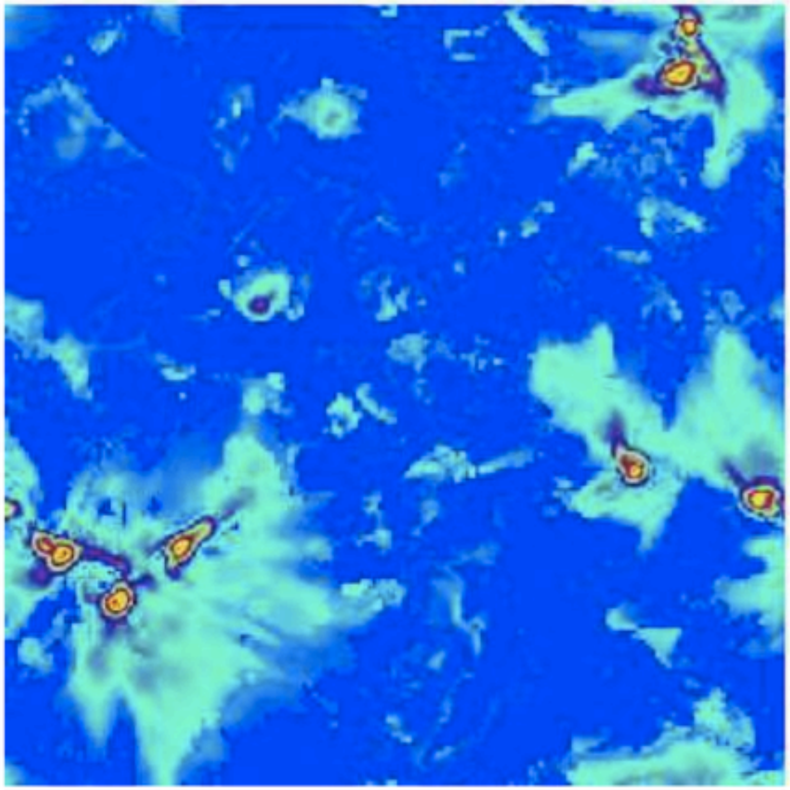
Moist convective closure:

One idea: start with cloud-resolving
(non-rotating) radiative-convective equilibrium
Compare with single column model from a GCM



Courtesy of P. Blossey, C. Bretherton

Non-rotating RCE with standard 25km HiRAM



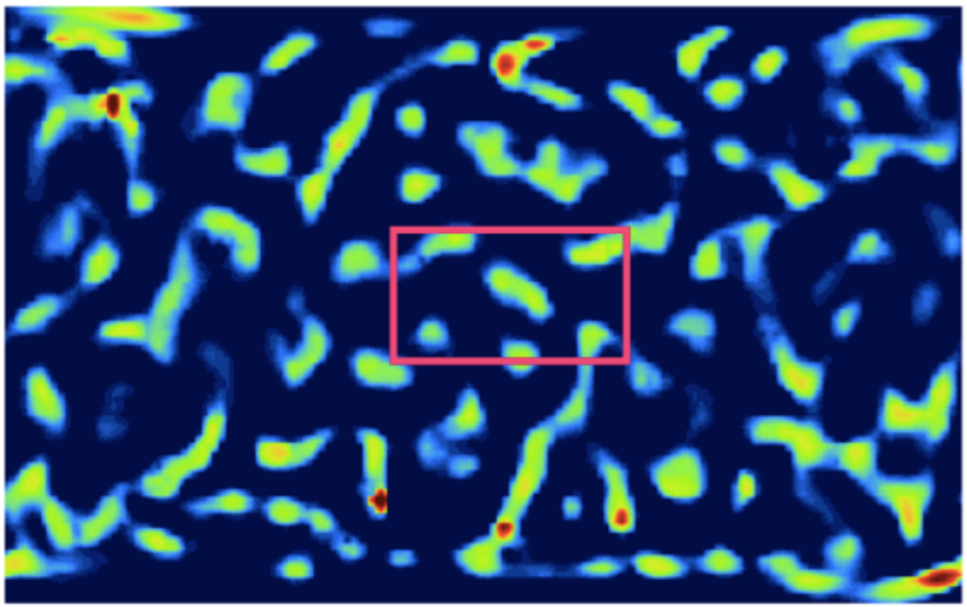
2500km x 2500km domain

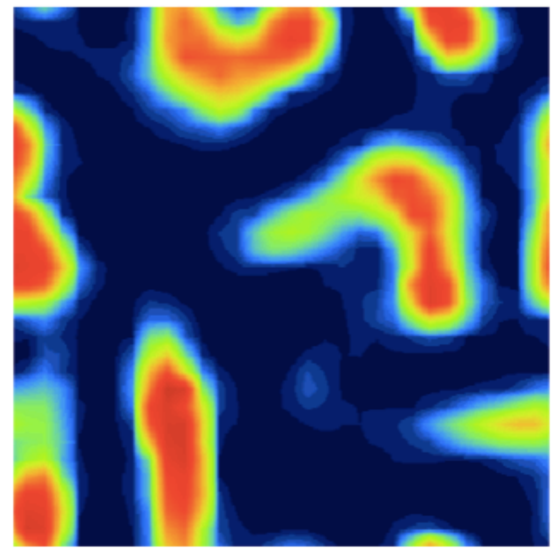
Instantaneous precipitation

Horizontally uniform state unstable

Doubly-periodic: 200km grid, 32 x 32

global





Fixed sst

Held, Zhao, Wyman, 2007 – RCE in a GCM (AM2.1)

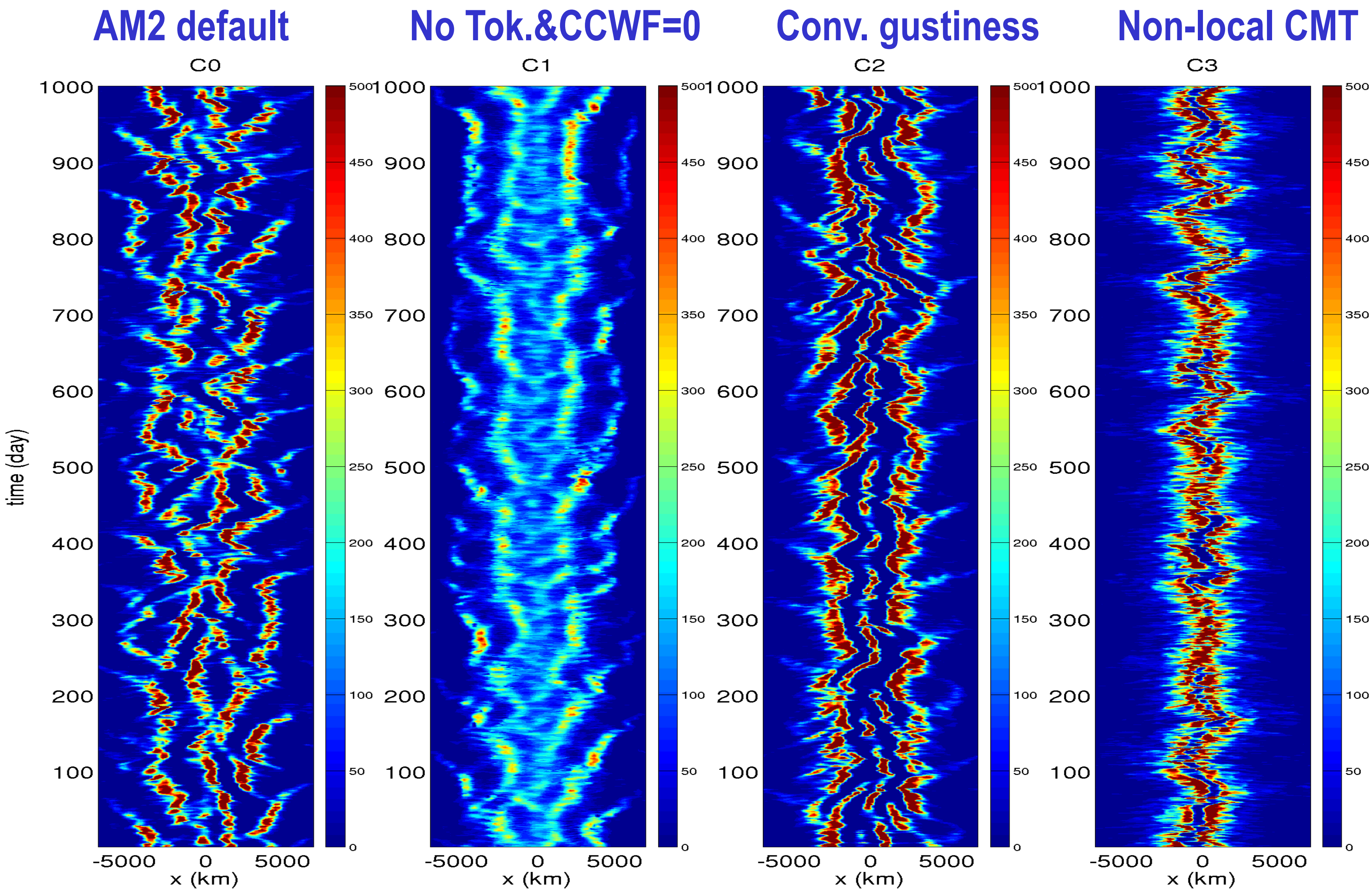
Possible way forward:

Minimize convective aggregation by removing radiative-cloud interactions, surface flux dependence on surface wind

If aggregation in CRM can be controlled, should GCM’s RCE (making the same simplifications) be stable??
If not, compare GCM’s RCE with CRM RCE

Z. Kuang proposal: perturb RCE with horizontally uniform steady heat and moisture sources with all possible vertical structures and compare GCM and CRM response matrices

Then move away from homogeneity by looking at Mock (non-rotating) Walker cell: SST(x) = A + B sin(x)



2-step process:
1) Optimize CRM against field programs

2) Test GCM closure against CRM;

allows a more systematic analysis
(developing closure becomes a theoretical problem)

Complimentary to
comparing GCM directly with observations