Turbulence and cloud formation in the atmospheric boundary layer

Some aspects of stratified turbulence

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Morphology and dynamics of anisotropic flows

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  - Center for Advanced Computing Research at Caltech

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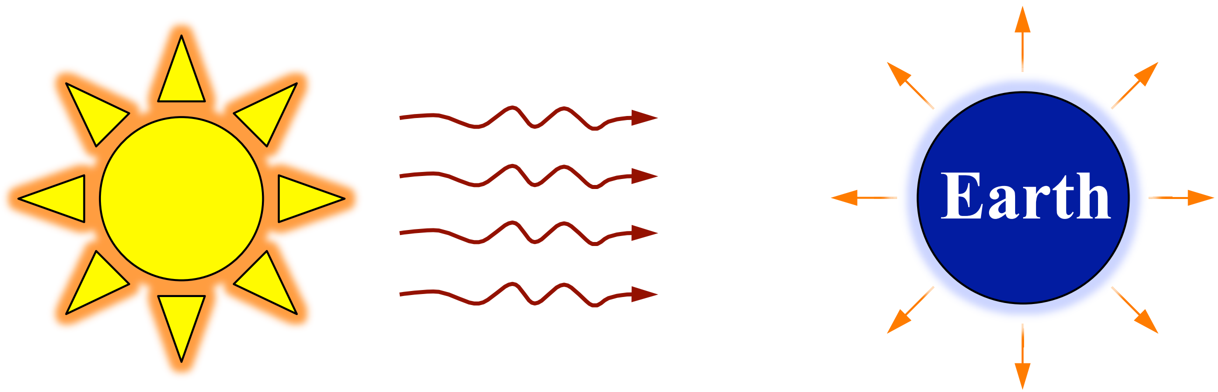
  - Office of Naval Research
  - NASA MAP Program

Outline

- Large scale $\longrightarrow$ small scale with main focus on the atmosphere
  - Fundamentals of climate dynamics
    (physical processes: radiation + rotation + multiple phases + stratification)
    - Planetary radiation balance
    - Climate forcing and feedbacks
    - Global circulation
  - The role of the atmospheric boundary layer
    (liquid & gas phase + stratification)
  - Stably stratified turbulence
    (stratification in a Boussinesq fluid + homogeneity assumption)
    - Phenomenology
    - Scaling relations
    - Small-scale anisotropy statistics in homogeneous stratified turbulence
- Stably stratified turbulence is a very active area of research
  - Some concepts are still not widely accepted
  - Use some of our own (unpublished) work to better illustrate some aspects of the theory

Mean equilibrium temperature

- Radiative equilibrium: Energy input = energy output



$$f\pi R^2(1-a) = 4\pi R^2\sigma T^4$$

Energy from sun ≈ 1400 W m^{-2}

Stefan–Boltzmann constant ≈ 5.67×10^{-8} W m^{-2} K^{-4}

Earth's albedo ≈ 0.31

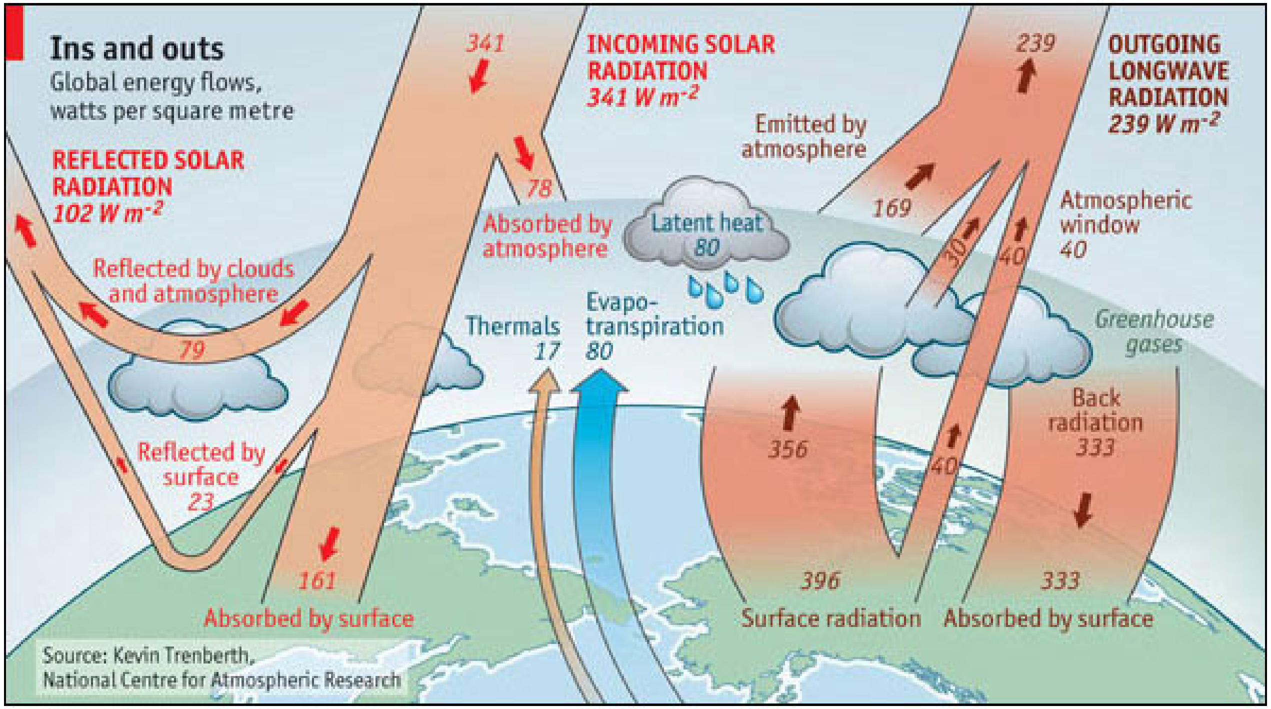
Earth's radius

$$T = \left(\frac{(1-a)f}{4\sigma}\right)^{1/4} = 255\,\mathrm{K} = -18\,^{\circ}\mathrm{C}$$

Too cold!!!

The "greenhouse" effect

- The atmosphere is a processor of energy
- Water droplets are very efficient absorbers and emitters of radiation
- Currently, there is an imbalance of ~ 0.9 W m^{-2}
  - Solar radiation on a clear summer day around noon at Cargèse is about 1000 W m^{-2}

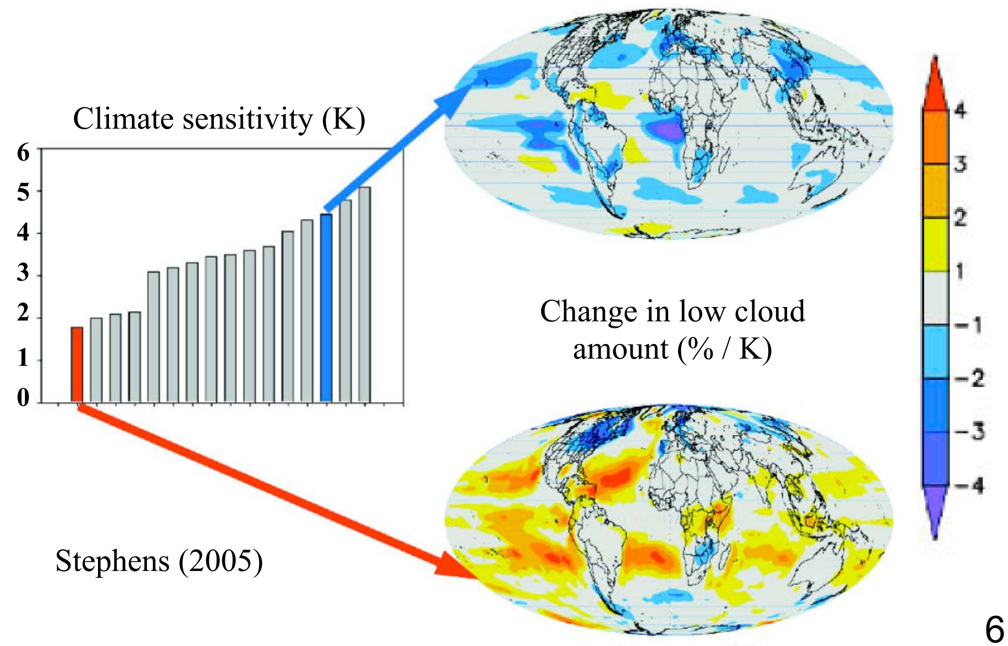


Graphic: The Economist, "The clouds of unknowing," March 18th, 2010
Data/analysis: Trenberth, Fasullo & Kiehl (2009)

Climate forcing and feedbacks

- Increase in greenhouse gases because of human activity has let to an increase in radiative forcing
  - CO_2 +40% and CH_4 +148% since pre-industrial levels (1750)
- Water vapor is the most dominant greenhouse gas
  - ...but we cannot control the amount of water in the atmosphere (thermodynamics, circulation, ocean, etc) → **feedback**
- Clouds have both warming and cooling effects depending on their characteristics (cloud thickness, cloud top height, etc…)
- Boundary layer clouds have a strong cooling effect
- Intergovermental Panel on Climate Change (IPCC) 2007:

  "Cloud feedbacks (particularly from low clouds) remain the largest source of uncertainty [in climate models]."



Stephens (2005)

Transport of mass and energy – Large-scale circulation

- Earth is a sphere (to a very good approximation)
  - Non-uniform heating
    - Equator receives more energy than the poles

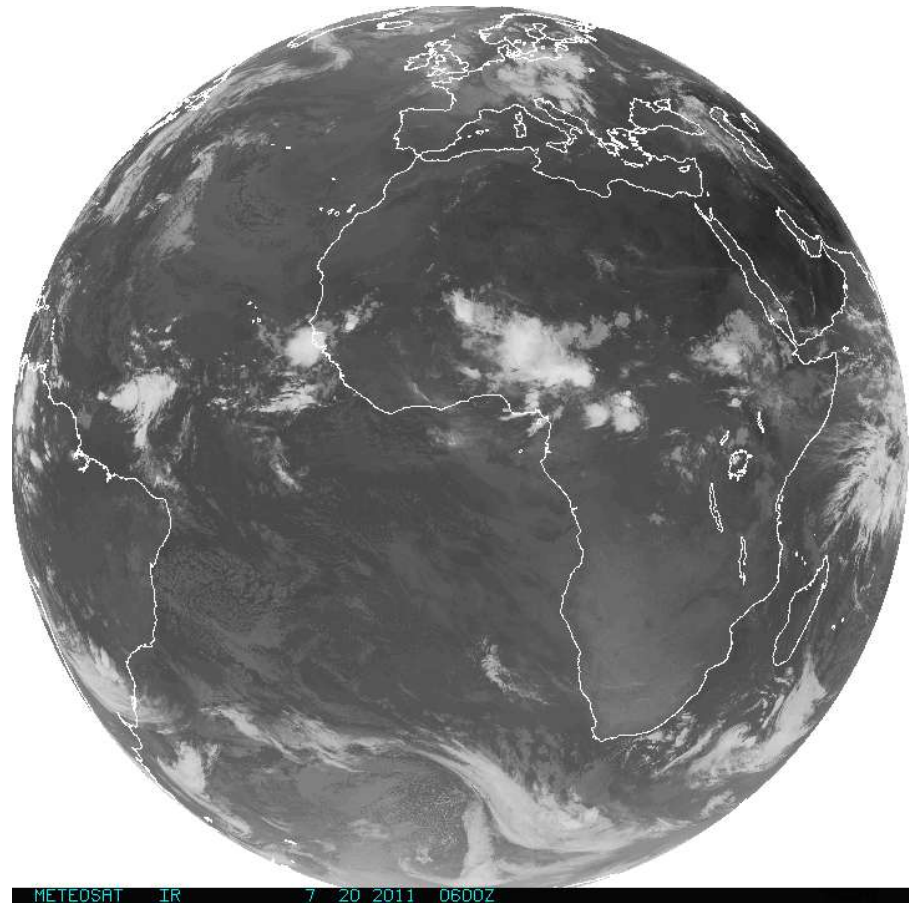
- Earth rotates
  - Angular momentum is conserved for zonal (east–west) motions, but…
  - Adjustment is necessary for meridional (north–south) motions – Coriolis force

Rossby number: $Ro = \dfrac{U}{fL}$

Coriollis parameter: $f = 2\Omega \sin\phi$

latitude

rotation rate



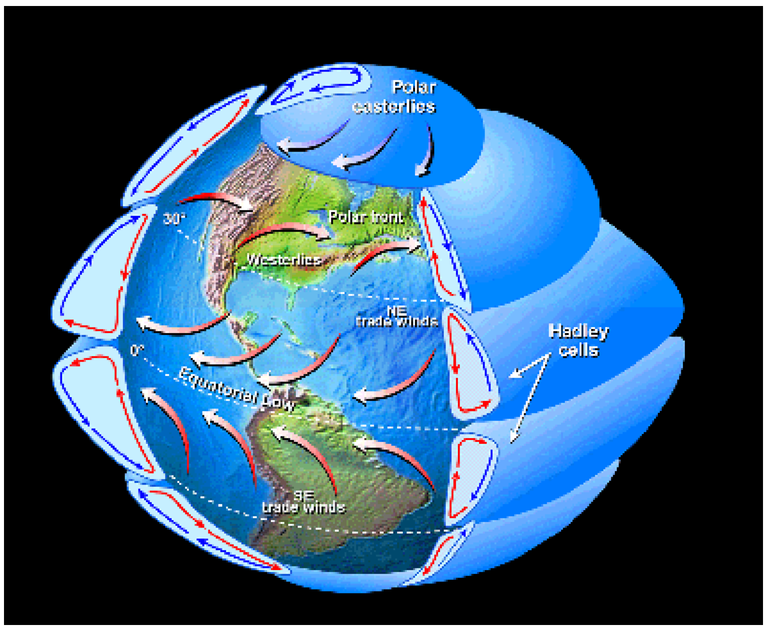
Infrared – 4 hours ago

METEOSAT/European Organisation for the Exploitation of Meteorological Satellites (EUMETSAT)

Large-scale circulation

Theory

Mean atmospheric circulation



www.larc.nasa.gov

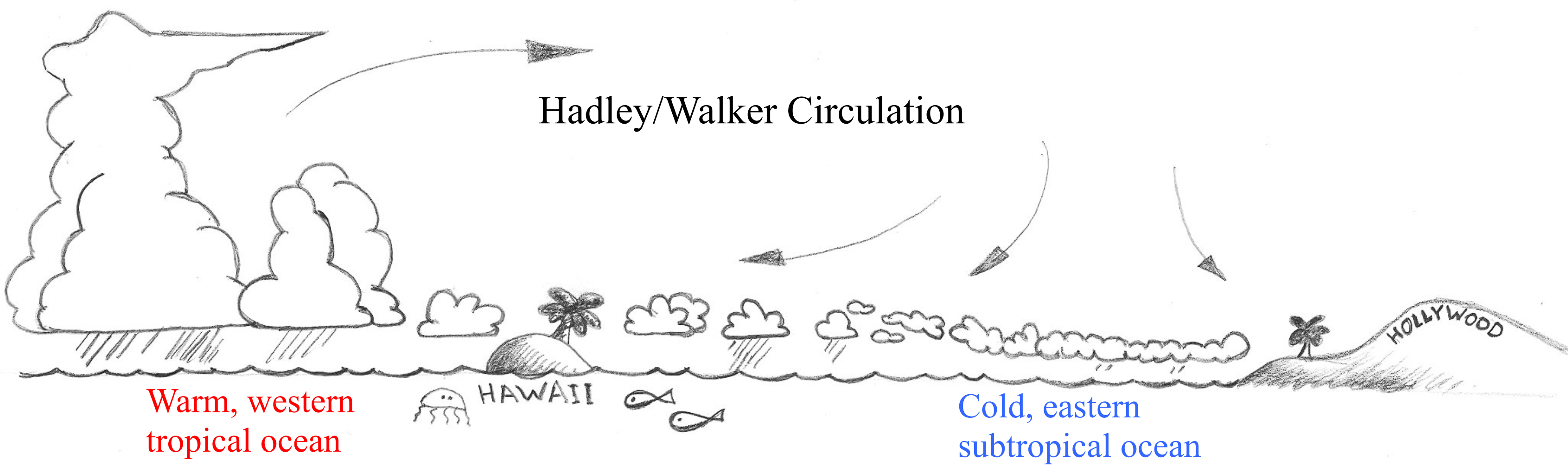
Application

Christopher Columbus' voyage to the New World in 1492

www.columbusnavigation.com
maps.google.com

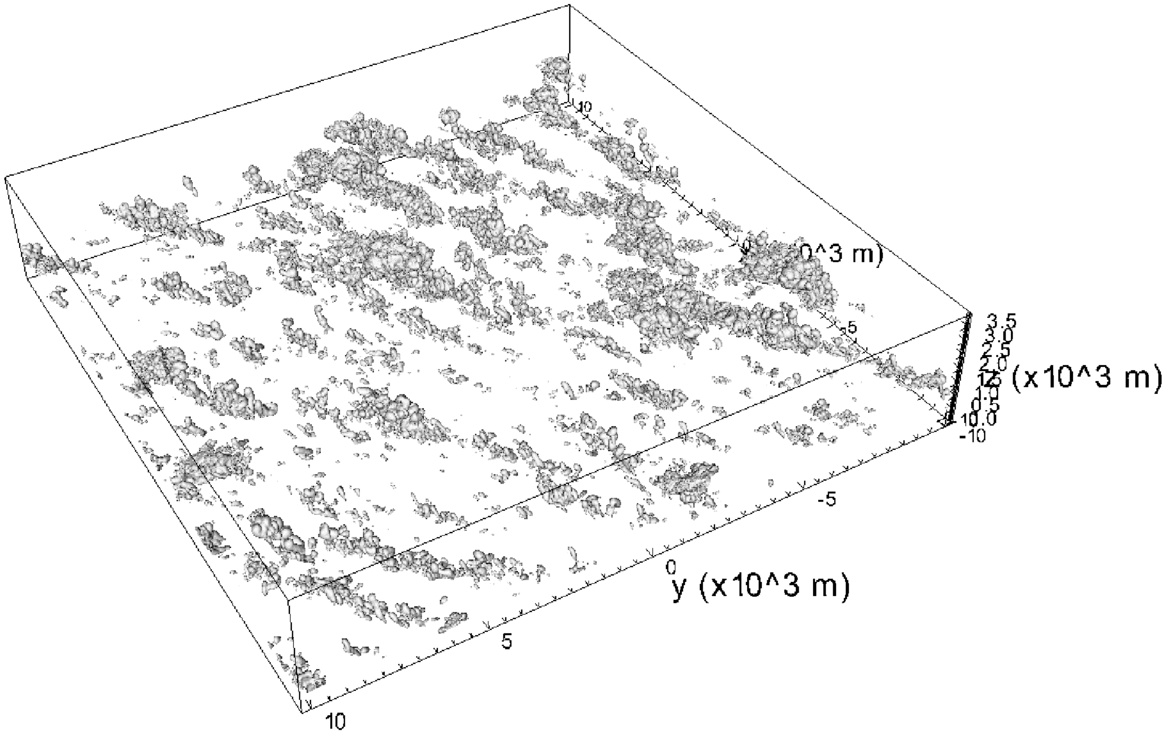
Boundary layer clouds and the large-scale circulation

- Clouds form when air cools below the saturation temperature and water vapor condenses to liquid droplets or ice crystals
  - Latent heat is released when condensation/deposition occurs: air becomes more buoyant
  - Condensation occurs on solid particles, **condensation nuclei**, otherwise RH > 100%
- Cloud droplets scatter visible light (geometric scattering), clouds appear white
  - Larger droplets are less effective scatterers
  - The ‘white stuff’ is not a passive tracer!



Adapted from Arakawa (1975) & Stevens (2005)

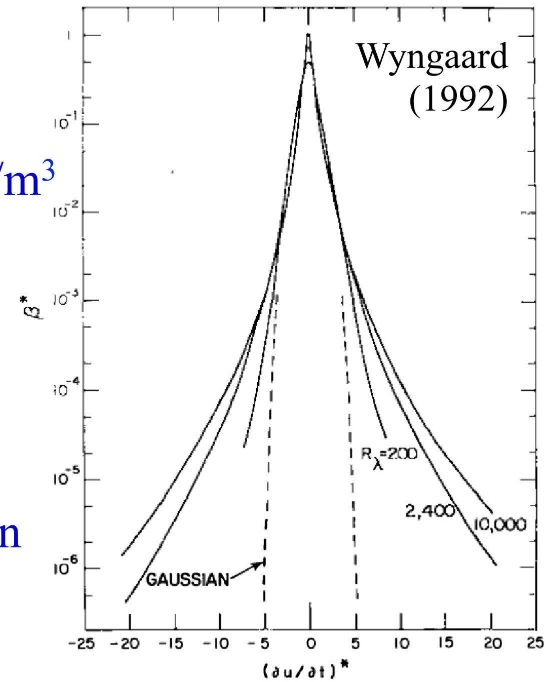
Cumulus clouds – Macrophysics

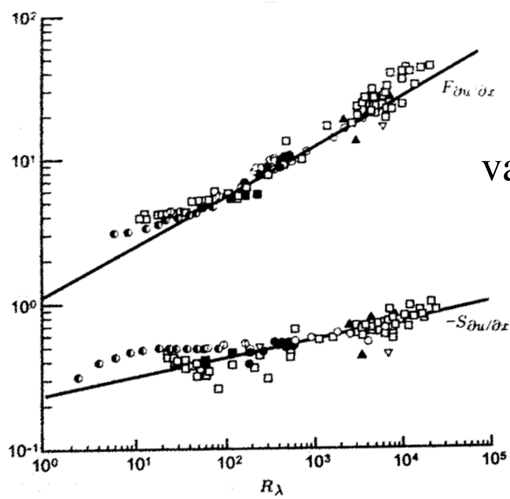


Matheou, Chung, Nuijens, Stevens & Teixeira, *Mon. Wea. Rev.* (2011)

Cloud microphysics

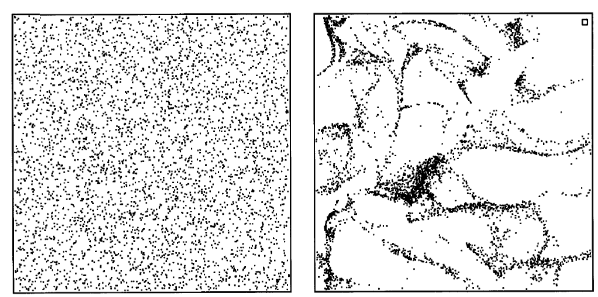
- Reynolds number in a cumulus cloud, $Re \approx 10^8$
- Kolmogorov scale $\eta \approx 1$ mm
- Cloud droplet diameter $d \approx 10$ µm and $\sim 100 \times 10^6$ drops/m^3
  - Droplet size distribution evolves, droplets grow larger with time, mainly through collisions and coalescence
  - Larger droplets:
    - Decrease in albedo: more light absorption and less scattering
    - Large sedimentation/fall speeds => formation of rain
- The small-scale character of turbulence plays a key role in the collision-coalescence process





Sreenivasan & Antonia (1997)

variance of droplet velocity divergence $\sim \tau_d^2 F_{\partial u/\partial x} \left(\dfrac{\varepsilon}{\nu}\right)^2$



Shaw et al. (1998)
Shaw (2003)
Wyngaard (2010)

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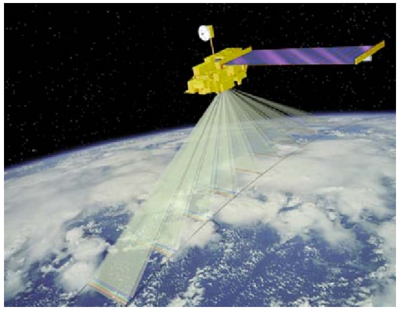
Interaction between cloud macro- and microphysics
Rain and cloud organization

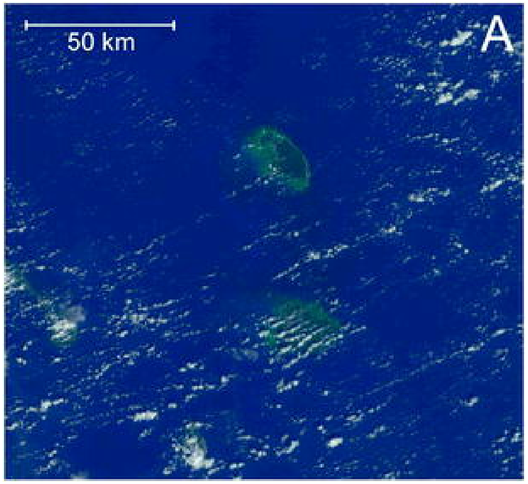
- Multi-angle Imaging SpectroRadiometer (MISR) images
- Observations during Rain In Cumulus over Ocean (RICO) campaign (Rauber et al. 2007)



Rain increases

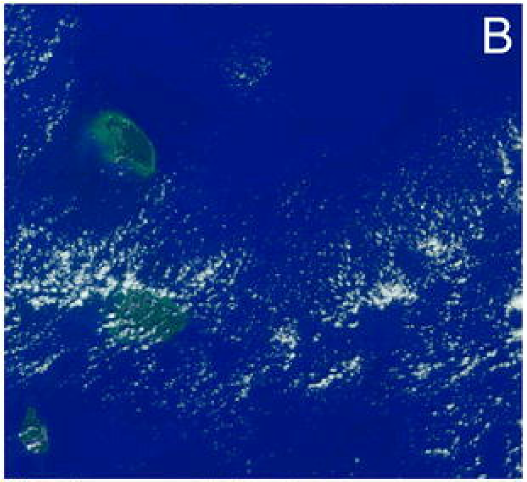
Wind-parallel cloud streets

Rarely produce precipitation



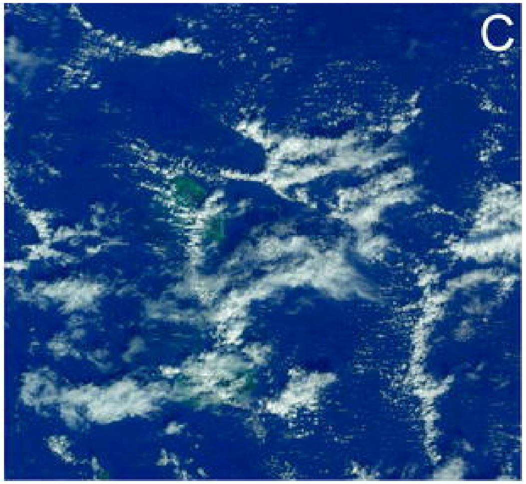
Cloud clusters

Light to moderate precipitation



Cloud clusters aligned in arc-shaped formations

Significant precipitation



Wind direction

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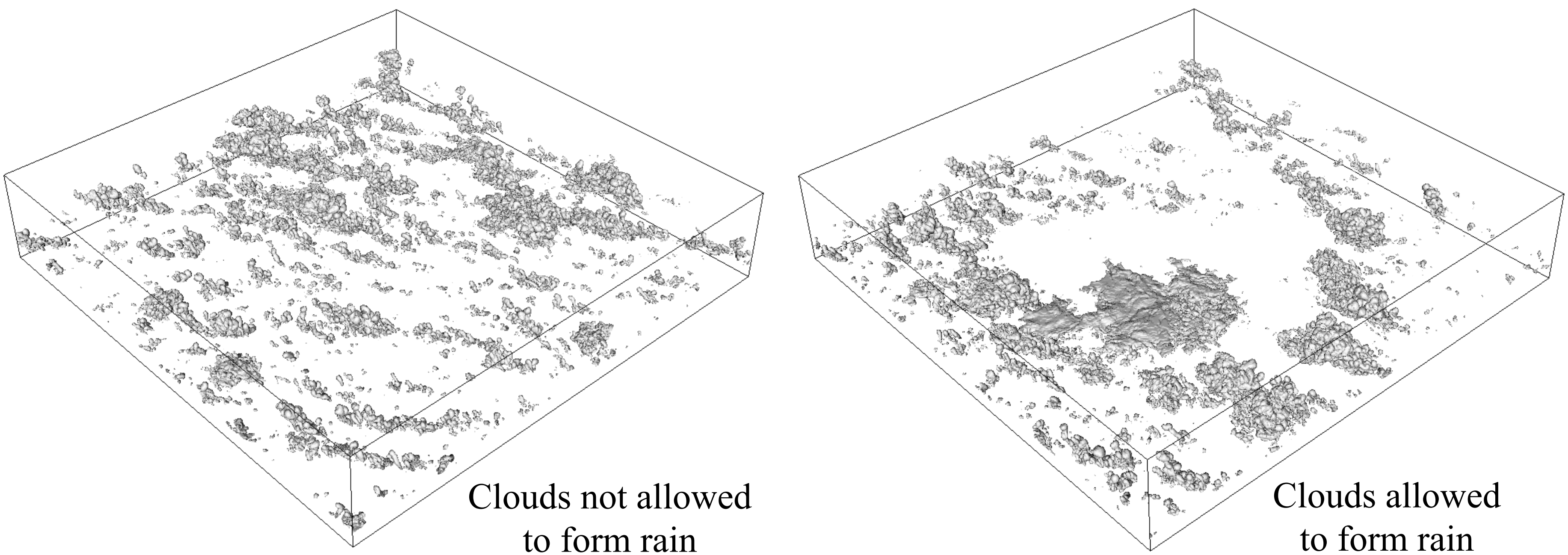
Effects of precipitation – LES

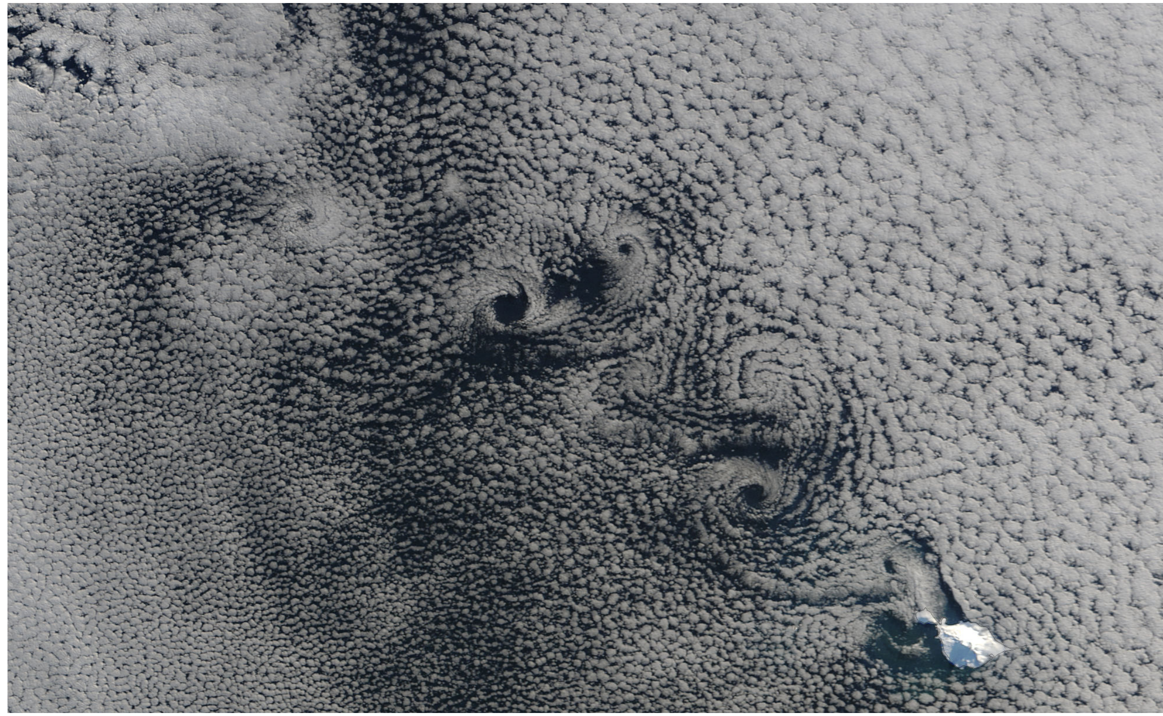
- About 20% of the total precipitation in the subtropics is produced by cumulus clouds
- Simulations with and without precipitation
- Simulations show that the onset of precipitation significantly alters the organization of convection



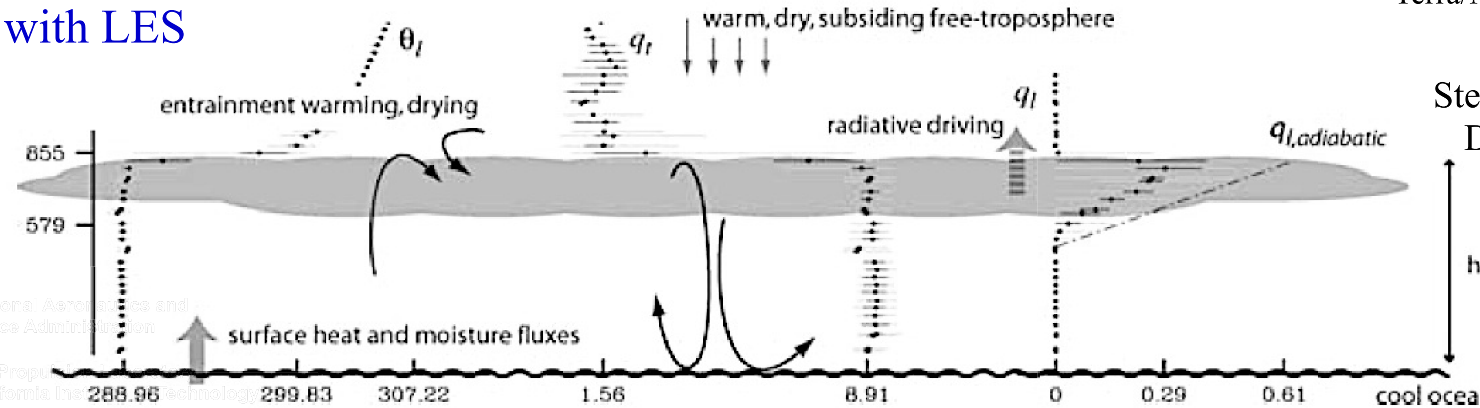
Matheou et al. (2011)

Stratocumulus clouds

- Large albedo
  - Note variability!
- Very low cloud tops
- Strong inversion at top of boundary layer
  - Cloud evaporates when warm and dry air mixes with BL air
- Very important cloud regime for climate but extremely difficult to simulate even with LES



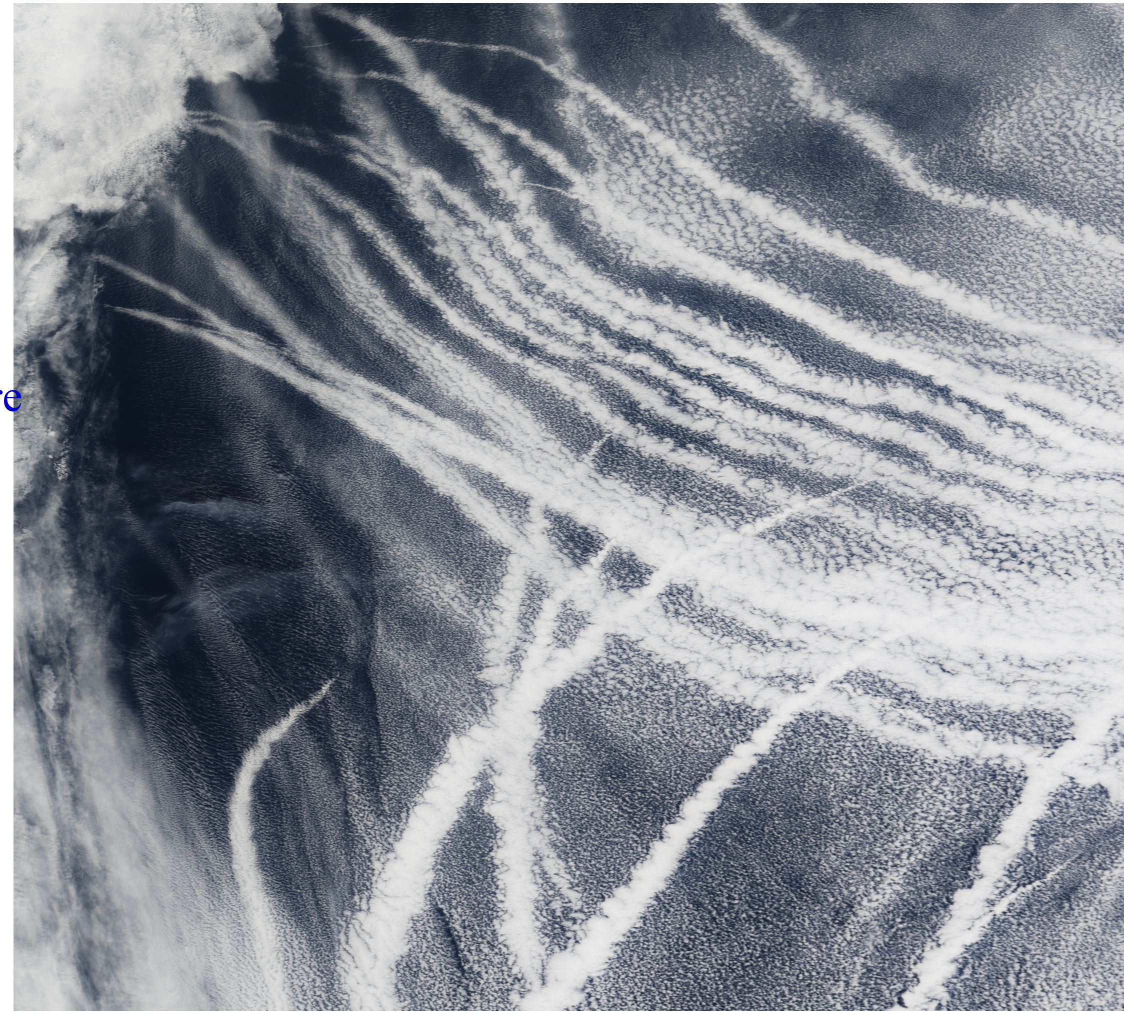
Terra/MODIS image



Stevens (2005)
DYCOMS II

Ship tracks

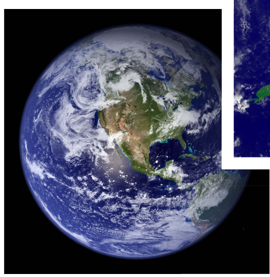
- In remote areas of the ocean cloud condensation nuclei (CCN) are scarce
- Less and larger cloud droplets
- However… particles produced in ship exhausts are more numerous than the natural CCN over the ocean
- The amount of water in the cloud remains the same but the cloud becomes brighter
- What if we purposely brighten clouds?
  - Geoengineering



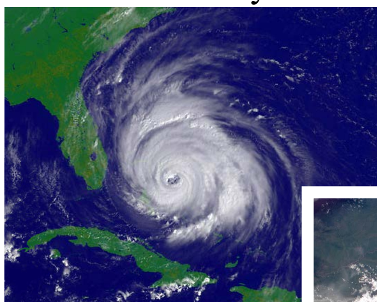
Terra/MODIS image
Northern Pacific

Scales of atmospheric motions and models

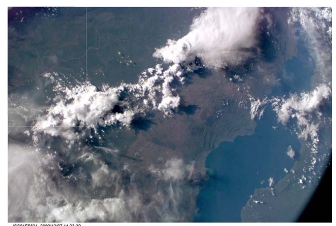
Global circulation



Mesoscale organization
Weather/storm systems



Convection Organization – Thunderstorms



Individual Clouds

Cloud–Environment interactions

Small-scale turbulence

10 000 km — 1000 km — 10 km — 1 km — 100 m — 1 mm

$L/\Delta x \approx 10^8$

Global circulation model (GCM)
$L/\Delta x \approx 10^2$

Resolved | **Modeled**

Large-eddy simulation
$L/\Delta x \approx 10^3$

Input from large scales | **Resolved** | **Modeled**

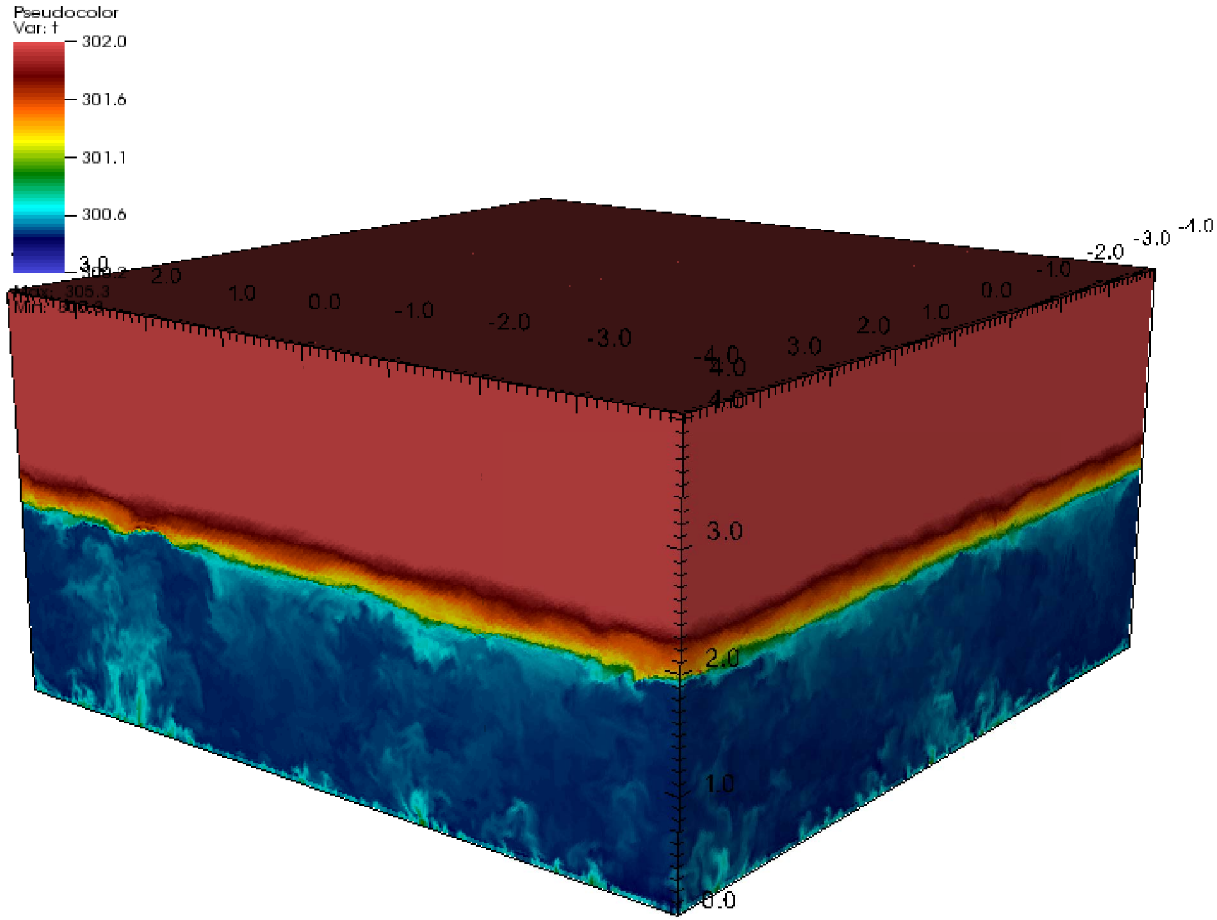
Direct numerical sim.
$L/\Delta x \approx 10^3$

Forcing / Turbulence production /Large scales

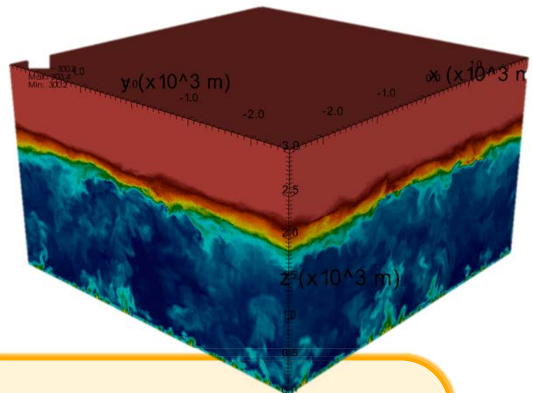
Computational cost increases ~ $(L/\Delta x)^4$

Images: NASA, B. Stevens, Chung & Matheou

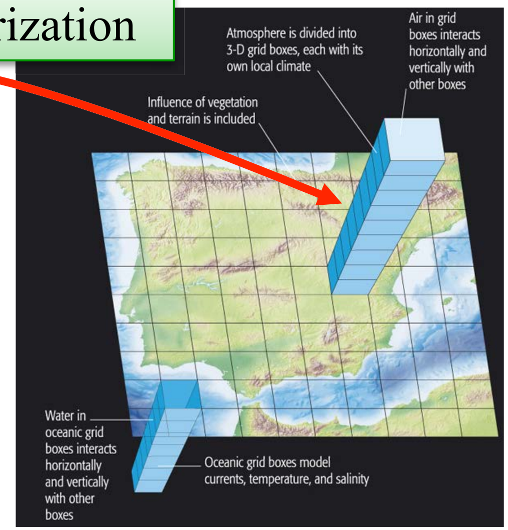
Dry convective boundary layer



Formulating and evaluating a parameterization



parameterization



https://www.e-education.psu.edu/meteo469/node/140
Mann & Kump, *Dire Predictions*

LES: Three-dimensional

$$\frac{\partial \tilde{\theta}(t,x,y,z)}{\partial t} = \ldots$$

$$\Delta x = 5-100\,\mathrm{m}$$

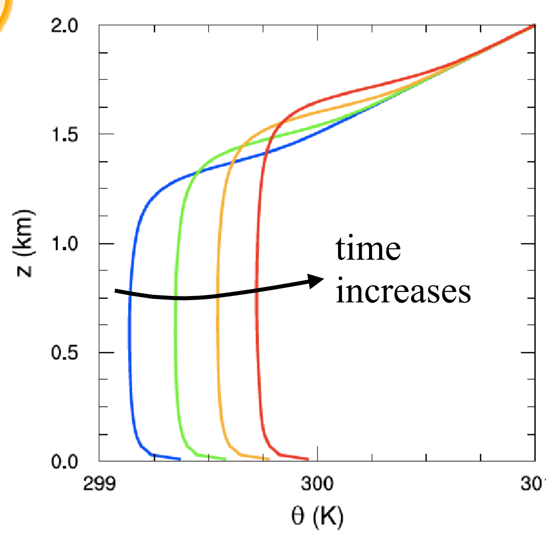
Single column model: one-dimensional

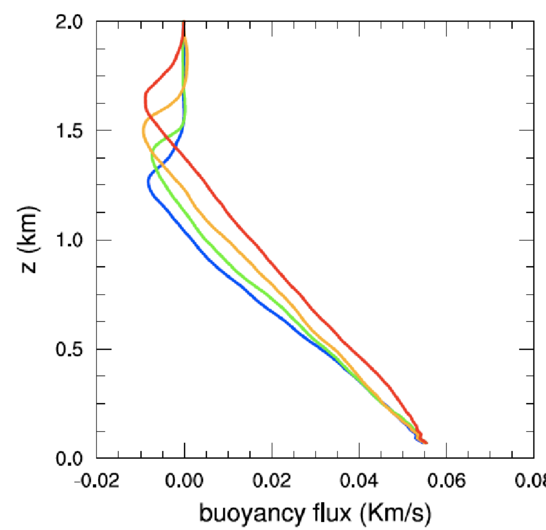
$$\frac{\partial \overline{\theta}(t,z)}{\partial t} = \frac{\partial \overline{w'\theta'}}{\partial z} + F_\theta$$

$$\Delta x \approx 100\,\mathrm{km}$$

Potential temperature:

$$\theta = T\left(\frac{p_{z=0}}{p}\right)^{\frac{\gamma-1}{\gamma}}$$





Eddy-diffusivity/Mass flux approach (EDMF)

- Goal: Evolution of boundary layer given surface conditions and free troposphere

$$\frac{\partial \overline{\theta}(t,z)}{\partial t} = -\frac{\partial \overline{w'\theta'}}{\partial z} + F_\theta$$

- Decompose turbulence in the boundary layer into two types of motions
  - Strong organized updrafts
  - Remaining (more isotropic) turbulent motions
- Turbulent flux can be decomposed as (Siebesma & Cuijpers 1995)

$$\overline{w'\theta'} = a_{\mathrm{u}}\overline{w'\theta'}^{\mathrm{u}} + (1-a_{\mathrm{u}})\overline{w'\theta'}^{\mathrm{e}} + \underbrace{a_{\mathrm{u}}(w_{\mathrm{u}} - \overline{w})}_{M\text{: convective mass flux}}(\theta_{\mathrm{u}} - \theta_{\mathrm{e}})$$

updraft area << 1

$$\overline{w'\theta'} \approx \overline{w'\theta'}^{\mathrm{e}} + M(\theta_{\mathrm{u}} - \theta_{\mathrm{e}})$$

$$\overline{w'\theta'} \approx -K\frac{\partial \overline{\theta}}{\partial z} + M(\theta_{\mathrm{u}} - \overline{\theta})$$

- Eddy diffusivity closure

$$K(z) = 0.25\, l\left(\tfrac{1}{2}\overline{u_i u_i}\right)^{1/2}$$

$$\frac{1}{l} = \frac{1}{l_1} + \frac{1}{l_2} \qquad l_1(z) = 0.5\frac{z_i}{w_*}\left(\tfrac{1}{2}\overline{u_i' u_i'}\right)^{1/2} \qquad l_2(z) = \kappa z\left(1.0 - 100.0\frac{z}{L}\right)^{0.2}$$

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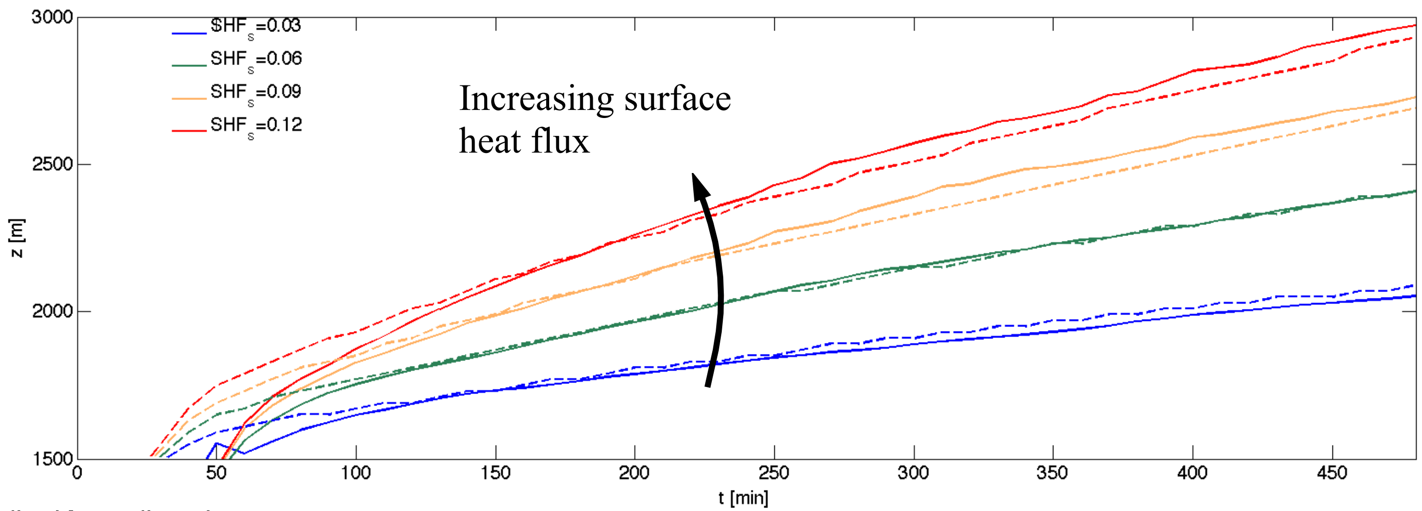
Updraft parameterization

- Mass flux: $$\frac{\partial \theta_u}{\partial z} = -\varepsilon_l\left(\theta_u - \bar{\theta}\right)$$ $$w_u \frac{\partial w_u}{\partial z} = -\varepsilon_l w_u^2 + 2.0g\left(\frac{\theta_u}{\bar{\theta}} - 1\right)$$
- Lateral entrainment coefficient is a key parameter $\varepsilon_l(z) = \frac{0.7}{l}$
- More details in:

  Witek, Teixeira & Matheou, 2011: An integrated TKE-based eddy-diffusivity/mass-flux boundary layer closure for the dry convective boundary layer, *J. Atmos. Sci.*, **68**(7), 1526–1540.

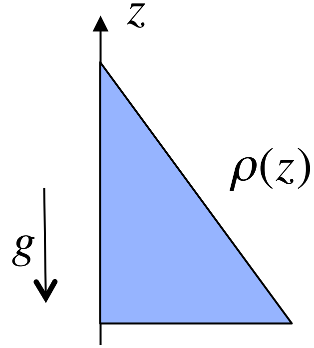


LES:
continuous lines

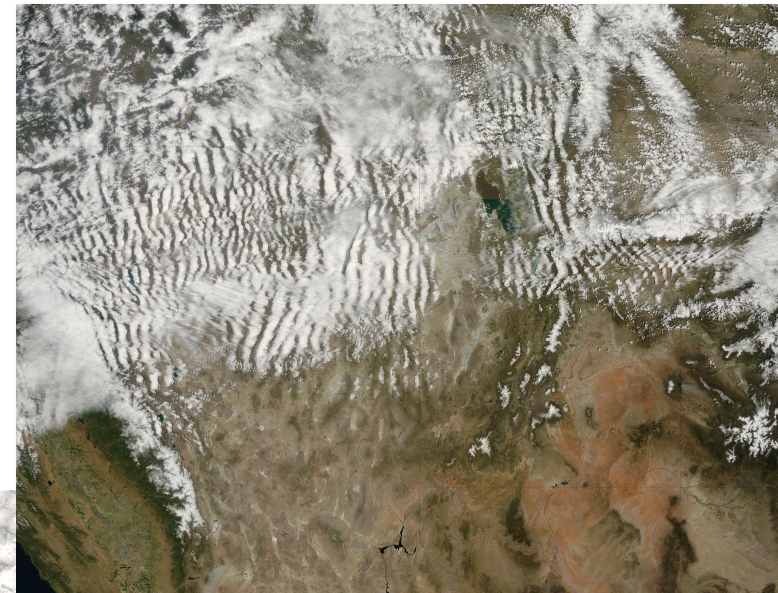
Single-column model:
dashed lines

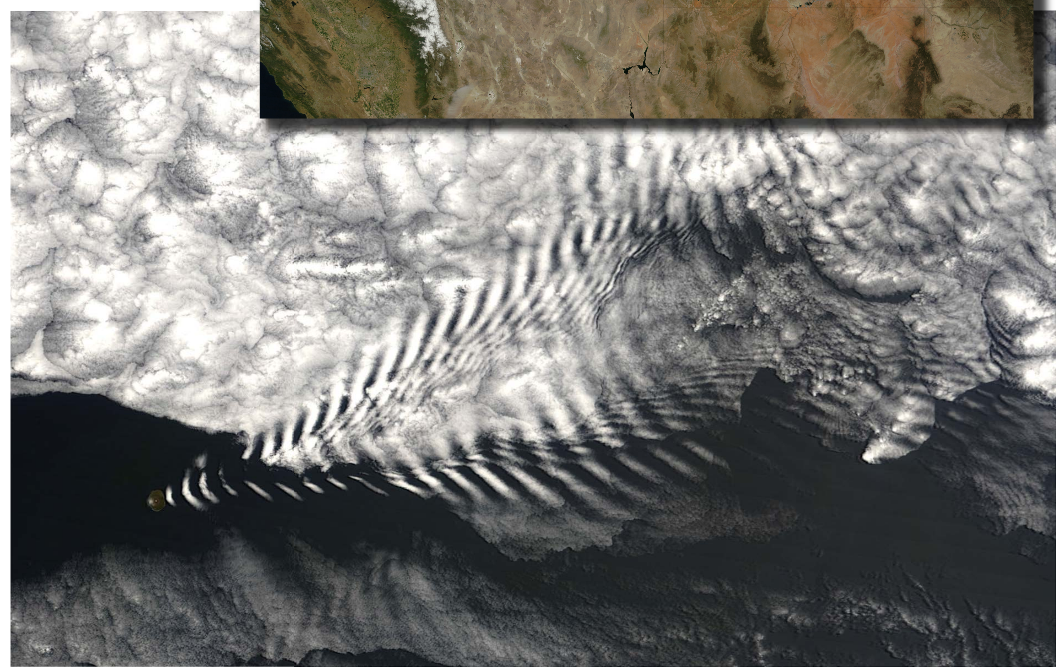
Buoyancy frequency: a measure of stratification

$$N \equiv \sqrt{\frac{\partial \bar{b}}{\partial z}} = \sqrt{-\frac{g}{\rho_0}\frac{\partial \bar{\rho}(z)}{\partial z}}$$



- Stably stratified fluids support waves
- Waves can travel for large distances





Waves become visible when water vapor condenses at the crests

Length scales and non-dimensional numbers

- Length scales
  - Ozmidov
    - Buoyancy effects strongly felt for scales > l_O
    - Associated with the vertical extent of overturning motions

$$l_O \equiv \left(\frac{\varepsilon}{N^3}\right)^{1/2}$$

  - Buoyancy
    - Thickness of layers, also associated with instabilities (e.g., Billant & Chomaz 2001)

$$l_b \equiv U / N$$

  - Obukhov
    - Length (height) at which buoyancy becomes important

$$L \equiv \frac{u_*^2}{\kappa b_*} = \frac{\left(\overline{u'w'}\right)^{3/2}}{\kappa \overline{b'w'}}$$

  - Corrsin
    - Anisotropic shear effects strongly felt for scales > l_C

$$l_C \equiv \left(\frac{\varepsilon}{S^3}\right)^{1/2} \quad \text{with } S = \frac{\partial \overline{u}}{\partial z}$$

- Non-dimensional parameters
  - Richardson number

$$Ri \equiv \frac{N^2}{S^2}$$

    - "Overall" Ri (Turner 1973, p12)

$$Ri_o \equiv g\frac{\rho'}{\rho_0}\frac{L}{U^2} = \frac{1}{Fr^{1/2}}$$

  - Flux Richardson number

$$Rf \equiv \frac{\overline{b'w'}}{\overline{u'w'}\,S}$$

  - Buoyancy Reynolds number
    - Also called the intensity parameter, I, by oceanographers

$$\mathcal{R} \equiv \frac{\varepsilon}{\nu N^2} \sim \left(\frac{l_O}{\eta}\right)^{4/3}$$

The critical Richardson number

- Assume:
  - Homogeneous turbulence with constant N^2 and $S = \partial u / \partial z$
  - Stationary turbulence

$$\frac{1}{2}\frac{\partial \overline{u_i' u_i'}}{\partial t} = -\overline{u'w'}S - \overline{w'b'}N^2 + \varepsilon = 0$$

$$\frac{1}{2}\frac{\partial \overline{b_i' b_i'}}{\partial t} = -\overline{w'b'}N^2 - \chi = 0$$

- Relate dissipation terms to large-scale: $\varepsilon = \left(\overline{u_i' u_i'}\right)^{3/2} \big/ l_u$ and $\chi = \overline{b_i' b_i'}\left(\overline{u_i' u_i'}\right)^{1/2} \big/ l_b$
- Assume normalized fluxes are constant, i.e., independent of stratification (Townsend 1958) **not quite right!**

$$k_u = \frac{\overline{u'w'}}{\overline{u_i' u_i'}} = \text{const} \qquad k_b = \frac{\overline{b'w'}}{\left(\overline{u_i' u_i'}\right)^{1/2}\left(\overline{b_i' b_i'}\right)^{1/2}} = \text{const}$$

- Relation between flux and gradient Ri

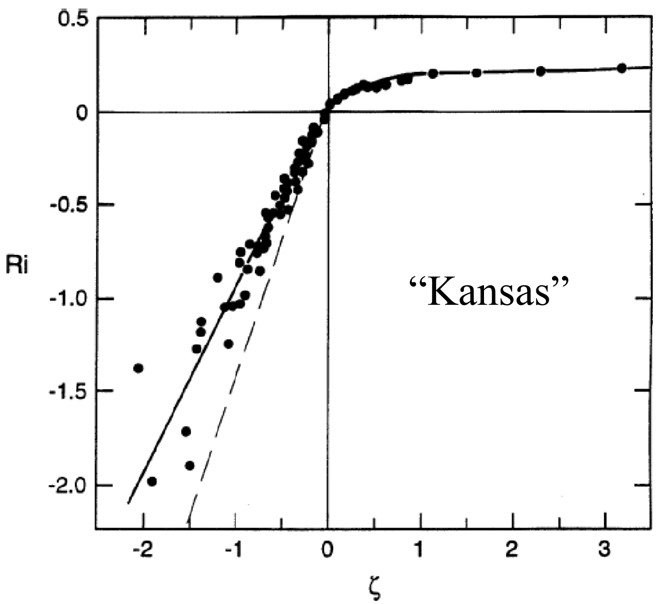
$$Rf = \frac{1}{2}\left[1 - \left(1 - 4\frac{l_b}{l_u}\frac{k_b^2}{k_u^2}Ri\right)^{1/2}\right]$$

- For $\overline{u_i' u_i'} > 0$ Ri attains a critical value. Ri_{crit} is the largest Ri value for which solutions of the balance equations with non-zero TKE can be found

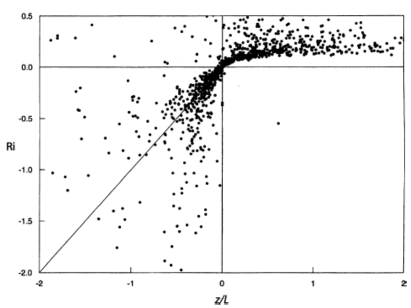
$$Ri_{\text{crit}} = \frac{1}{4}\frac{l_u}{l_b}\frac{k_u^2}{k_b^2} \approx 0.2$$

Miles–Howard instability criterion for inviscid shear flow: $Ri < 1/4$

Measurements of Ri_{crit}

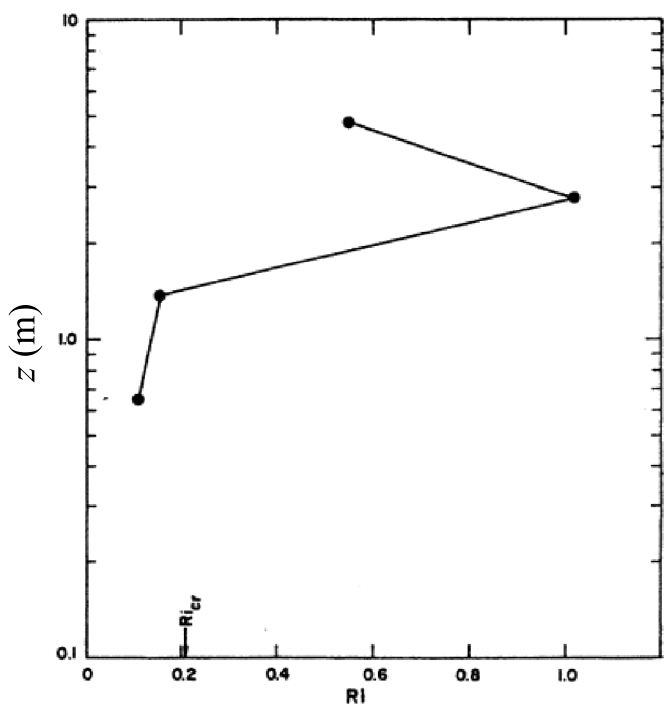


“Kansas”



“Wyoming”

Businger & Arya (1974)
Businger (2005)



Laminar

Turbulent

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Sources of anisotropy

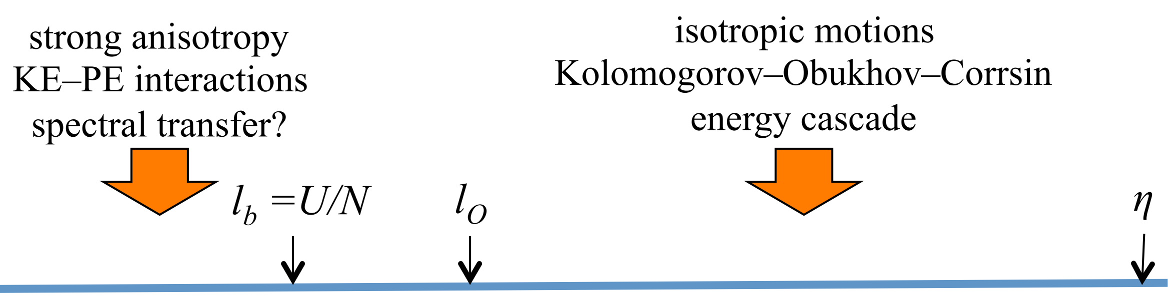
- Shear is typically a production mechanism
  - Large scale anisotropy because of shear, Corrsin scale

$$l_C \equiv \left(\frac{\varepsilon}{S^3}\right)^{1/2}$$

- Vertical stratification inhibits vertical motions since vertical displacements occur at the expense of potential energy
  - No penalty for horizontal motions
  - This is a stabilizing mechanism
- The Ozmidov scale is used to characterize the largest scale that can overturn
  - See also discussion in Appendix of Riley & Lindborg (2008)
  - Scales smaller than l_O are expected to be nearly isotropic

$$l_O \equiv \left(\frac{\varepsilon}{N^3}\right)^{1/2}$$

- As stratification increases motions become “flatter”
  - Pancake eddies, layers, steps, quasi-horizontal motions
- Stratified turbulence is not the same as 2D turbulence!
  - Dissipation mechanism is markedly different



DNS Param.:

$Re_\lambda \approx 400$

$$S^* = S\frac{\overline{u_i'u_i'}^2}{\varepsilon} = 4-12$$

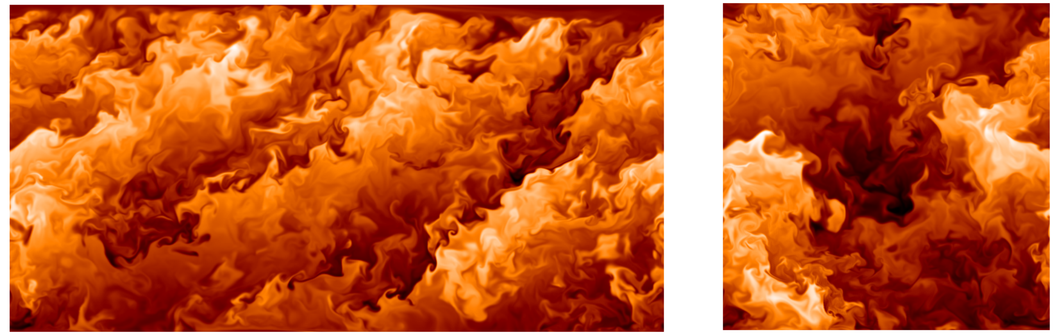
$Sc = 0.7$

$k_{\max}\eta = 1.2$

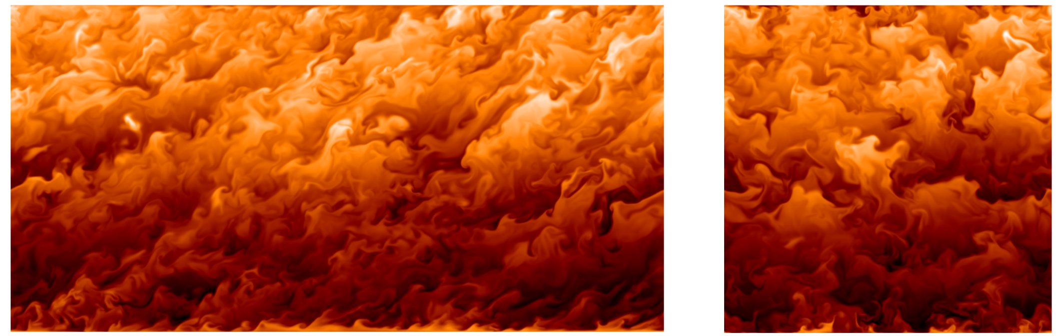
$$L_* = \frac{L u_*}{\nu}$$

Chung & Matheou, in review, *JFM*

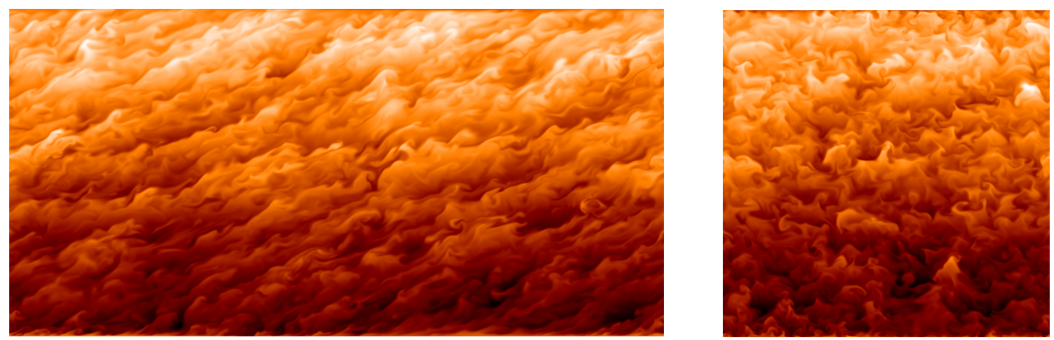
$Ri = 0.026$
$Rf = 0.026$
$\mathcal{R} = 10^4$
$L_* = 10^4$



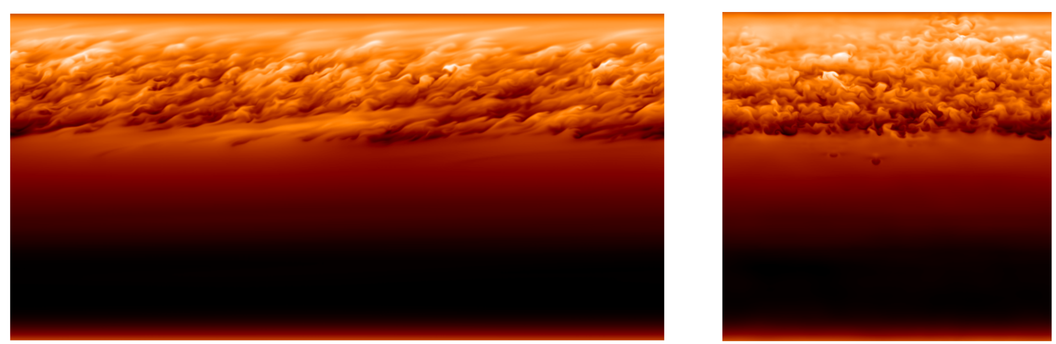
$Ri = 0.053$
$Rf = 0.062$
$\mathcal{R} = 10^3$
$L_* = 1500$



$Ri = 0.125$
$Rf = 0.134$
$\mathcal{R} = 10^2$
$L_* = 200$



$Ri = 0.157$
$Rf = 0.161$
$\mathcal{R} = 10$
$L_* = 30$



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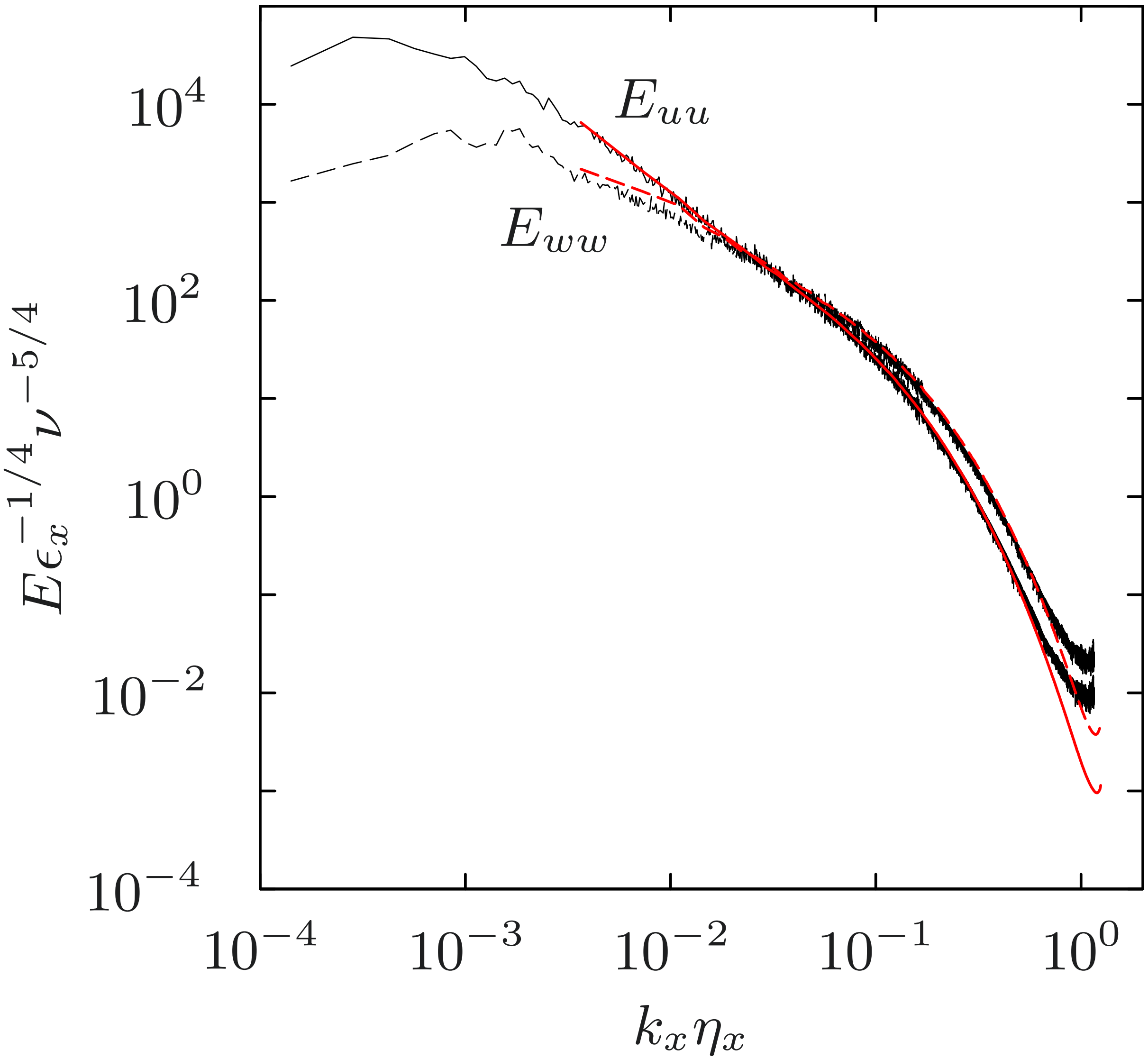
Validation – Isaza et al. (2009)



Turbulence collapse

- What parameter characterizes a stably stratified flow?

- Can *Ri* characterize the state of turbulence in a stably stratified flow?
  - Not really: need a parameter that accounts for the destabilizing action of *Re* and the stabilizing action of *Ri*
  - Need to consider two parameters, e.g., *Re* and *Ri*. But in the strongly stratified regime *Ri* drops out.
- Criteria for turbulence collapse
  - Buoyancy Reynolds number criterion, $\mathcal{R} > 10$
    - Laval et al. (2003), Riley & de Bruyn Kops (2003), Shih et al. (2005), Brethouwer et al. (2007), Ivey et al. (2008)
    - There are earlier references in the Oceanography literature!
  - $\mathcal{R} = \varepsilon/\nu N^2 \sim Re\, Ri \sim Re\, Fr^{-2}$
  - Flores & Riley (2011) find $L_* \approx 100$ a criterion for turbulence collapse

$$L_* = \frac{L u_*}{\nu}$$

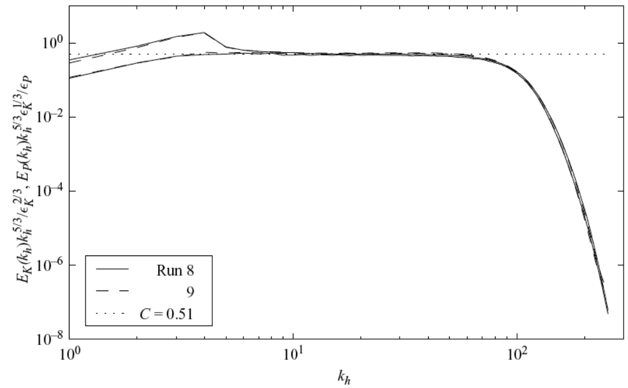
Energy transfer and spectra in stratified turbulence

- Lindborg (2006)

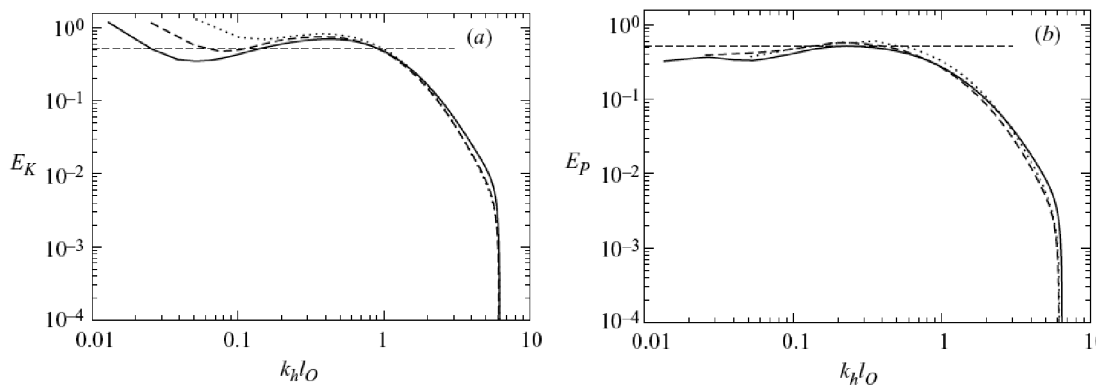
$$E_K = 0.5\varepsilon^{2/3} k_h^{-5/3}$$

$$E_P = 0.5\chi\,\varepsilon^{-1/3} k_h^{-5/3}$$

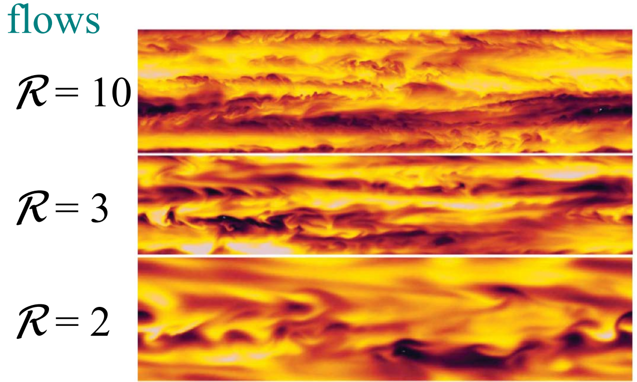
  – Forward (towards smaller scales) kinetic and potential energy cascades



- Brethouwer et al. (2007)
  – Confirm $\sim k^{-5/3}$ and forward energy cascade

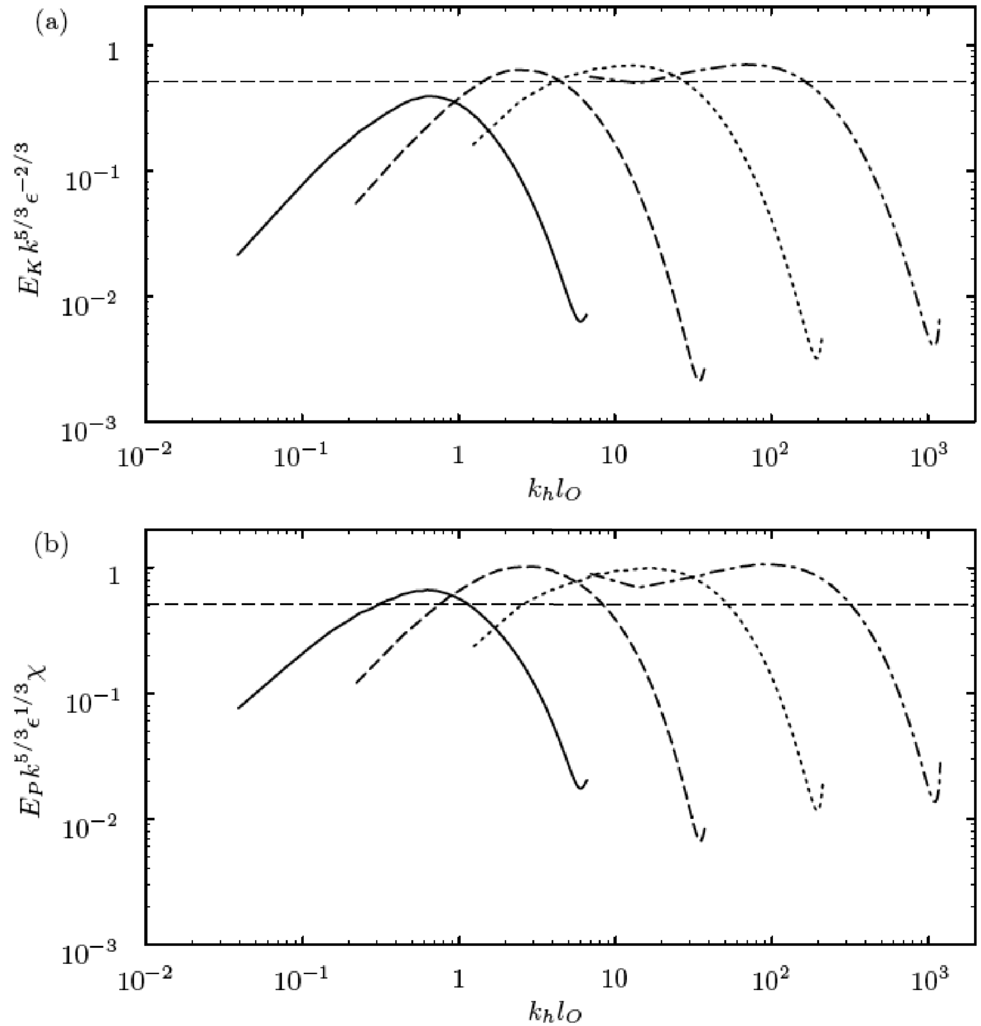


- Riley & Lindborg (2008)
  – Why $\sim k^{-5/3}$ if kinetic–potential energy interactions?
  – Possible links to measured spectra in geophysical flows
- It is difficult to infer the dynamics of atmospheric turbulence based on forced strongly stratified turbulence "DNS" studies

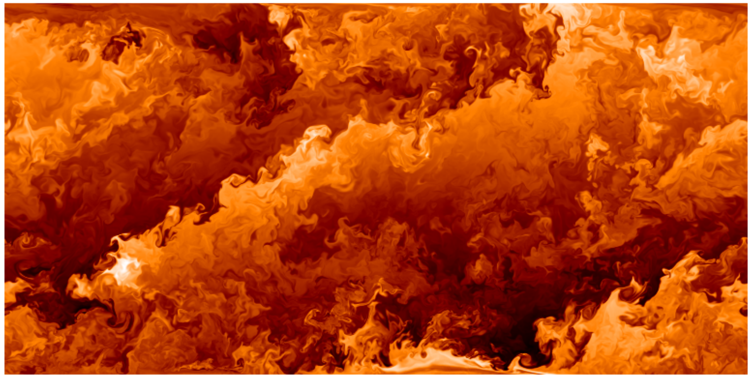


Horizontal spectra

- Lindborg scaling for present DNS



A surface layer view



- The simulation domain imposes a length scale to a flow that does not have an outer scale
- The group $\dfrac{L_z}{u_*}\dfrac{\partial \overline{u}}{\partial z}$ is a universal constant
- Use $\dfrac{\kappa z}{u_*}\dfrac{\partial \overline{u}}{\partial z} \equiv 1$ to define a confinement scale z

  in a similar way that the height z is a confinement scale for turbulent eddies in the surface layer
- Assumption: $z_* \equiv \dfrac{z u_*}{\nu} >> 1$
- Non-dimensional form of streamwise spectra

$$E_{uu}(k_x) = \alpha_1 \varepsilon^{2/3} k_x^{-5/3} \longrightarrow \frac{k_x E_{uu}(k_x)}{u_*^2} = \frac{\alpha_1}{(2\pi\kappa)^{2/3}} \phi_\varepsilon^{2/3} \left(\frac{k_x z}{2\pi}\right)^{-2/3} \qquad \phi_\varepsilon \equiv \frac{\varepsilon \kappa z}{u_*^3}$$

$$E_{bb}(k_x) = \beta_1 \varepsilon^{-1/3} \chi k_x^{-5/3} \longrightarrow \frac{k_x E_{bb}(k_x)}{b_*^2} = \frac{\beta_1}{(2\pi\kappa)^{2/3}} \phi_\varepsilon^{-1/3} \phi_\chi \left(\frac{k_x z}{2\pi}\right)^{-2/3} \qquad \phi_\chi \equiv \frac{\chi \kappa z}{u_* b_*^2}$$

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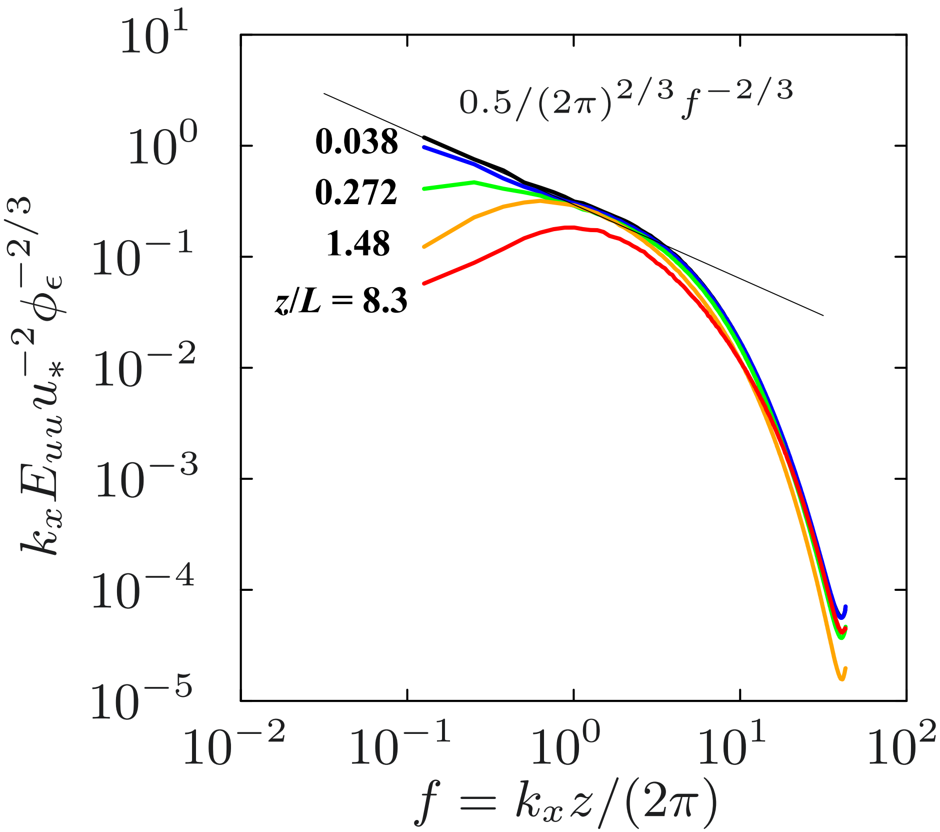
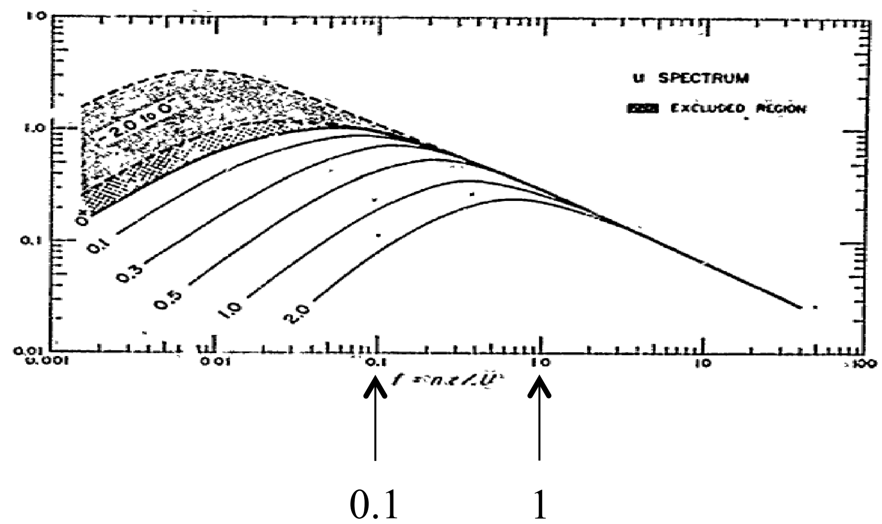
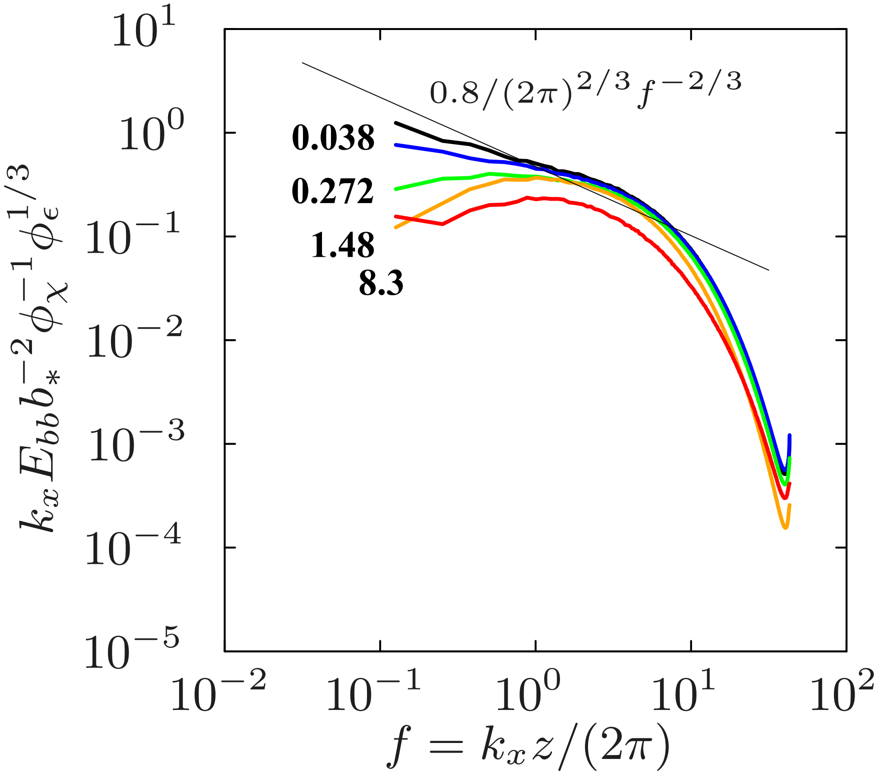
Kaimal et al. (1972) scaling – Streamwise velocity

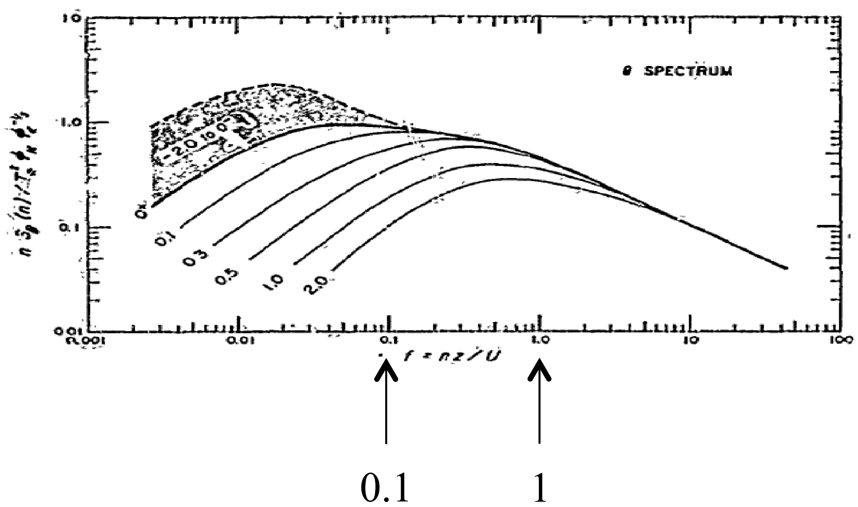




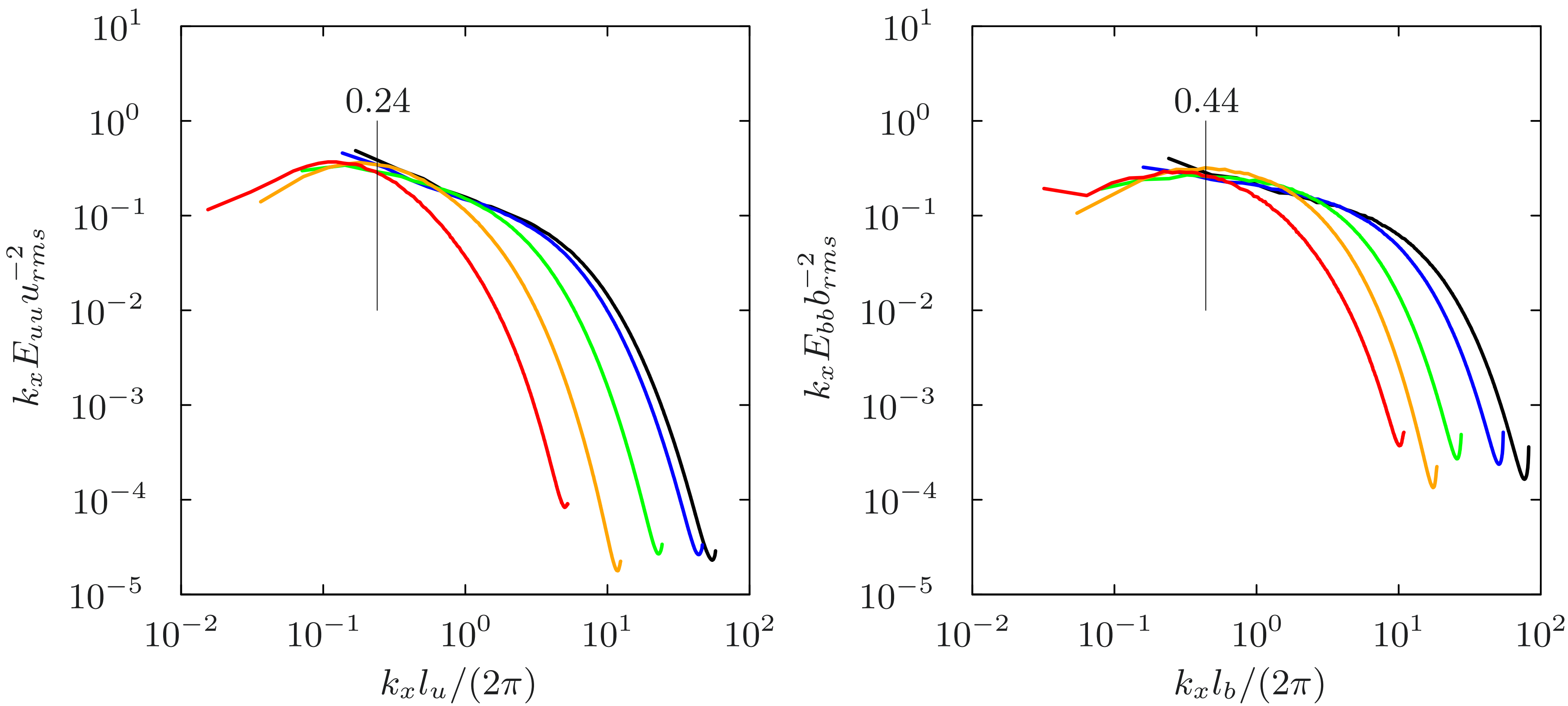
J. C. Kaimal, J. C. Wyngaard, Y. Izumi & O. R. Coté: 1972 Spectral characteristics of surface-layer turbulence, *Quart. J. R. Met. Soc.*, **98**, 563–589.

Kaimal et al. (1972) scaling – buoyancy





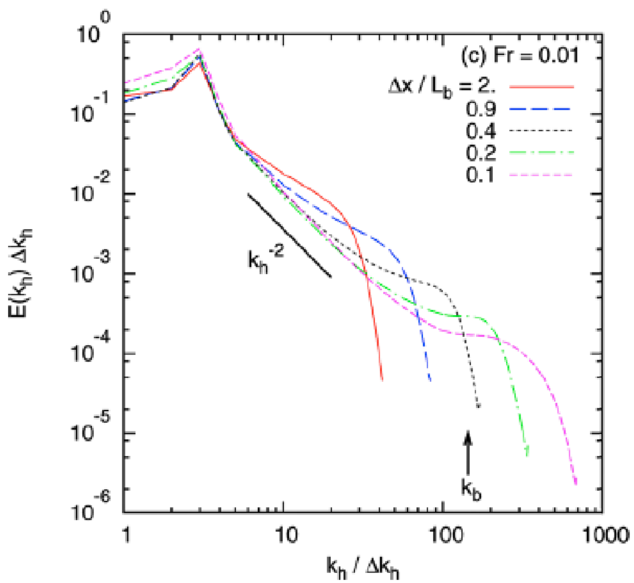
Scaling of energy peaks (Kaimal et al. 1972)

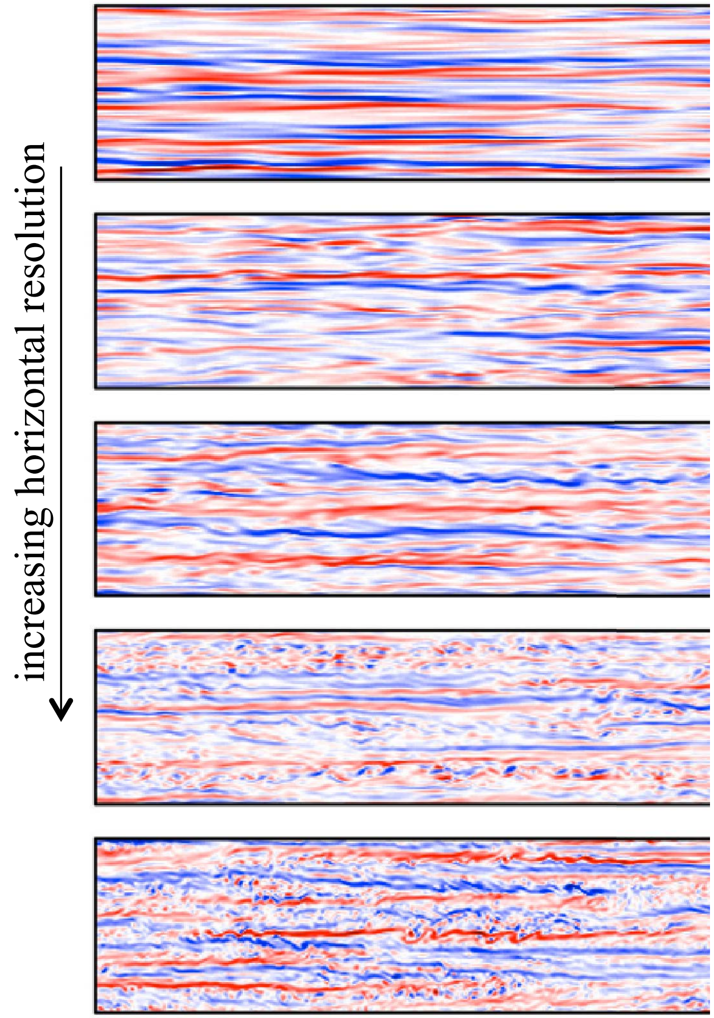


$$l_u \equiv \frac{u_{rms}^3}{\varepsilon} \qquad l_b \equiv \frac{b_{rms}^3 \varepsilon^{1/2}}{\chi^{3/2}}$$

Horizontal resolution effects (Waite 2011)

- Waite (2011) varied horizontal resolution with respect to the buoyancy scale, $l_b = U/N$
- Stratified turbulence simulations are sensitive to horizontal resolution, $\sim k_h^{-2}$ instead of $\sim k_h^{-5/3}$
- Postulate the existence of a microscale range between l_b and l_O
  - $l_O / l_b \sim Fr^{1/2}$
- Deloncle et al. (2008) "shortcut" to dissipation notion (direct transfer of energy from large to buoyancy scale) may explain some of the findings

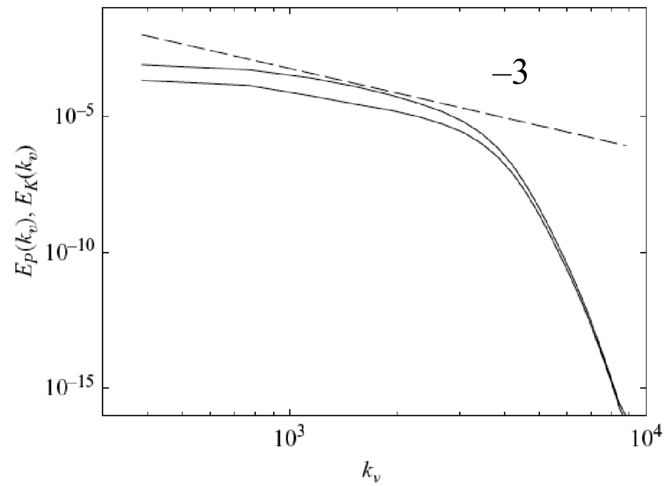


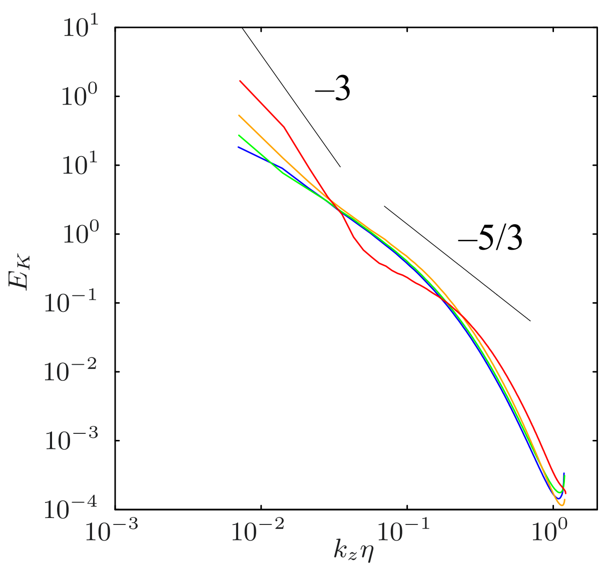


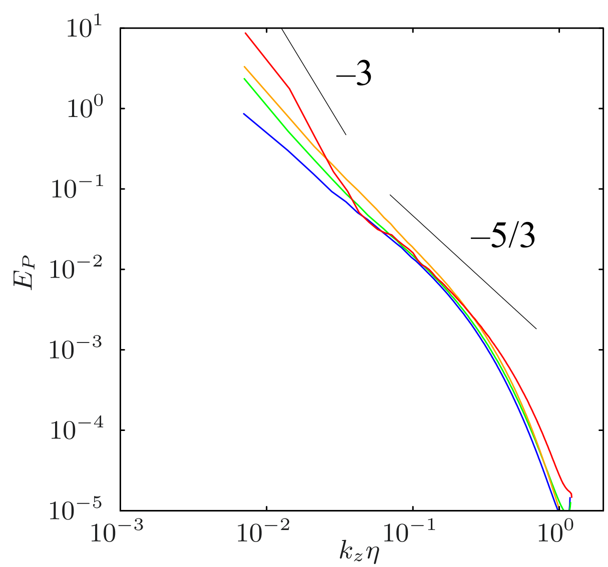
Vertical spectra

- Lindborg (2006)

$$E_K \sim E_P \sim N^2 k_z^{-3}$$





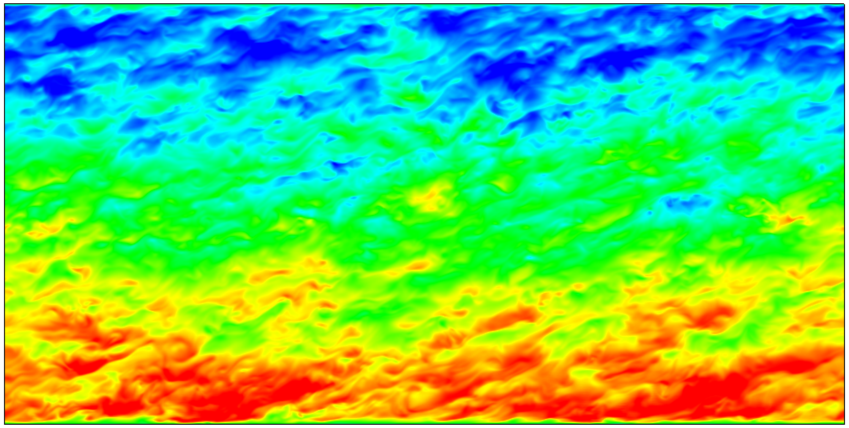


Formation of layers

- With increasing stratification an inverse transfer of energy to $k_h = 0$ and the formation of layers is observed
  - Godeferd & Cambon (1994), Laval et al. (2003), + *many others*
- What characterizes the formation of layers? Ri, $\mathcal{R} = Re\, Fr^{-2}$, Fr ?
  - Although commonly observed, the exact processes governing the dynamics of the layers in a fully turbulent flow is not well understood
- Formation of layers in a fully turbulent flow differs from energy transfer in a laminar flow transitioning and breaking down into a turbulent-like state (Deloncle et al. 2008)

$\mathcal{R}$	Re_λ	
	Volume	Plane
∞	376	370
10^4	377	248
10^3	381	144
10^2	787	84
10	2152	48

$\mathcal{R} = 100$ – Streamwise velocity



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Formation of layers and atmospheric turbulence

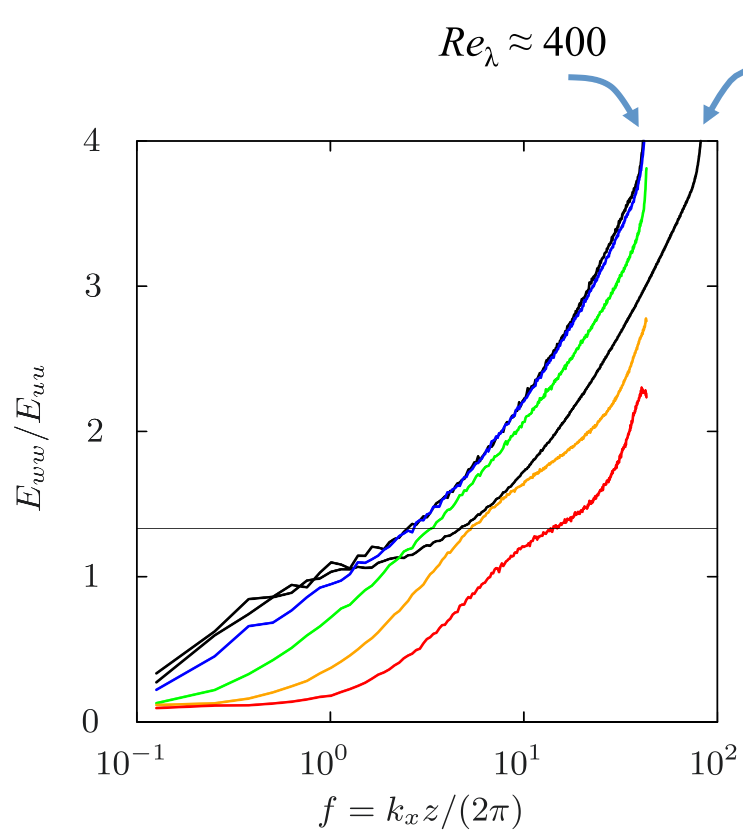
- Lumley & Panofsky (1964, p 211):

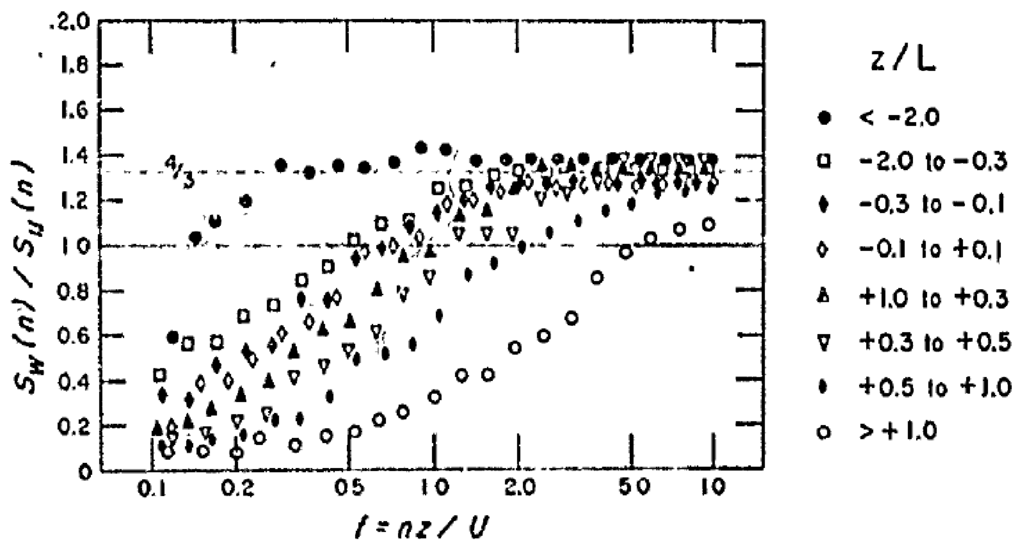
  "Finally we may speculate about the structure of turbulence some distance from the ground. Aside from clouds and the relatively infrequent clear-air turbulence, flights are generally smooth. Yet balloon soundings and detailed Doppler-radar wind measurements indicate some rather general variations of horizontal velocity components with horizontal scale of several miles and more. It is thus likely that the atmosphere is generally filled with quasi-horizontal eddies of sizes greater than 2 miles, which probably derive their energy from horizontal wind shear."

- …moreover, the effects of rotation cannot be ignored for mesoscale atmospheric motions

Isotropy – E_{ww}/E_{uu}

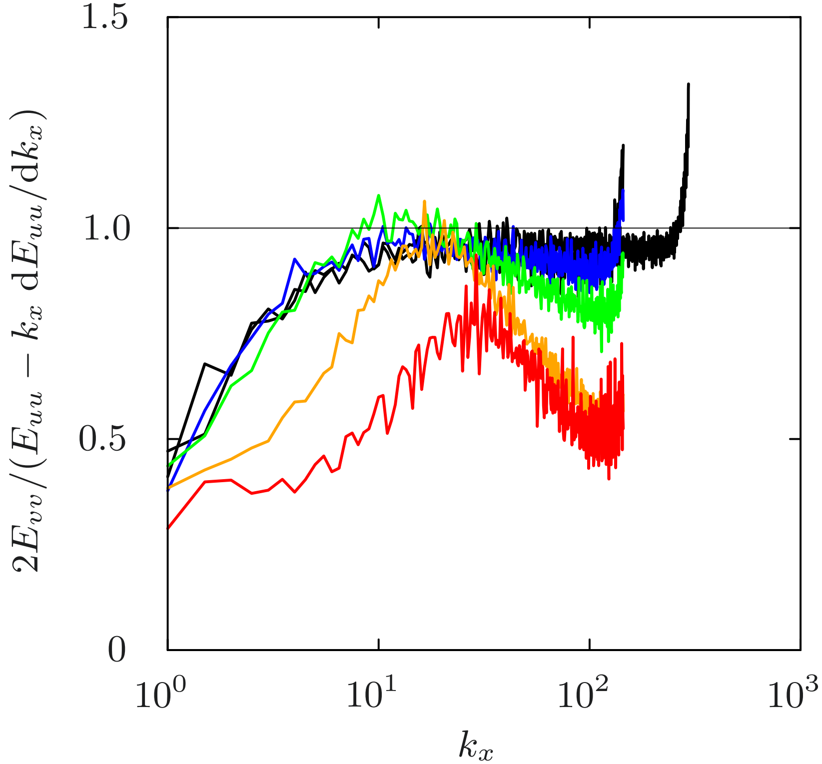
- DNS does not have an extensive inertial range

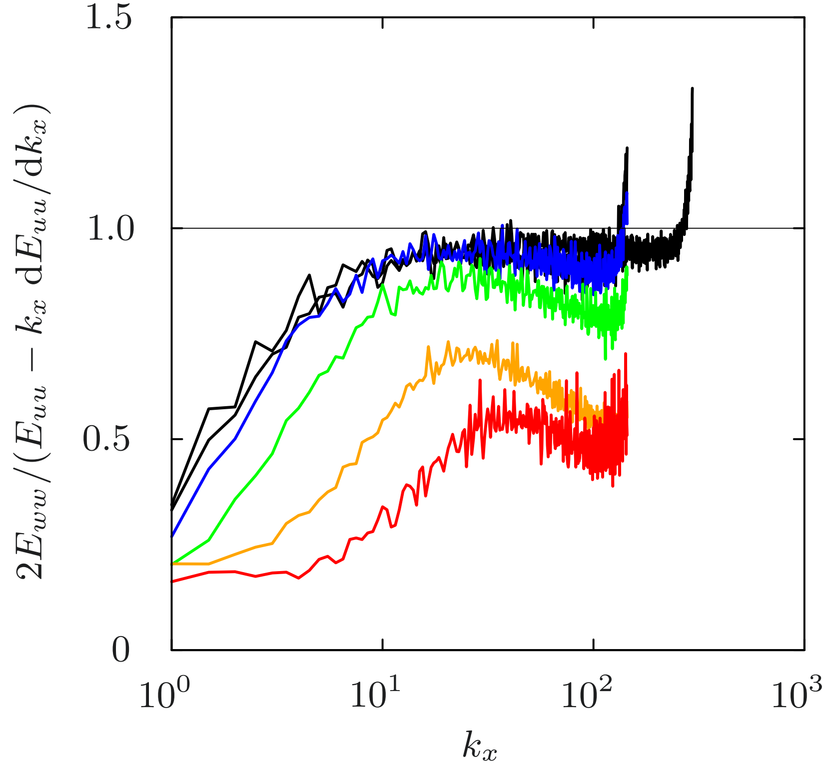




Isotropy – Spectra

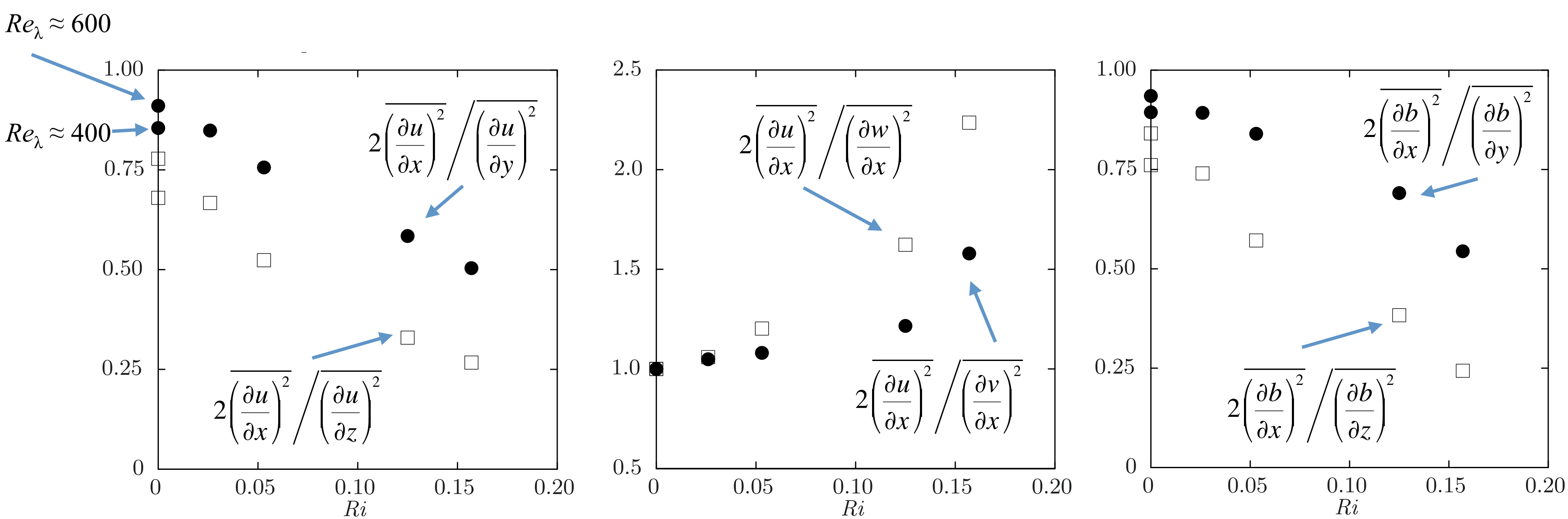
$$E_{vv} = E_{ww} = \frac{1}{2}\left(E_{uu} - k_x \frac{\mathrm{d}E_{uu}}{\mathrm{d}k_x}\right)$$





Local (small scale) isotropy statistics

- *Ri* and *Re* are not independent in the DNS!



Summary and conclusions

- We cannot understand and predict weather and climate without understanding the dynamics of atmospheric turbulence
  - Turbulent transport is a first order effect in the atmosphere and ocean

- Simulations of stratified turbulence are extremely difficult because of the presence of multiple length scales
  - Difficulties in achieving the necessary scale separation
  - What are the consequences for subgrid scale modeling?
    - What happens when the buoyancy and/or overturning scales become smaller than grid resolution?

- Need to be cautious
  - Many times investigators extrapolate their conclusions to atmospheric turbulence

- Clouds strongly affect the energy balance of the planet
  - Low cloud feedbacks is the largest uncertainty in present day climate models

Coming soon!