大气环流

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课程简介

- 课程要求
  - 熟悉大气环流的基本分布和形态
  - 掌握各主要环流系统的维持和变化机制
  - 建立各环流系统形成的物理模型
  - 了解现阶段的大气环流模式
  - 知道大气环流方向有待解决的科学问题

大气环流概述－观测资料

- 地面资料 （陆地，航船）
- 探空资料
- 卫星资料
- Aircraft report (AIREP)
- 海洋资料

data assimilation → "froze" analysis technique

technique always in development, e.g. using models with higher resolution, better parameterization

↓

reanalysis data

ERA-40

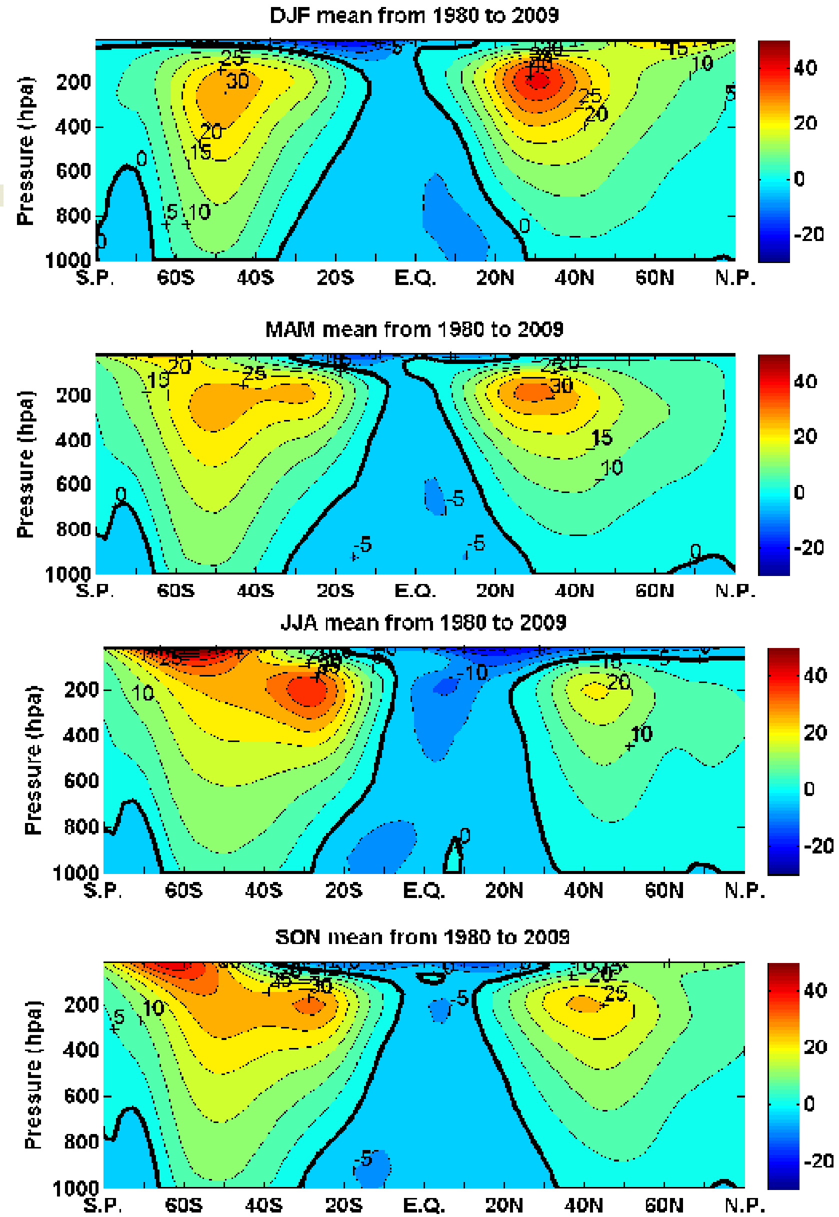
NCEP/NCAR

NCEP/NCAR 再分析资料

- 物理量及分类
  - **A** (strongly influenced by *observed data*, hence, in the most reliable class): geopotential height, T, u, v...
  - **B** (although there are *observational data* directly affecting the value of the variable, the *model* also has a strong influence): relative humidity, w(vertical velocity), lowest level u and v...
  - **C** (*no observations* directly affecting the variable, so that it is derived *solely from the model* forced by the data assimilation): radiation fluxes, surface heat fluxes, cloud forcing, precipitation rate...
  - **D** (*fixed* from climatological values and does *NOT depend on models*): surface roughness, surface geopotential height...

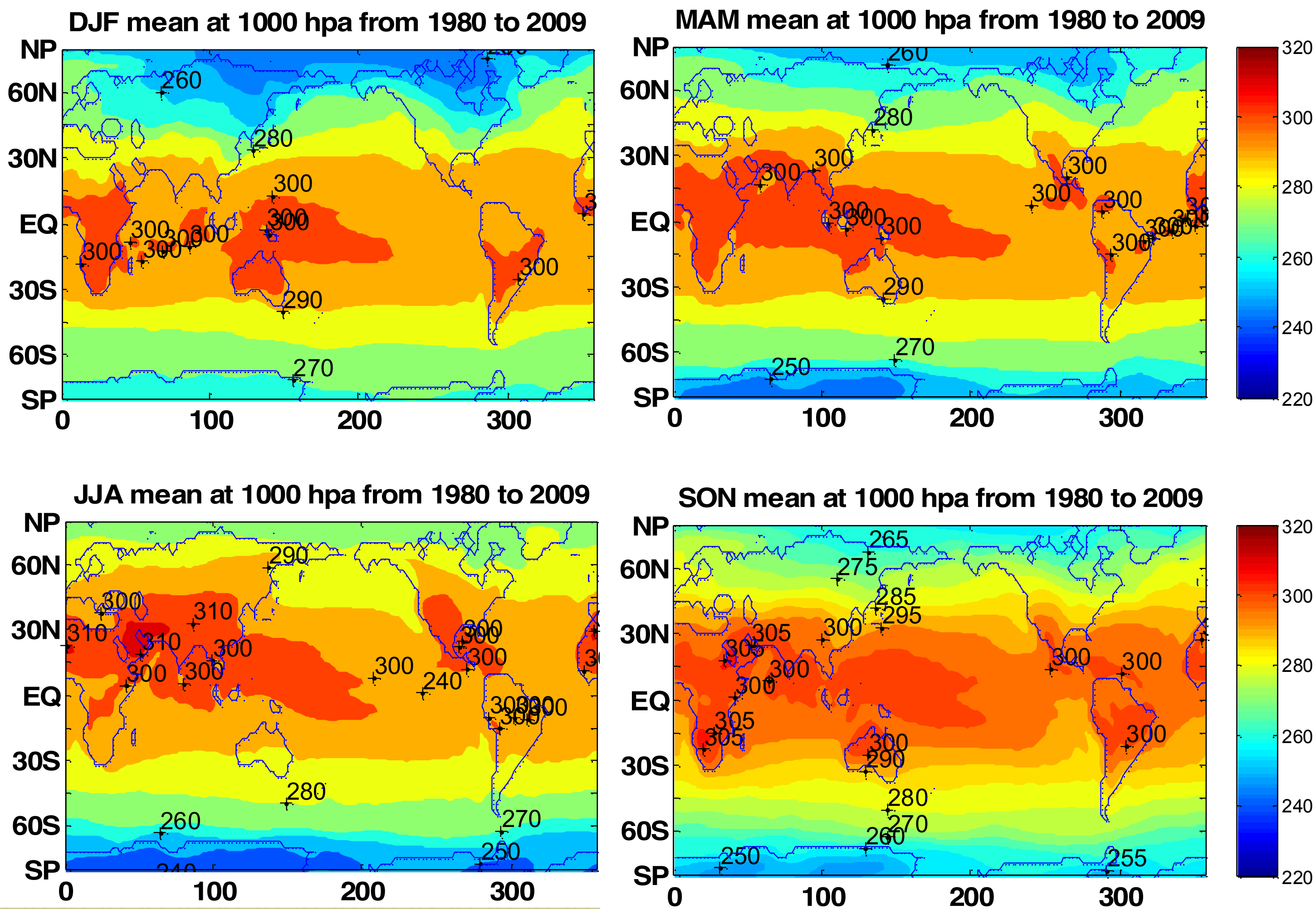
Question 1

- **风场:**
- 西风带，东风带
  - 对流层，平流层
  - 随纬度的变化
- 水平，垂直结构
- 季节变化：位置，强度，副热带与温带急流
- 与温度场的匹配

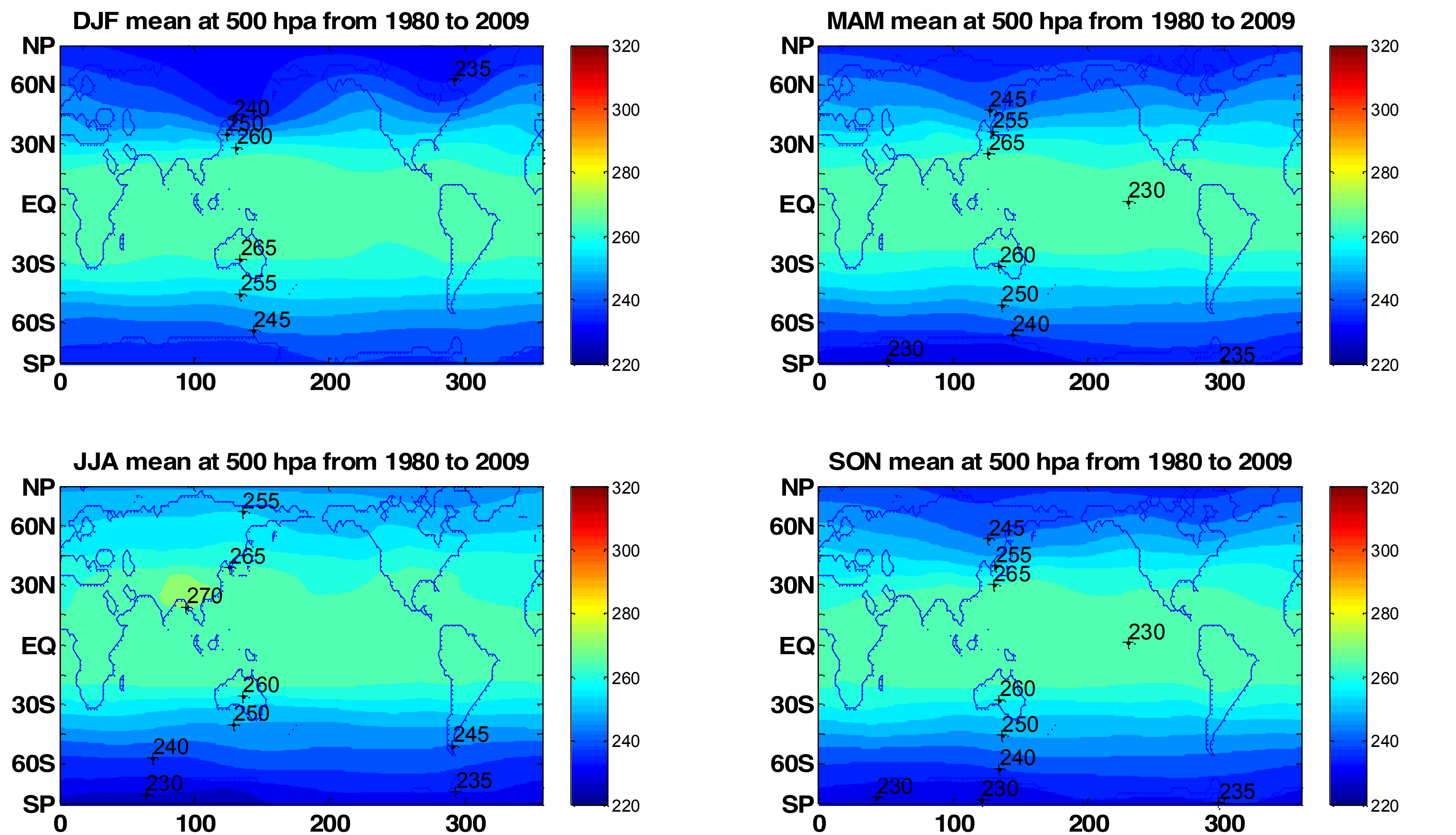


Question 2

- 温度场－纬向
  - 受海陆分
  - 垂直变化
  - 季节变化



Question 2


DJF mean at 500 hpa from 1980 to 2009
MAM mean at 500 hpa from 1980 to 2009
JJA mean at 500 hpa from 1980 to 2009
SON mean at 500 hpa from 1980 to 2009
NP
60N
30N
EQ
30S
60S
SP
0
100
200
300
220
240
260
280
300
320
235
240
250
260
265
255
245
230
270

大气环流概述－分析方法

- Decompose a field in both time and space: (the results depend on whether we average in time or zonally first.

$$
\begin{aligned}
\overline{[vT]} &= \overline{[([v]+v^*)([T]+T^*)]} = \overline{[v][T]} + \overline{[v^*T^*]} \\
&= \overline{([\bar{v}]+[v]')([\bar{T}]+[T]')} + \overline{[v^*T^*]} \\
&= [\bar{v}][\bar{T}] + \overline{[v]'[T]'} + \overline{[v^*T^*]}
\end{aligned}
$$

transport by the steady mean meridional circulation | transport by the transient mean meridional circulation | transport by the spacial eddy circulation

Alternatively,

$$
\begin{aligned}
[\overline{vT}] &= [\overline{(\bar{v}+v')(\bar{T}+T')}] = [\bar{v}\bar{T}] + [\overline{v'T'}] \\
&= [([\bar{v}]+\bar{v}^*)([\bar{T}]+\bar{T}^*)] + [\overline{v'T'}] \\
&= [\bar{v}][\bar{T}] + [\bar{v}^*\bar{T}^*] + [\overline{v'T'}]
\end{aligned}
$$

transport by stationary eddies | transport by transient eddies

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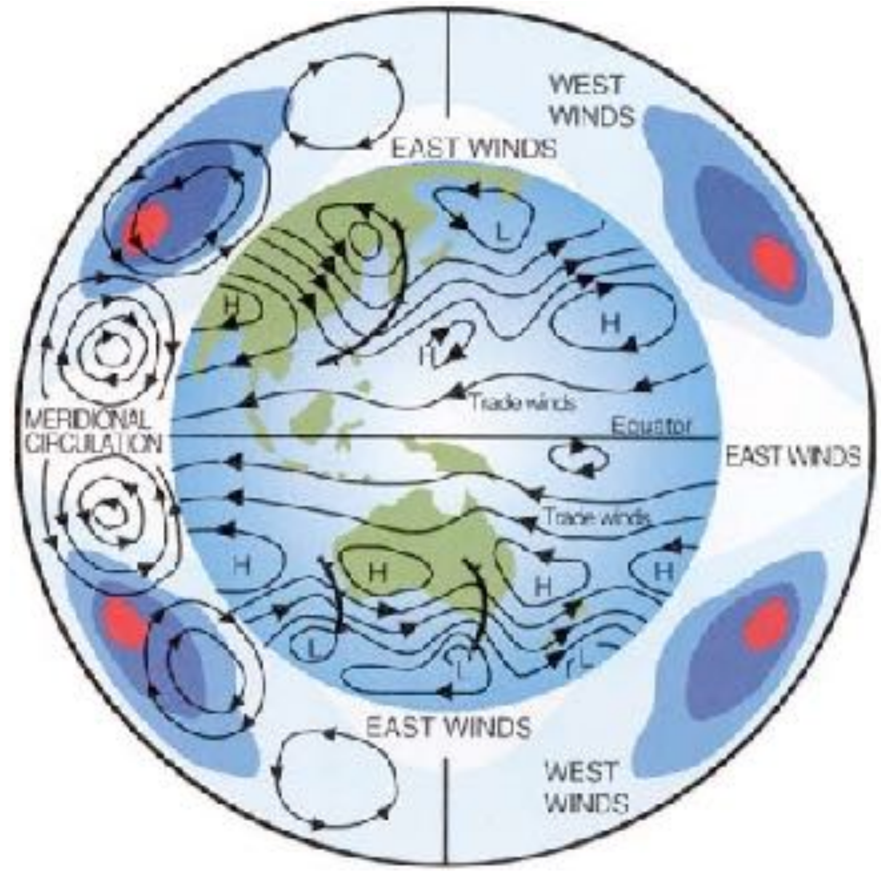
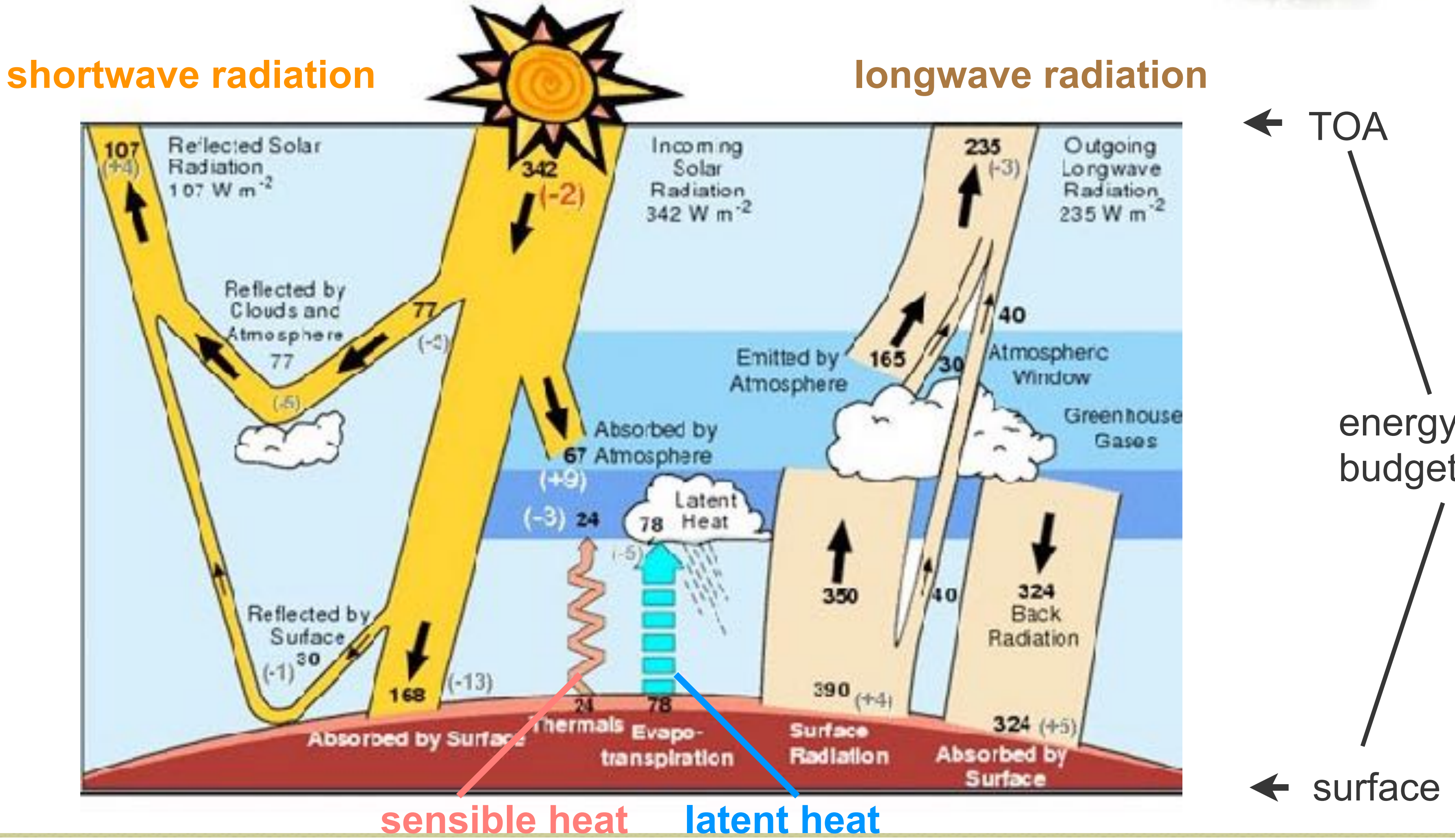
  - 辐射强迫
  - 下界面强迫
- 经向环流系统（纬向平均环流, zonally averaged circulations）：
  - Hadley 环流
  - Ferrel 环流、急流、波流相互作用
- 纬向环流系统（non-zonal circulations）：
  - Storm tracks
  - Monsoon
  - ENSO and Walker circulation
- 不同复杂度的大气环流模式
- 全球暖化背景下的大气环流

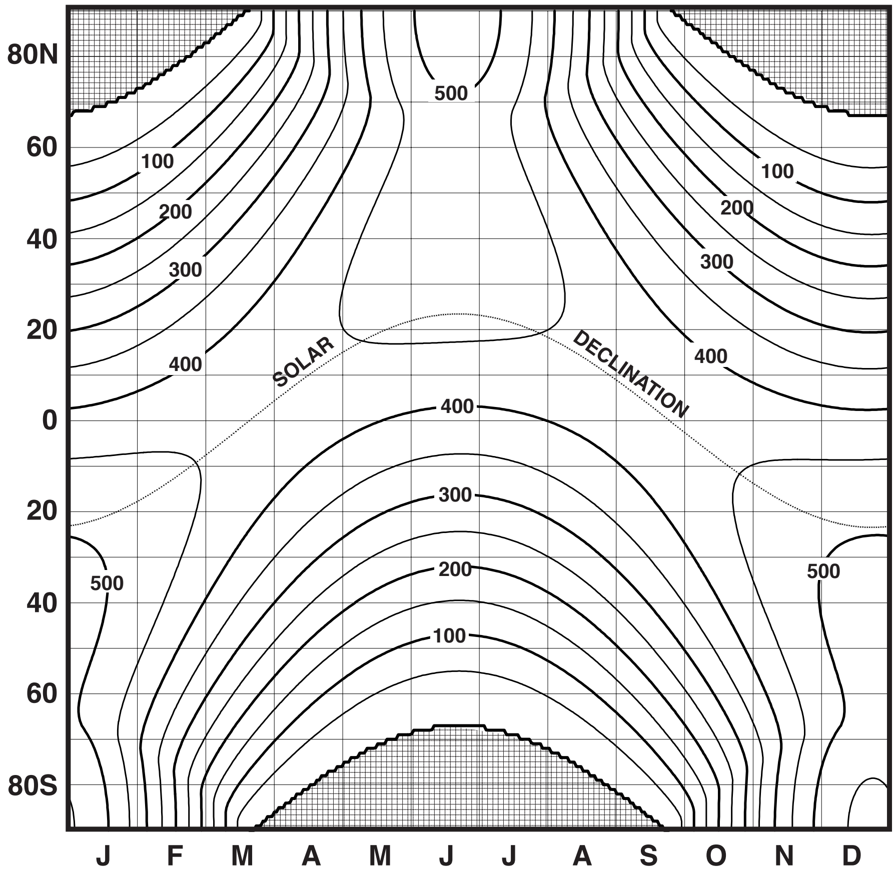


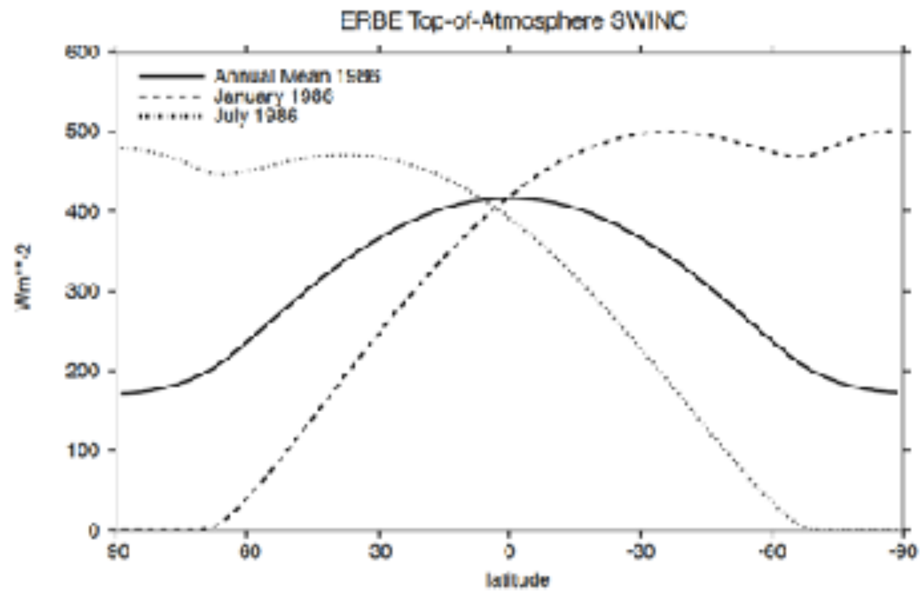
From the solar radiation…



From the solar radiation…

- At TOA





Figures: the zonally averaged incident solar radiation, observed in the Earth Radiation Budget Experiment (ERBE).

From the solar radiation…

Incident solar radiation	340 W/m^2
Planetary albedo	0.3
Absorbed solar radiation	240 W/m^2
Outgoing longwave radiation	240 W/m^2

Table: globally and annually averaged TOA radiation budget

SW ~ LW

$S(1-\alpha)$

← TOA

energy budget

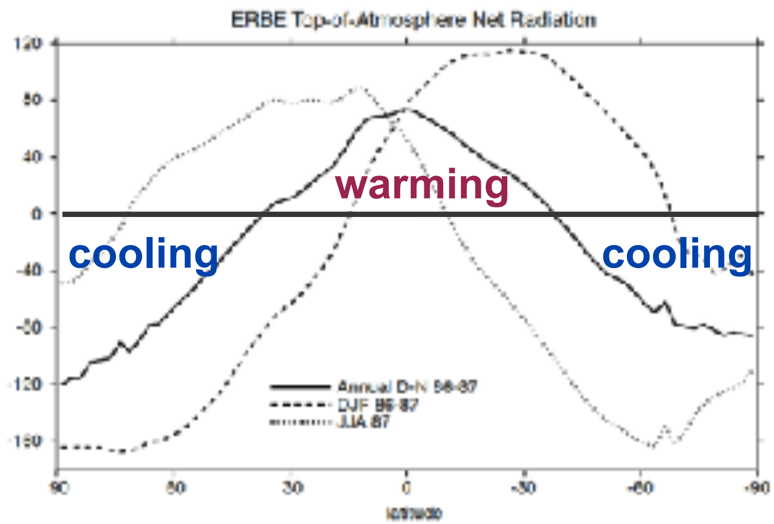
Absorbed solar radiation (240 - 176)	64 W m^{-2}
Net emitted terrestrial radiation (-240 + 73)	-167 W m^{-2}
Net radiative heating	-103 W m^{-2}
Latent heat input	79 W m^{-2}
Sensible heat input	24 W m^{-2}

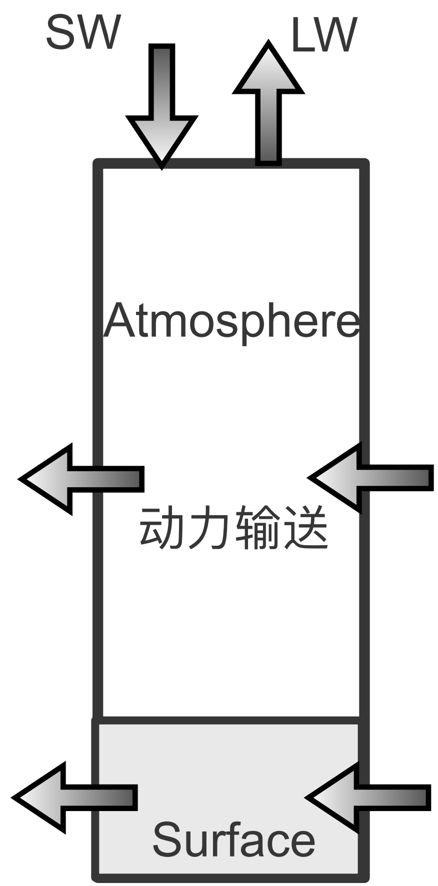
Table: globally and annually averaged atmosphere energy budget

Absorbed solar (SW)	176 W m^{-2}
Downward infrared (LW↓)	312 W m^{-2}
Upward infrared (LW↑)	-385 W m^{-2}
Net longwave (LW)	-73 W m^{-2}
Net radiation (SW + LW)	103 W m^{-2}
Latent heat (LH)	-79 W m^{-2}
Sensible heat (SH)	-24 W m^{-2}

e: SW(net) + LW(net) + LH +SH ~0 ← surface

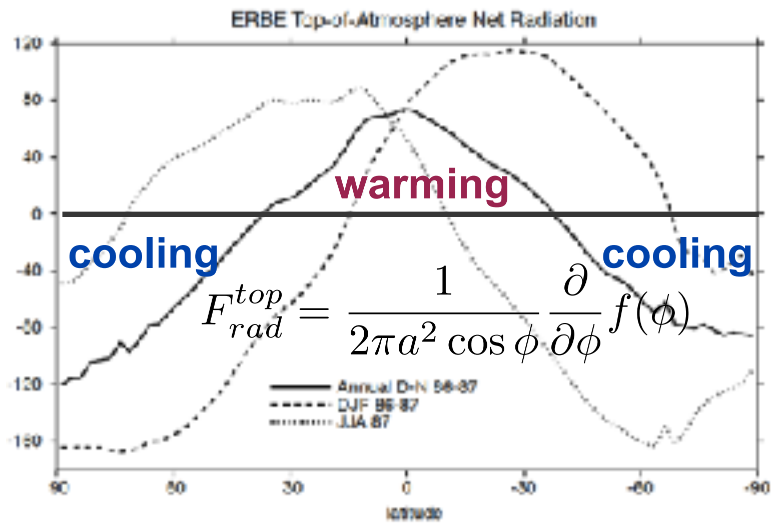
Table: globally and annually averaged surface energy budget

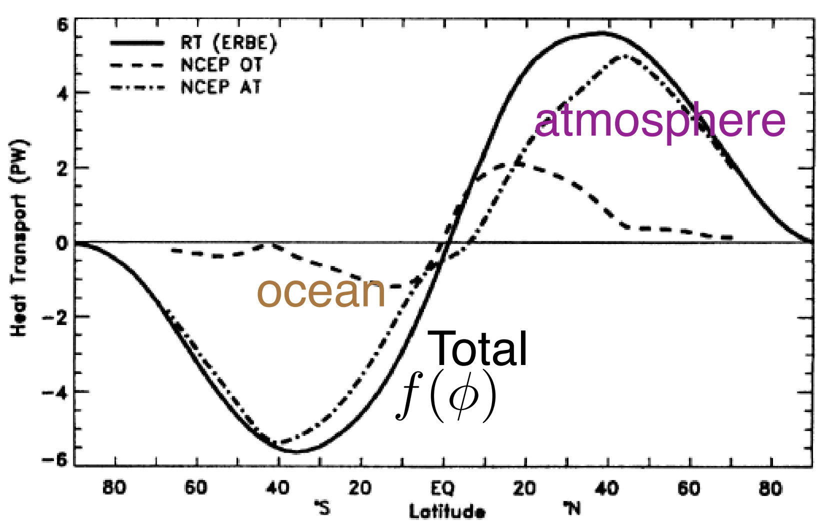




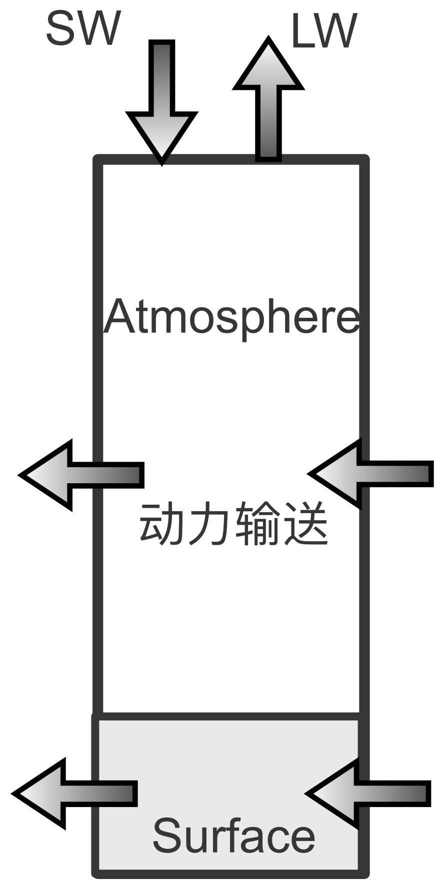
$$F_{rad}^{top} = \frac{1}{2\pi a^2 \cos\phi} \frac{\partial}{\partial \phi} f(\phi)$$

$f(\phi)$ – meridional energy transport by atmosphere and oceans





Wunsch (2005), J. Climate



Simple energy balance climate models

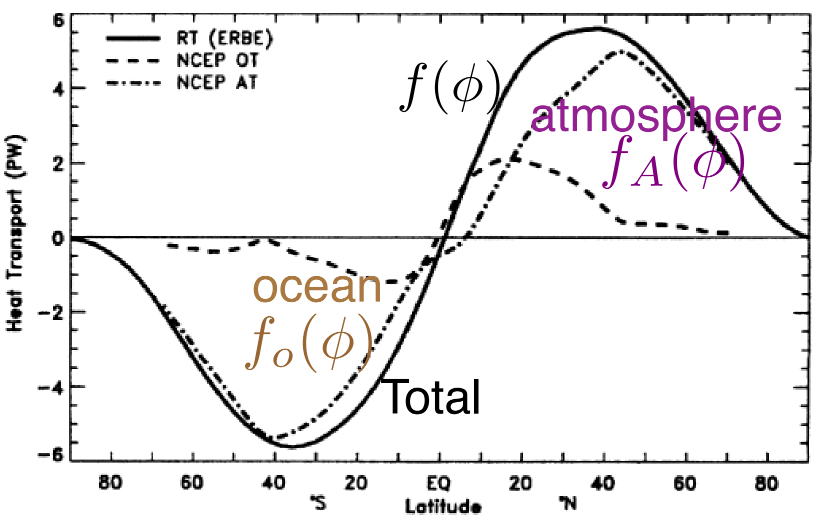
$$F_{rad}^{top} = \frac{1}{2\pi a^2 \cos\phi} \frac{\partial}{\partial \phi} f(\phi)$$

Atmosphere:

$$F_{rad}^{top} - F_{rad}^{sfc} + F_{LH} + F_{SH} = \frac{1}{2\pi a^2 \cos\phi} \frac{\partial}{\partial \phi} f_A(\phi)$$

Ocean:

$$F_{rad}^{sfc} - F_{LH} - F_{SH} = \frac{1}{2\pi a^2 \cos\phi} \frac{\partial}{\partial \phi} f_o(\phi)$$



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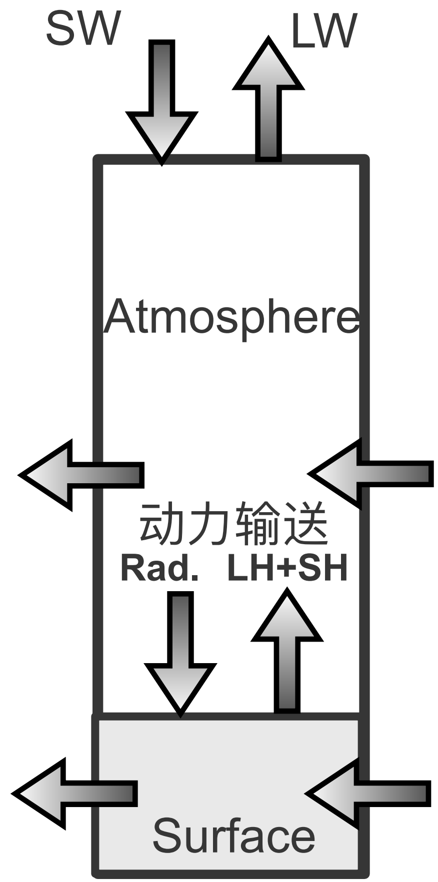
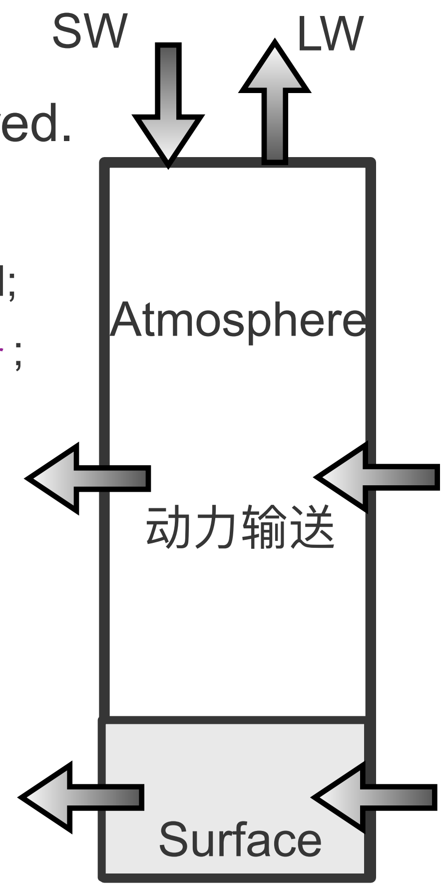


Simple energy balance climate models

- Simplest models in which the interactions between ***radiation*** and ***dynamic heat transport*** can be considered.
  - Assumptions are made below:
    - One-dimensional, only latitude dependences are considered;
    - Global energy budgets are assumed to be expressed in T_{sur} ;
    - Only annual mean conditions are considered;



$$\mathcal{C}\frac{\partial T(x,t)}{\partial t} = \text{solar radiation} - \text{infrared cooling}$$

$$-\text{divergence of heat flux}$$

$$x = \sin\phi, \text{ where } \phi \text{ is latitude.}$$

$$\mathcal{C}\frac{\partial T(x,t)}{\partial t} = F_{rad}^{top} - \frac{1}{2\pi a^2}\frac{\partial}{\partial x}f(x)$$

Simple energy balance climate models

$$\mathcal{C}\frac{\partial T(x,t)}{\partial t} = \text{solar radiation} - \text{infrared cooling}$$

$$-\text{divergence of heat flux}$$

$x = \sin\phi$, where ϕ is latitude.

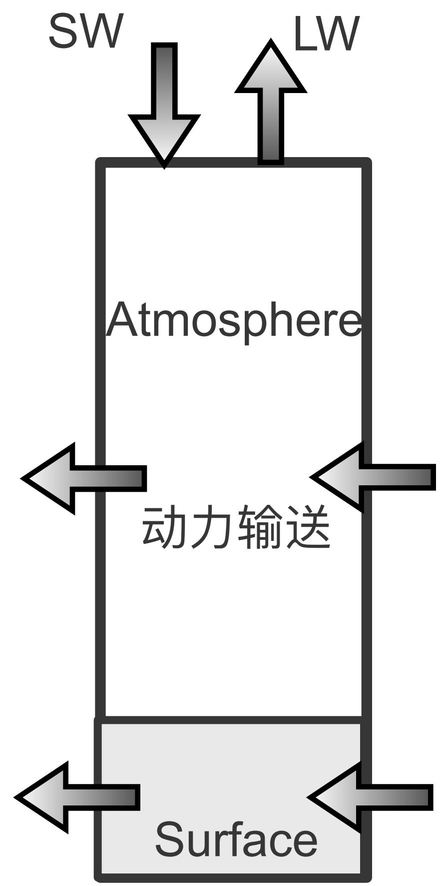
$$\text{solar radiation} = Qs(x)\mathcal{A}(T)$$

$s(x)$ – latitudinal distribution of SW, whose integral from the equator to pole is unity

$$\mathcal{C}\frac{\partial T(x,t)}{\partial t} = Qs(x)\mathcal{A}(T) - I(T) + F(T)$$

In equilibrium,

$$Qs(x)\mathcal{A}(T) - I(T) + F(T) = 0$$



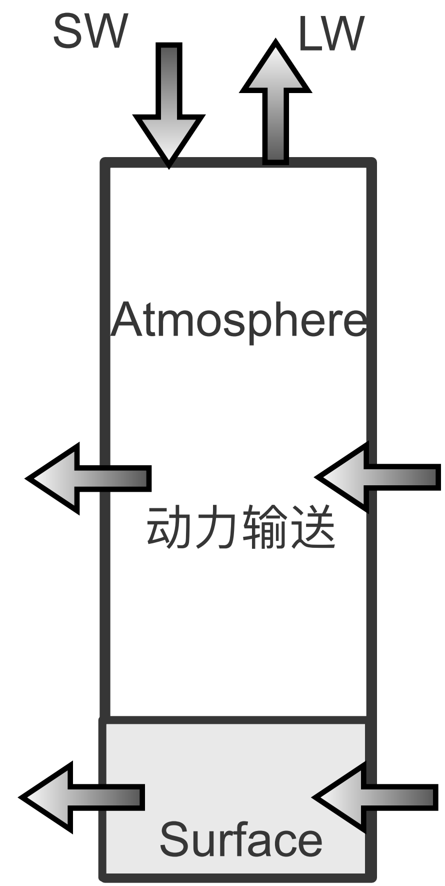
Simple energy balance climate models

In equilibrium,

$$Qs(x)\mathcal{A}(T) - I(T) + F(T) = 0$$

In real atmosphere:

$$Qs(\phi)\mathcal{A}(\phi) - I(\phi) = \frac{1}{2\pi a^2 \cos\phi}\frac{\partial}{\partial\phi}f(\phi)$$



Simple energy balance climate models

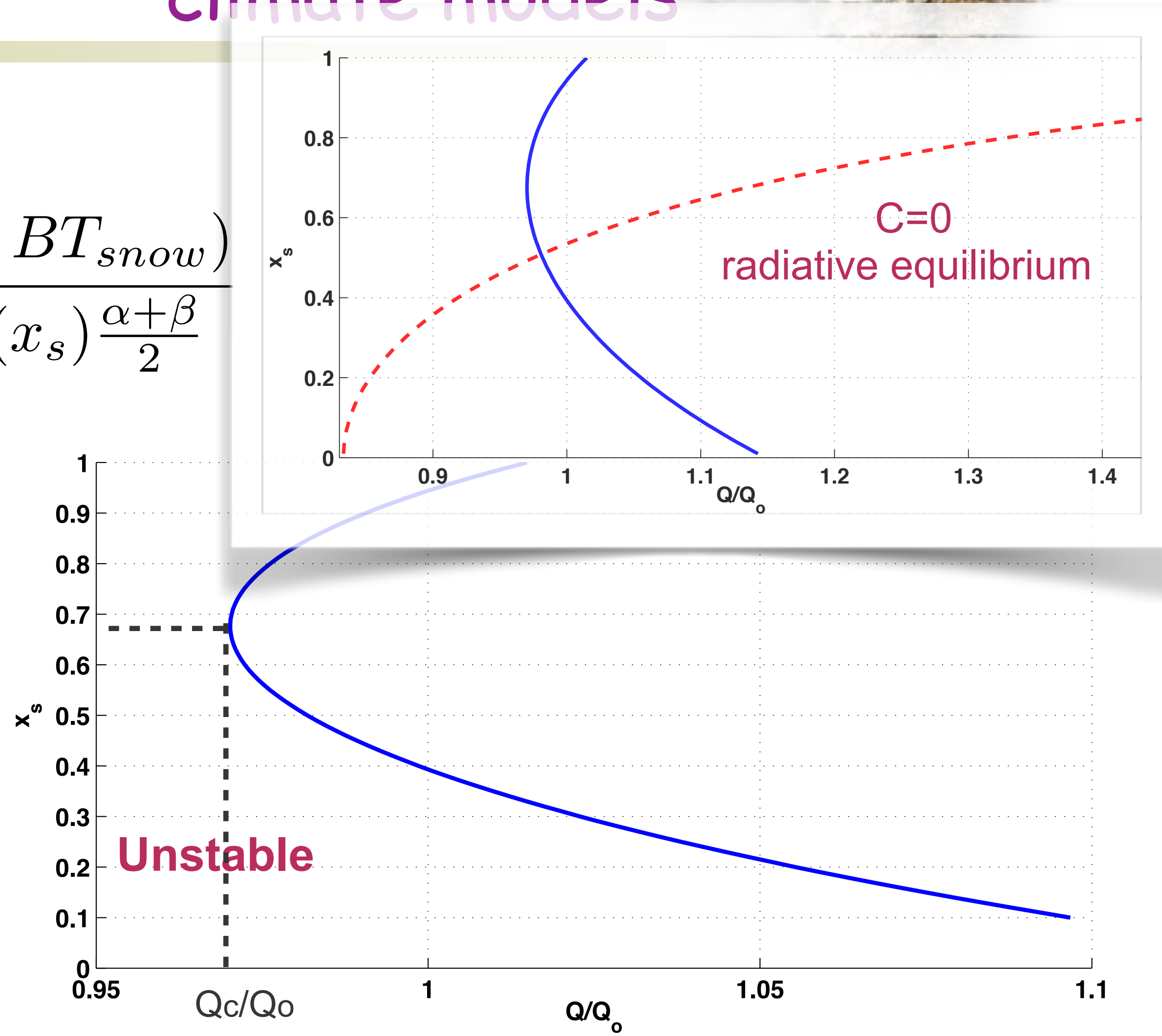
The snow line case:

$$Q(x_s) = \frac{(1+\frac{C}{B})(A+BT_{snow})}{\frac{C}{B}\bar{I}(x_s) + s(x_s)\frac{\alpha+\beta}{2}}$$

If C is nonzero,

The destabilizing effect of heat transport

There is a minimum value of Q, below which the climate will unstably proceed to a snow/ice covered earth.



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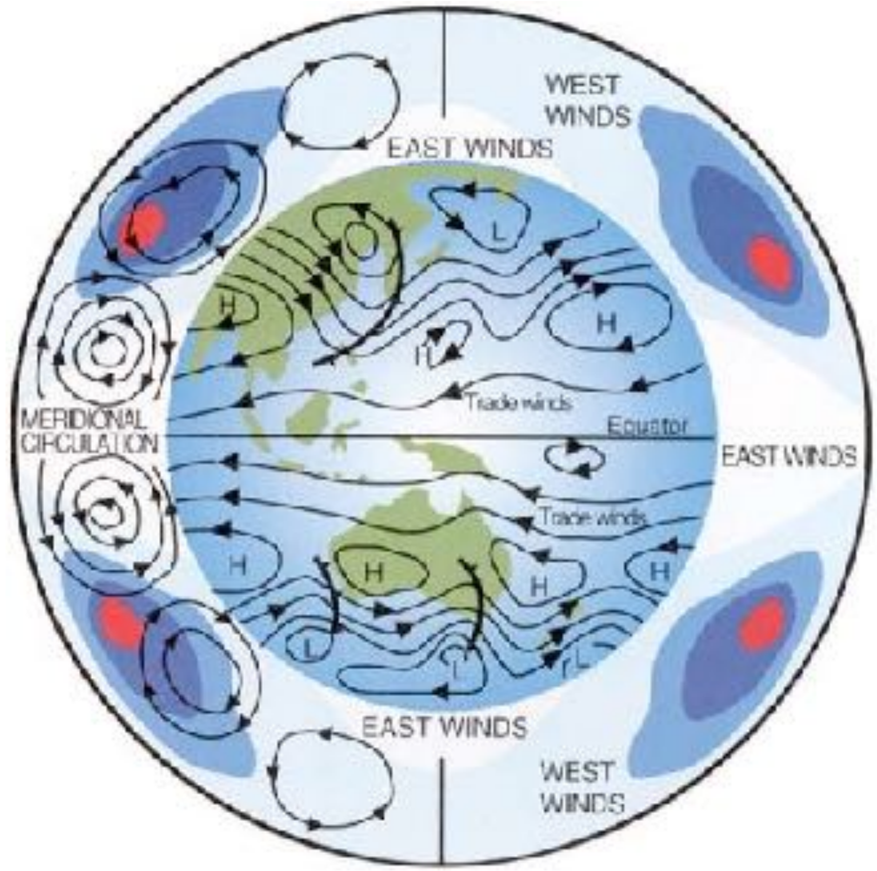


Hadley Cell

- Observations

- Summary （小结）
  - **Temperature field:** the equator-pole temperature gradient is much smaller than the RE temperature gradient.
  - **Wind fields:** **meridional winds** strongest at tropopause and surface; **vertical velocity** strongest at mid-level of the troposphere.
  - **Jets (zonal winds):** strong subtropical jet at **upper level** with its maximum in the latitudes at the edge or just poleward of the descending branch of the Hadley cell; **surface winds**-easterlies near the equator and westerlies in the extratropics.
  - **Strong seasonal variations:** in summer or winter, Hadley cell always appears as a strong single cell across the equator with the ascending branch in the tropics of the summer hemisphere.

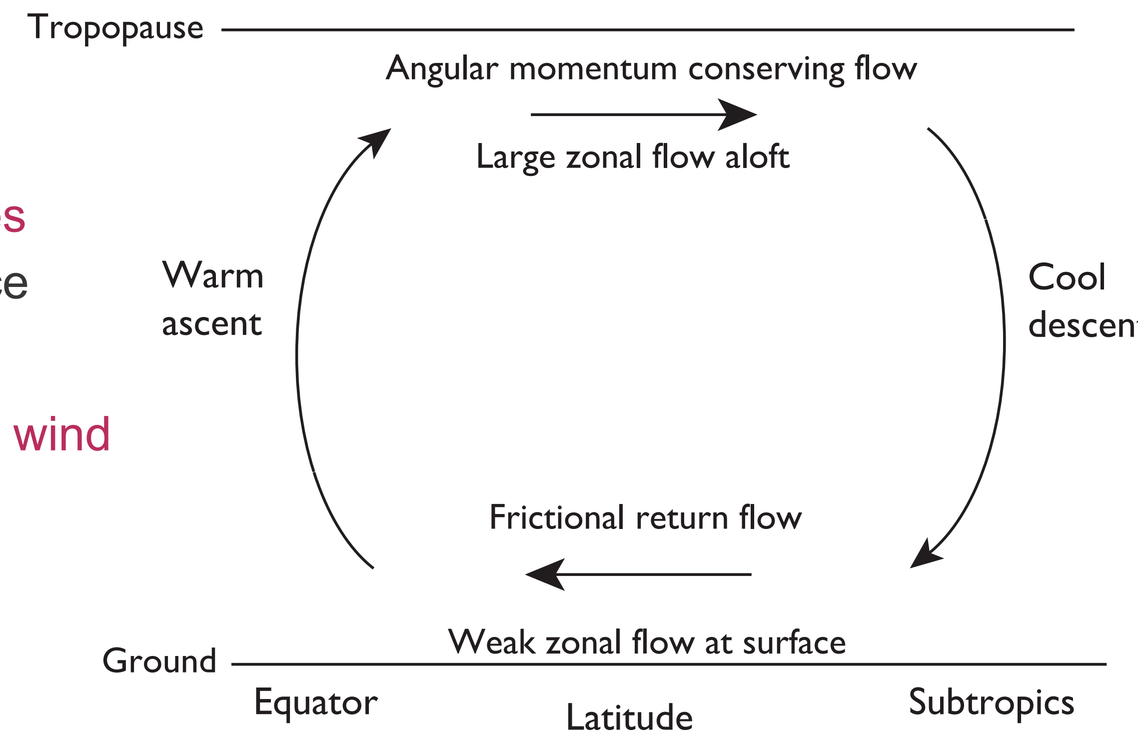
- Theories

Held-Hou model (1980)

Make assumptions:

(Vallis, 2006)

- the circulation is steady;
- the upper branch conserves angular momentum; surface zonal winds are weak;
- the circulation is in thermal wind balance.



Held-Hou model
-Angular momentum

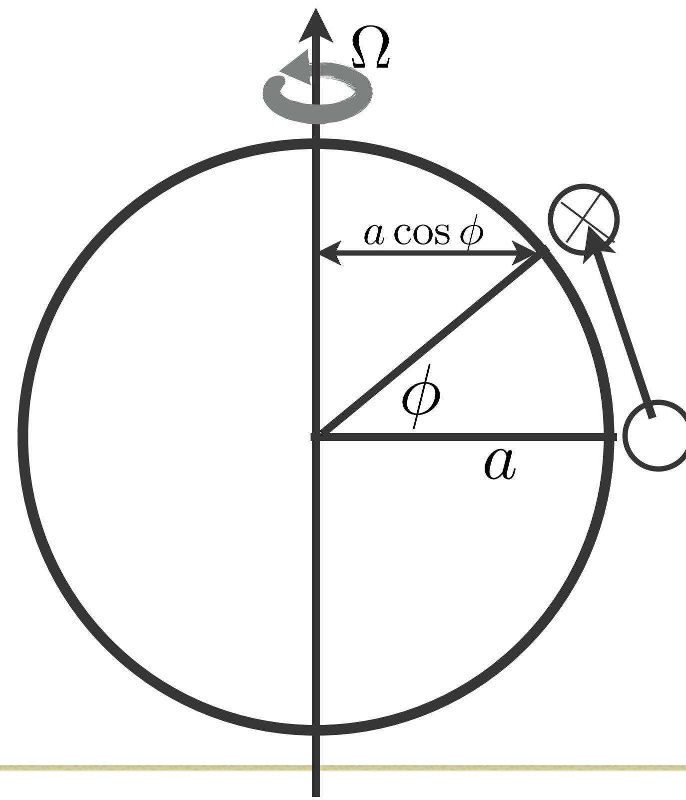
- The absolute angular momentum per unit mass is

$$M = (\Omega a \cos\phi + u) a \cos\phi$$

Due to earth's solid rotation

Deviation from the solid rotation

a is the radius of the earth



$$\frac{D}{Dt} M = -\frac{1}{\rho}\frac{\partial p}{\partial \lambda} + a\cos\phi F_\lambda$$

In an **axisymmetric flow** ([M]=M)

$$\frac{D}{Dt}[M] = a\cos\phi [F_\lambda]$$

In an inviscid (frictionless), **axisymmetric** flow, the angular momentum is conserved.

Held-Hou model

-Angular momentum

$$[M] = (\Omega a \cos\phi + [u]) a \cos\phi$$

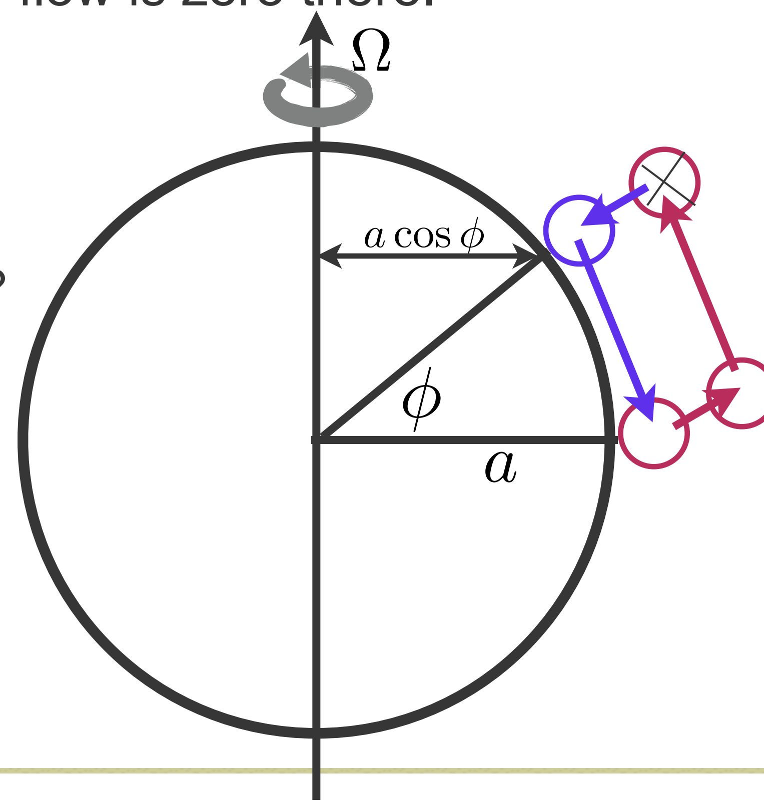
- At the equator, as the parcels rise from the surface, where the flow is weak, we assume that the zonal flow is zero there.

$$[u] = \Omega a \frac{\sin^2\phi}{\cos\phi} \equiv U_M$$

Then, what is the U_M at 10, 20, 30 degree?

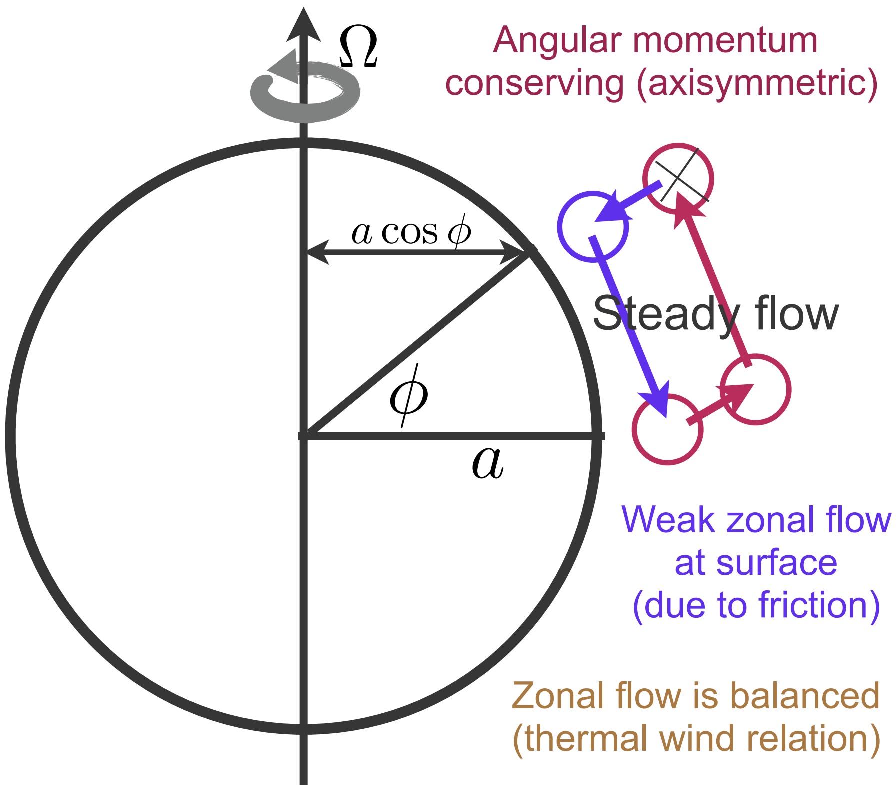
Answers: 14, 57, 134 m/s, respectively

Combined with the weak surface flow, this indicates strong vertical shear of the zonal wind.



- Theories

Held-Hou model (1980)



Meet the model (diagram)
↓
Conservation of angular momentum
↓
Thermal wind balance
↓
Distribution of temperature
↓
Latitude extent of Hadley Cell
↓
Strength of Hadley Cell
↓
Distribution of upper westerly
↓
Distribution of surface winds

Held-Hou model (review)

-Summary

- Distribution of temperature constrained by the conservation of angular momentum and thermal wind balance.

$$\frac{\tilde{\Theta}(0)-\tilde{\Theta}(\phi)}{\Theta_o}=\frac{\Omega^2 a^2}{2gH}\frac{\sin^4\phi}{\cos^2\phi}$$

Smaller than the RE temp gradient

- Extent of Hadley Cell:

$$\phi_H=\left(\frac{5}{3}\frac{gH\Delta_H}{\Omega^2 a^2}\right)^{1/2}$$

- Strength of Hadley Cell:

$$v\sim\frac{(gH)^{3/2}\,\Delta_H^{5/2}}{a^2\Omega^3\tau\Delta_V}$$

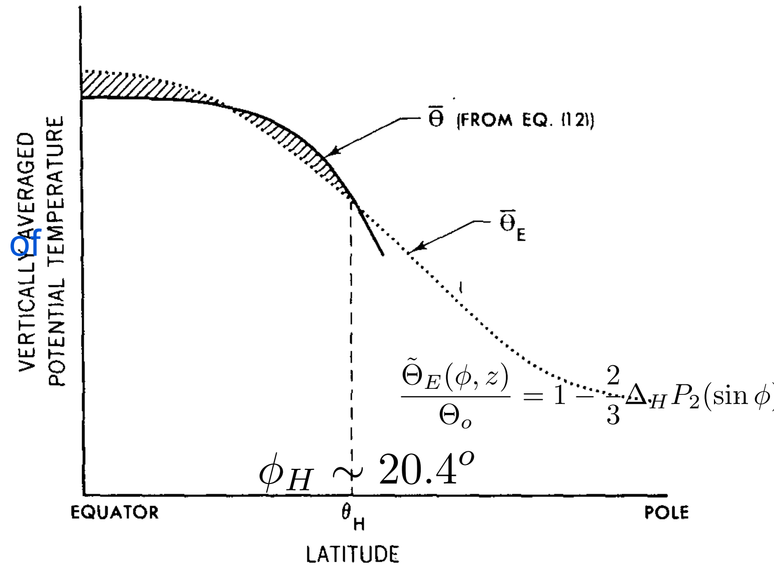
- Upper jet:

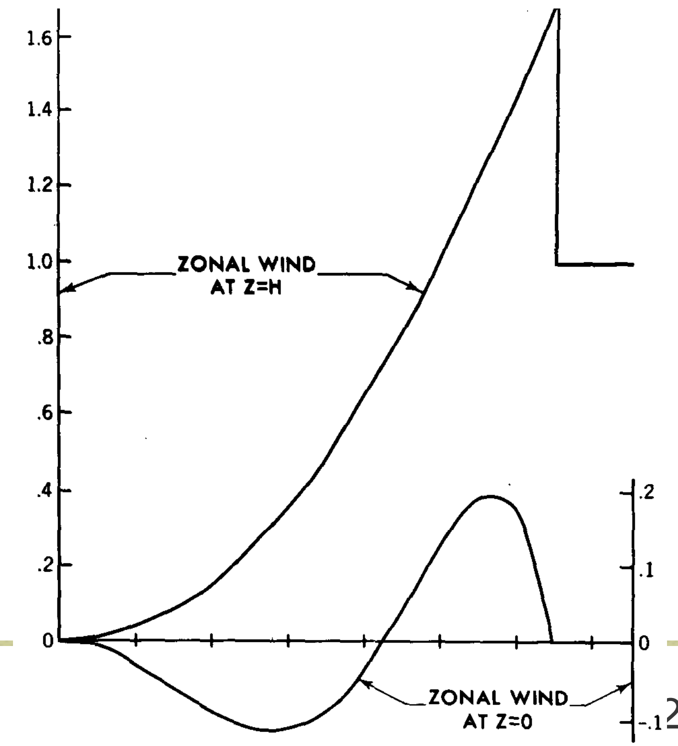
$$[u]=\Omega a\frac{\sin^2\phi}{\cos\phi}\equiv U_M$$

- Surface winds:

$$Cu(0)\approx-\frac{25}{18}\frac{g^2H^3\Delta_H^3}{a^3\Omega^3\tau\Delta_V}\left[\left(\frac{\phi}{\phi_H}\right)^2-\frac{10}{3}\left(\frac{\phi}{\phi_H}\right)^4+\frac{7}{3}\left(\frac{\phi}{\phi_H}\right)^6\right]$$

surface easterlies $\phi<\left(\frac{3}{7}\right)^{1/2}\phi_H$





第四章:

中纬度的经向环流系统

- Ferrel cell, baroclinic eddies and the westerly jet

授课教师: 张洋

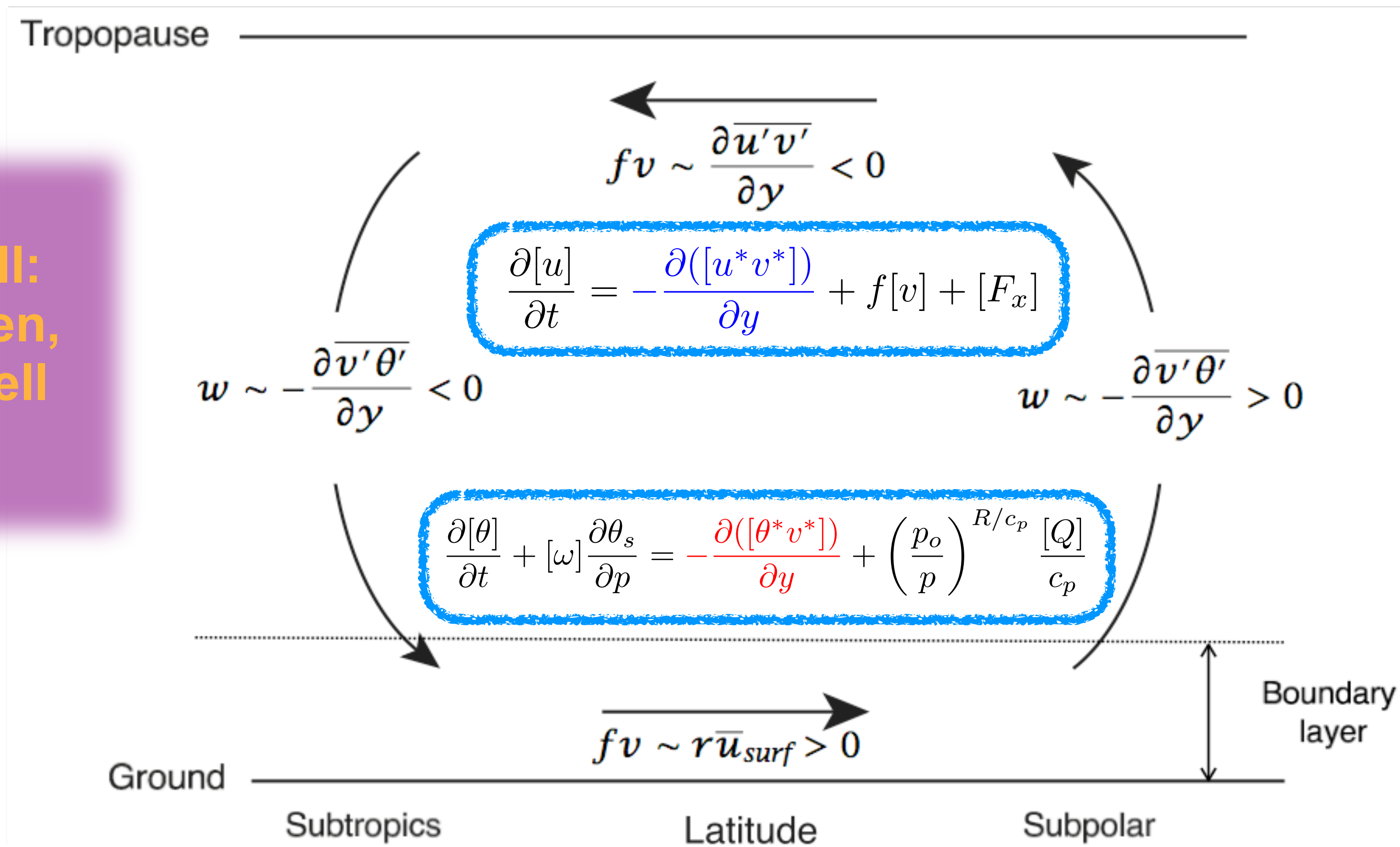
Observations

- Summary:
  - Zonal-mean flow:
    - Ferrel Cell: an indirect cell centered at 40-60 degree, with strong seasonal variation in N.H.
    - Westerly jet: surface westerlies centered at 40-60 degree
  - Eddies: transient eddies are dominant with stationary eddies only obvious in N.H.
    - Kinetic energy
    - Momentum flux
    - (Sensible) Heat flux

The Ferrel Cell

- The balance equations:

Ferrel Cell: eddy-driven, indirect cell



Baroclinic eddies

- linear baroclinic instability

- Conclusions:

Necessary condition for baroclinic instability: PV gradient changes sign in the interior or boundaries (Charney-stern theory), according to which the midlatitude atmosphere is baroclinic unstable. Different models. i.e. Eady and Charney models have more rigorous conditions.

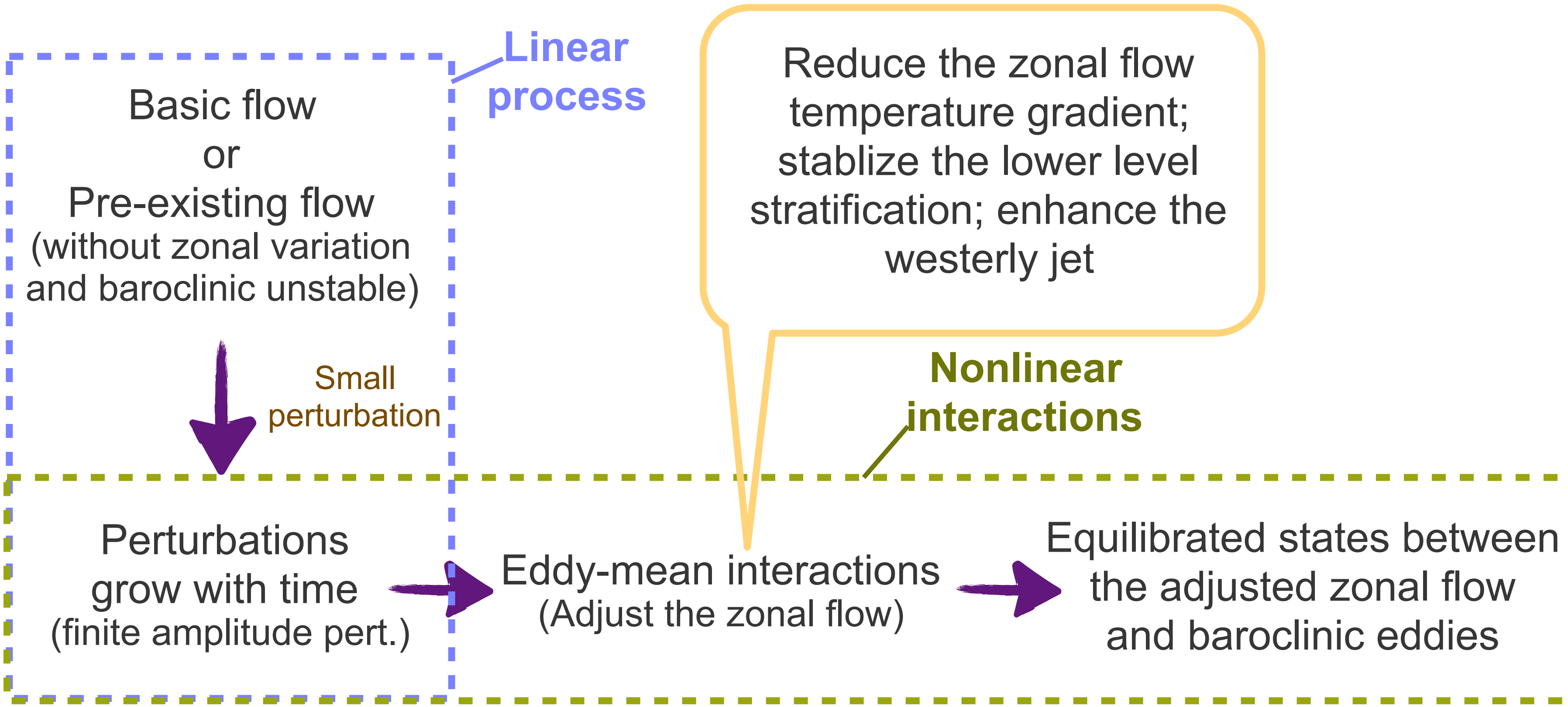
Growth rate: $\sigma = kc_i \approx 0.3\,\Lambda \dfrac{f_o}{N}$ in both Eady and Charney models!

Eady number/Eady growth rate

Most unstable mode: $k_{\max}^{-1} \propto L_d^{-1} = \left(\dfrac{NH}{f_o}\right)^{-1}$ (Eady) $\qquad k_{\max}^{-1} \propto \Lambda \dfrac{f_o}{\beta N}$ (Charney)

Baroclinic eddies

- From linear to nonlinear



Baroclinic eddies

- E-P flux

- In a QG, **steady**, **adiabatic** and **frictionless** flow:

- Momentum equation:

$$f[v] - \frac{\partial([u^*v^*])}{\partial y} = 0$$

- Continuity equation:

$$\frac{\partial[v]}{\partial y} + \frac{\partial[\omega]}{\partial p} = 0 \rightarrow \nabla \cdot \mathcal{F} = 0$$

- Thermodynamic equation:

$$[\omega]\frac{\partial\theta_s}{\partial p} + \frac{\partial([\theta^*v^*])}{\partial y} = 0$$

$$[v] = \frac{1}{f}\frac{\partial}{\partial y}([u^*v^*])$$

$$[\omega] = -\frac{\partial}{\partial y}\left(\frac{[\theta^*v^*]}{\partial\theta_s/\partial p}\right)$$

Define *Eliassen-Palm flux:*

$$\mathcal{F} \equiv -[u^*v^*]\,\mathbf{j} + f\,\frac{[v^*\theta^*]}{\partial\theta_s/\partial p}\,\mathbf{k}$$

E-P flux, TEM and Residual Circulation - Summary

- E-P flux: $$\mathcal{F} \equiv -[u^*v^*]\,\mathbf{j} + f\frac{[v^*\theta^*]}{\partial\theta_s/\partial p}\,\mathbf{k}$$
- In a **steady**, **adiabatic** and **frictionless** flow:

$$[v] = \frac{1}{f}\frac{\partial}{\partial y}([u^*v^*]) \qquad [\omega] = -\frac{\partial}{\partial y}\left(\frac{[\theta^*v^*]}{\partial\theta_s/\partial p}\right) \qquad \nabla\cdot\mathcal{F} = 0$$

#1
$$[v^*q^*] = -\frac{\partial}{\partial y}[u^*v^*] + f_o\frac{\partial}{\partial p}\frac{[v^*\theta^*]}{\partial\theta_s/\partial p} = \nabla\cdot\mathcal{F}$$

#2
$$\frac{\partial\mathcal{A}}{\partial t} + \nabla\cdot\mathcal{F} = 0$$

#3
$$\vec{\mathcal{F}} = \vec{\mathbf{c_g}}\mathcal{A} \;\longrightarrow\; \frac{\partial\mathcal{A}}{\partial t} + \nabla\cdot(\mathcal{A}\vec{\mathbf{c_g}}) = 0$$

- Residual mean circulations:

$$[\tilde{\omega}] = [\omega] + \frac{\partial}{\partial y}\left(\frac{[v^*\theta^*]}{\partial\theta_s/\partial p}\right), \qquad [\tilde{v}] = [v] - \frac{\partial}{\partial p}\left(\frac{[v^*\theta^*]}{\partial\theta_s/\partial p}\right)$$

- TEM equations: $$\frac{\partial[u]}{\partial t} = f[\tilde{v}] + \nabla\cdot\mathcal{F} + [F_x], \qquad \frac{\partial[\theta]}{\partial t} = -[\tilde{\omega}]\frac{\partial\theta_s}{\partial p} + \left(\frac{p_o}{p}\right)^{R/c_p}\frac{[Q]}{c_p}$$

Baroclinic eddies

- TEM

$$\tilde{\psi} = \psi_m + \frac{[v^*\theta^*]}{\partial\theta_s/\partial p}$$

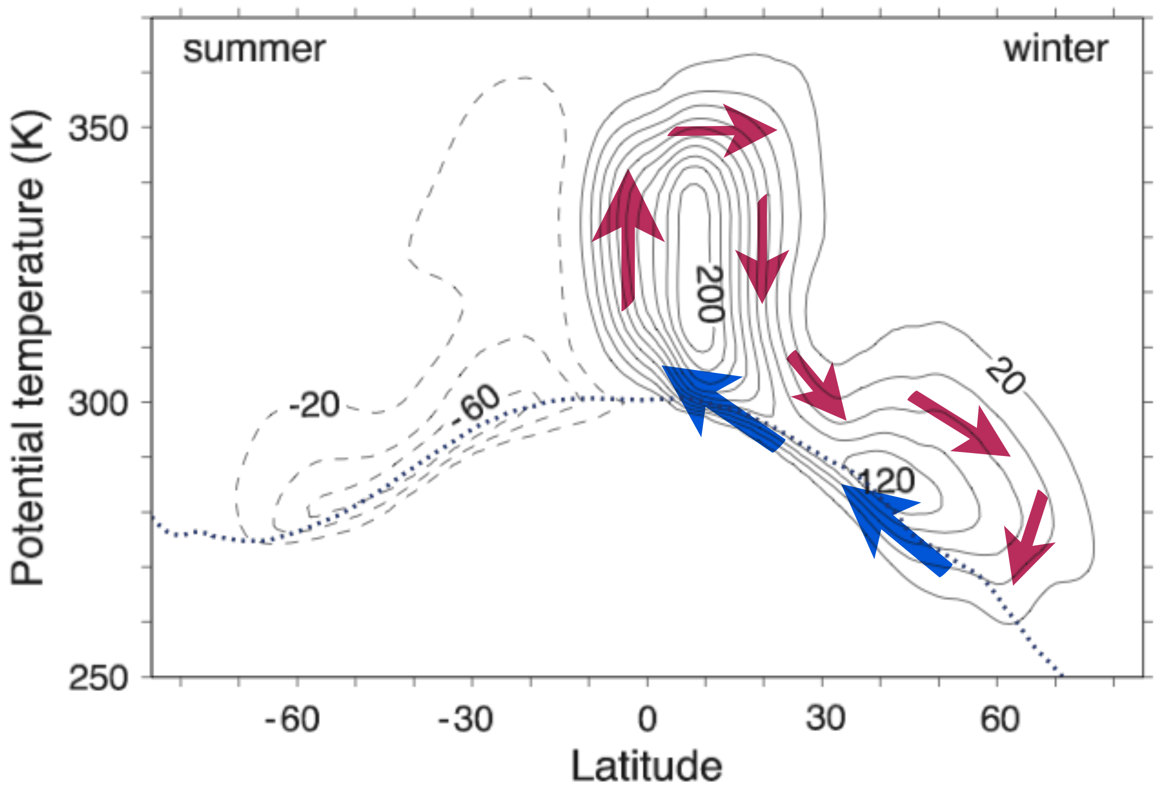
- In **isentropic** coordinate

$$\frac{D}{Dt} = \frac{\partial}{\partial t} + \mathbf{u}\cdot\nabla_\theta + \frac{D\theta}{Dt}\frac{\partial}{\partial\theta}$$
$$= \frac{\partial}{\partial t} + \mathbf{u}\cdot\nabla_\theta + \dot{\theta}\frac{\partial}{\partial\theta}$$

zero for adiabatic flow

The Ferrel cell in the isentropic coordinate is essentially reflect the *Residual Mean Circulation*.

Case 2: Observed circulation



(Fig.11.4, Vallis, 2006)

E-P flux

- The westerly jet

$$\mathcal{F} \equiv -[u^*v^*]\,\mathbf{j} + f\,\frac{[v^*\theta^*]}{\partial\theta_s/\partial p}\,\mathbf{k}$$

$$\vec{\mathcal{F}} = \vec{\mathbf{c_g}}\mathcal{A}$$

$$\frac{\partial[u]}{\partial t} = f[\tilde{v}] + \nabla\cdot\mathcal{F} + [F_x]$$

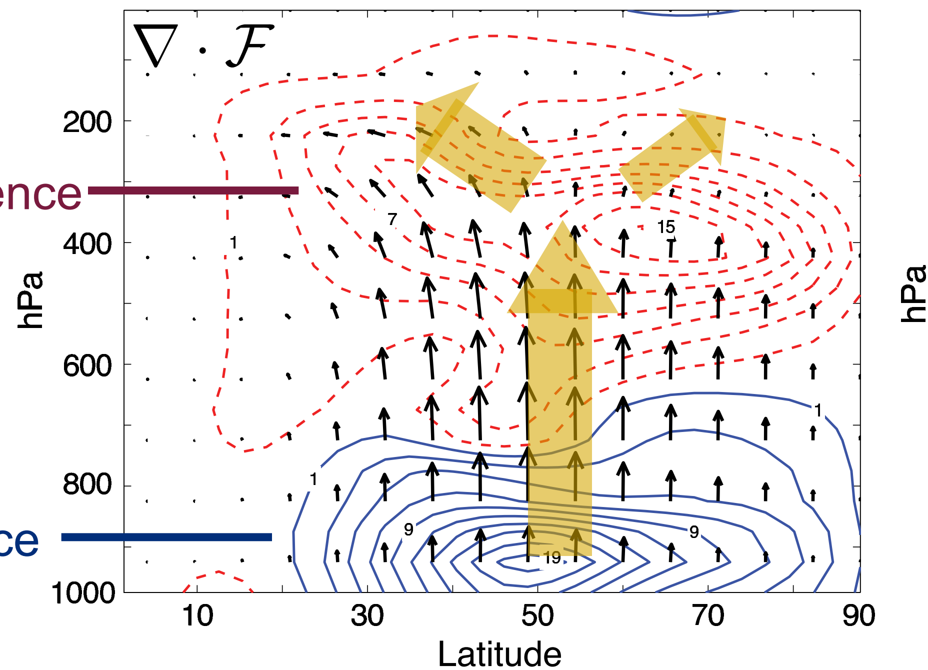
Wave energies: propagate **upwards** and **away** from the center of the jet

In the vertical direction:

Accelerating the lower jet
decelerating the upper jet
reduce the vertical shear of U

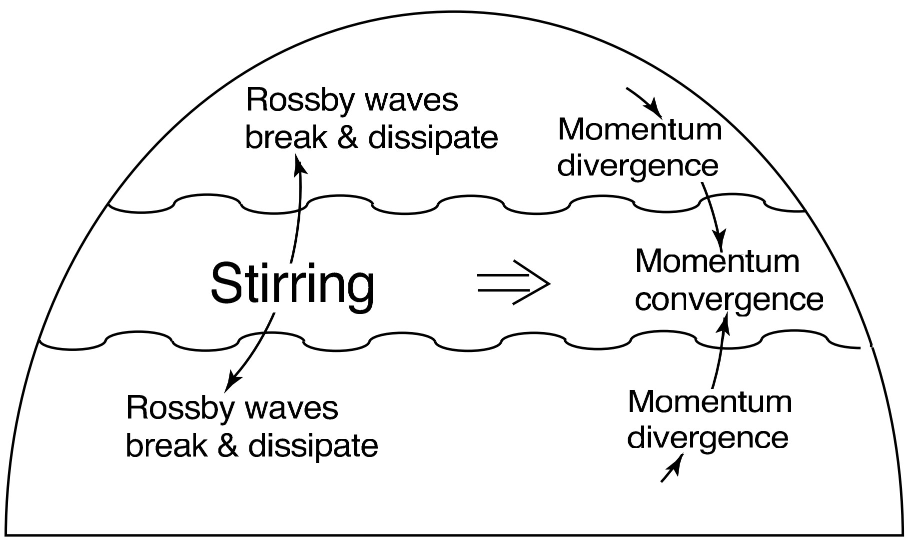
Convergence

Divergence



Eddy-driven jet:
- the momentum budget

$$\mathcal{F} \equiv -[u^* v^*]\,\mathbf{j} + f\frac{[v^*\theta^*]}{\partial\theta_s/\partial p}\,\mathbf{k}$$



$$\frac{\partial}{\partial t} < [u] >= -\frac{\partial}{\partial y} < [u^* v^*] > -r[u_{\rm surf}]$$

< > means vertical average

Wave energies: propagate **upwards** and **away** from the center of the jet

$$\vec{\mathcal{F}} = \vec{\mathbf{c_g}}\mathcal{A}$$

In equilibrium:

$$r[u_{\rm surf}] \sim -\frac{\partial}{\partial y} < [u^* v^*] >$$

There MUST be **surface westerlies** at midlatitudes.

E-P flux and the eddy-driven jet

-summary

$$\mathcal{F} \equiv -[u^* v^*]\,\mathbf{j} + f\,\frac{[v^*\theta^*]}{\partial\theta_s/\partial p}\,\mathbf{k}$$

$$\vec{\mathcal{F}} = \vec{\mathbf{c_g}}\mathcal{A}$$

- Numerical results and observations: eddies generate in the lower level, propagate **upwards** and **away** from the eddy source region.

$$\frac{\partial[u]}{\partial t} = f[\tilde{v}] + \nabla\cdot\mathcal{F} + [F_x]$$

- Accelerating the lower jet, decelerating the upper jet, reduce the vertical shear of U

- Momentum budget indicates that there MUST be surface westerlies in the eddy source latitude.

Energy cycles
in the baroclinic eddy-mean flow interactions

- Zonal mean and eddy components:
  - Kinetic energy (动能): $K_{\mathrm{M}} = \frac{1}{2}\left([u]^2 + [v]^2\right) \qquad K_{\mathrm{E}} = \frac{1}{2}\left([u^{*2}] + [v^{*2}]\right)$
  - Available potential energy (有效位能):

$$P_{\mathrm{M}} = \frac{c_p}{2}\Gamma\left([T] - T_s\right)^2 \qquad P_{\mathrm{E}} = \frac{c_p}{2}\Gamma[T^{*2}]$$

- Tendency equations under the QG assumption:

$$\frac{\partial}{\partial t}\int K \mathrm{d}m = -R\int \frac{\omega T}{p}\mathrm{d}m + \int (uF_x + vF_y)\mathrm{d}m$$

$$\frac{\partial}{\partial t}\int P \mathrm{d}m = R\int \frac{\omega T}{p}\mathrm{d}m + \int \Gamma(T - T_s)(Q - Q_s)\mathrm{d}m$$

Q - diabatic heating

Energy cycles

in the baroclinic eddy-mean flow interactions

- Equations under the **Quasi-geostrophic** assumption:

$$\frac{\partial}{\partial t}\int K\mathrm{d}m = -R\int\frac{\omega T}{p}\mathrm{d}m + \int(uF_x + vF_y)\mathrm{d}m$$

$$\frac{\partial}{\partial t}\int K_{\mathrm{M}}\mathrm{d}m = -R\int\frac{[\omega][T]}{p}\mathrm{d}m + \int[u^*v^*]\frac{\partial[u]}{\partial y}\mathrm{d}m + \int([u][F_x] + [v][F_y])\mathrm{d}m$$

$$\frac{\partial}{\partial t}\int K_{\mathrm{E}}\mathrm{d}m = -R\int\frac{[\omega^*T^*]}{p}\mathrm{d}m - \int[u^*v^*]\frac{\partial[u]}{\partial y}\mathrm{d}m + \int([u^*F_x^* + v^*F_y^*])\mathrm{d}m$$

$$\frac{\partial}{\partial t}\int P\mathrm{d}m = R\int\frac{\omega T}{p}\mathrm{d}m + \int\Gamma(T - T_s)(Q - Q_s)\mathrm{d}m$$

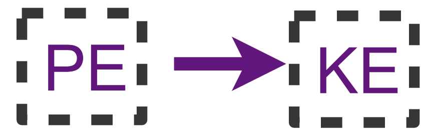
$$\frac{\partial}{\partial t}\int P_{\mathrm{M}}\mathrm{d}m = R\int\frac{[\omega][T]}{p}\mathrm{d}m + c_p\int\Gamma[v^*T^*]\frac{\partial[T]}{\partial y}\mathrm{d}m + \int\Gamma([T] - T_s)([Q] - Q_s)\mathrm{d}m$$

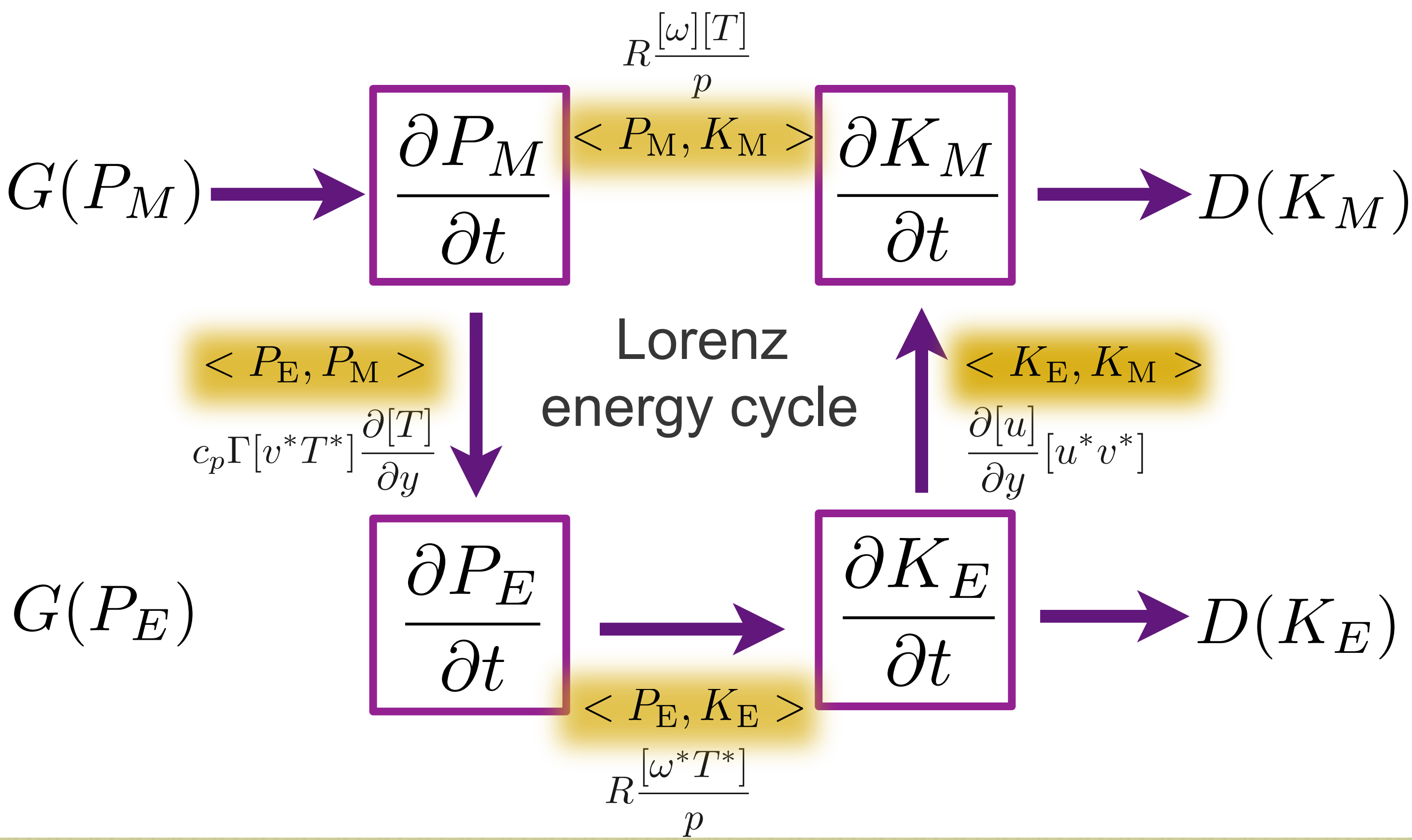
$$\frac{\partial}{\partial t}\int P_{\mathrm{E}}\mathrm{d}m = R\int\frac{[\omega^*T^*]}{p}\mathrm{d}m - c_p\int\Gamma[v^*T^*]\frac{\partial[T]}{\partial y}\mathrm{d}m + \int\Gamma[T^*Q^*]\mathrm{d}m$$

Energy cycles

in the baroclinic eddy-mean flow interactions

- Lorenz energy cycle:

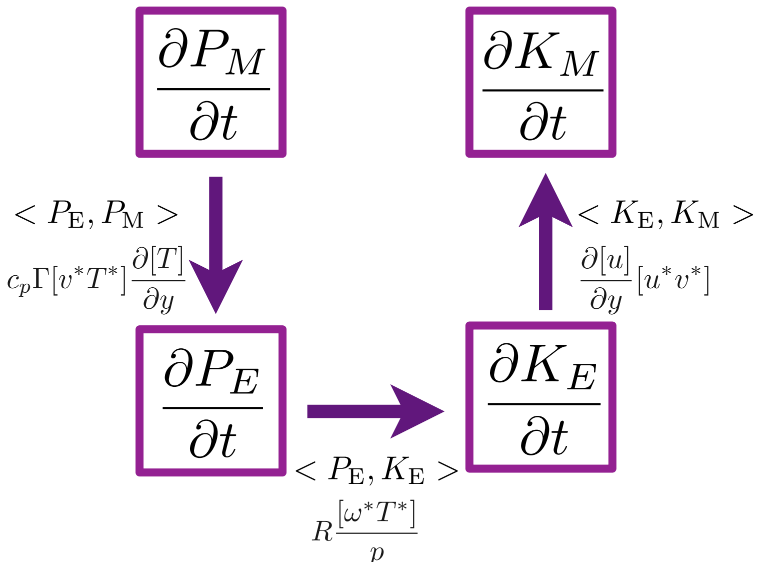


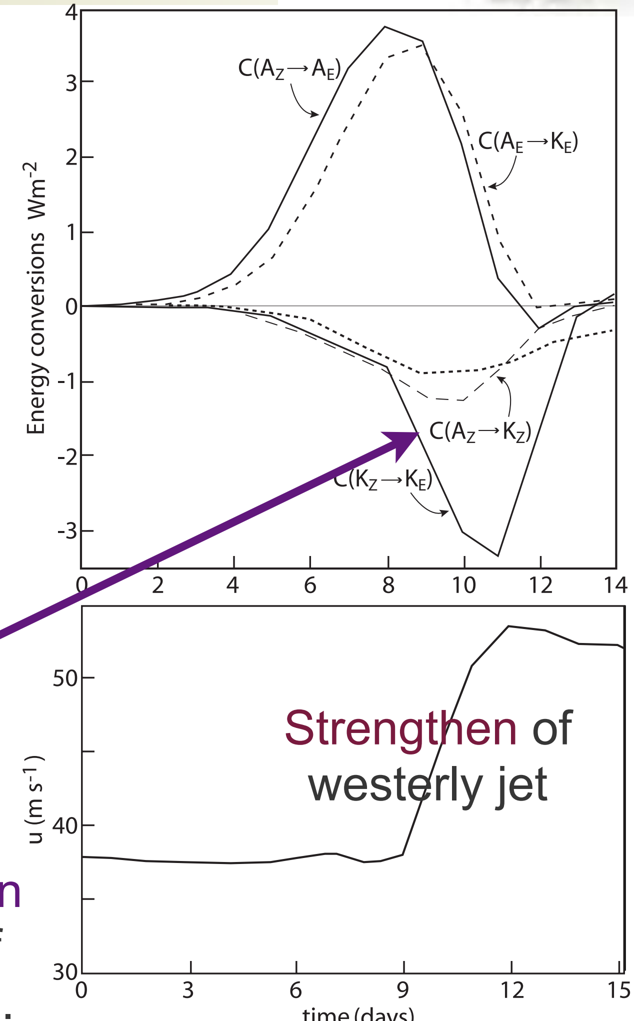


Baroclinic eddies

- baroclinic eddy life cycle

- Westerly jet and energy cycle:





Eddy momentum flux grows, which extracts kinetic energy from the eddies to the zonal mean flow, then the growth of the eddy energy ceases.

Numerical results from
Simmons and Hoskins, 1978, JAS

第六章:

能量与水汽的
分布、平衡与输送

授课教师：张洋

Distribution of each component

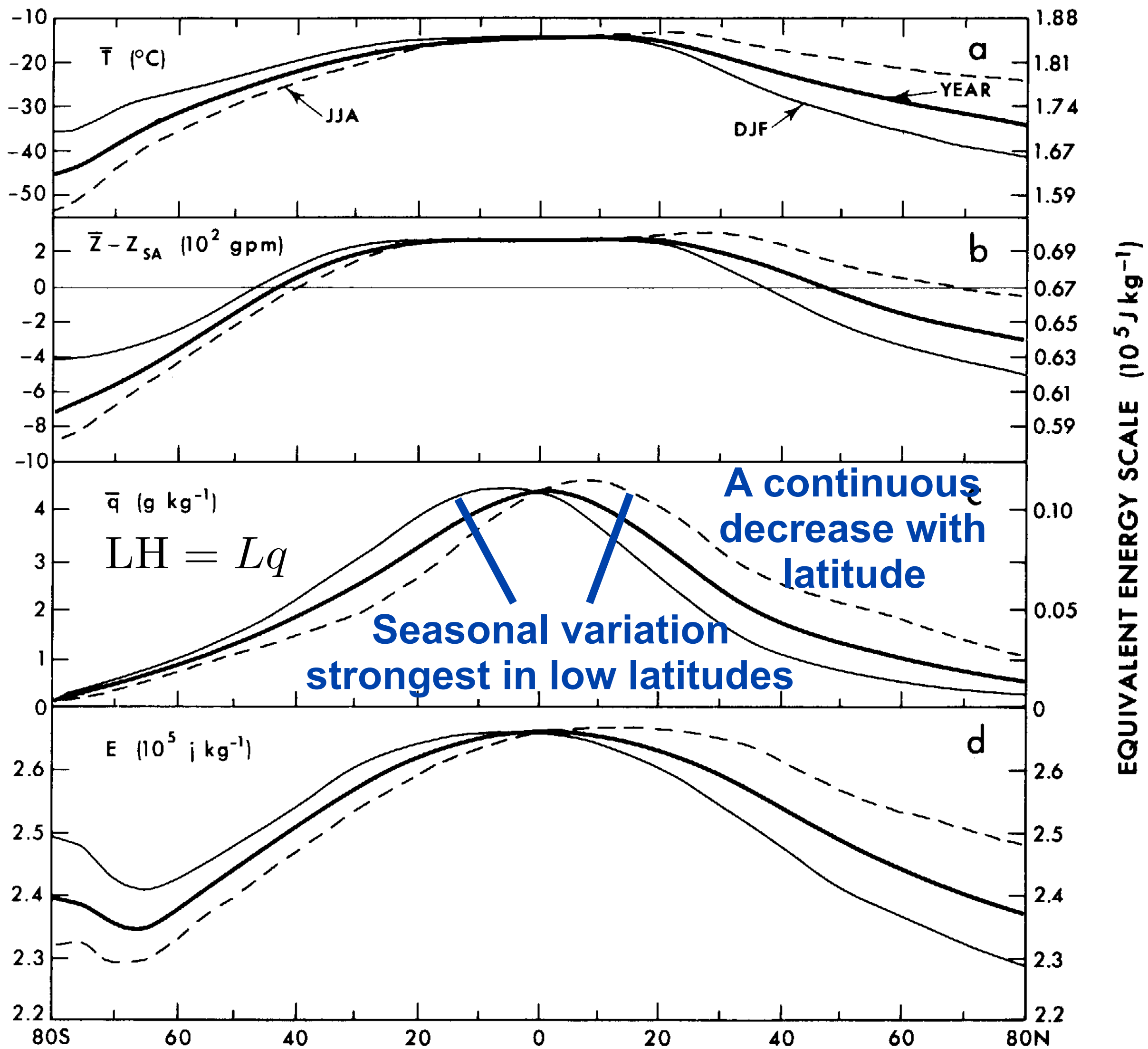
- Total energy:

$$E = I + \Phi + \mathrm{LH} + K$$

70.4% 27.1% 2.5% 0.05%

However, only **0.5%** are ***available*** to be converted for the general circulation.

- Meridional distribution



(Vertical averaged; PO

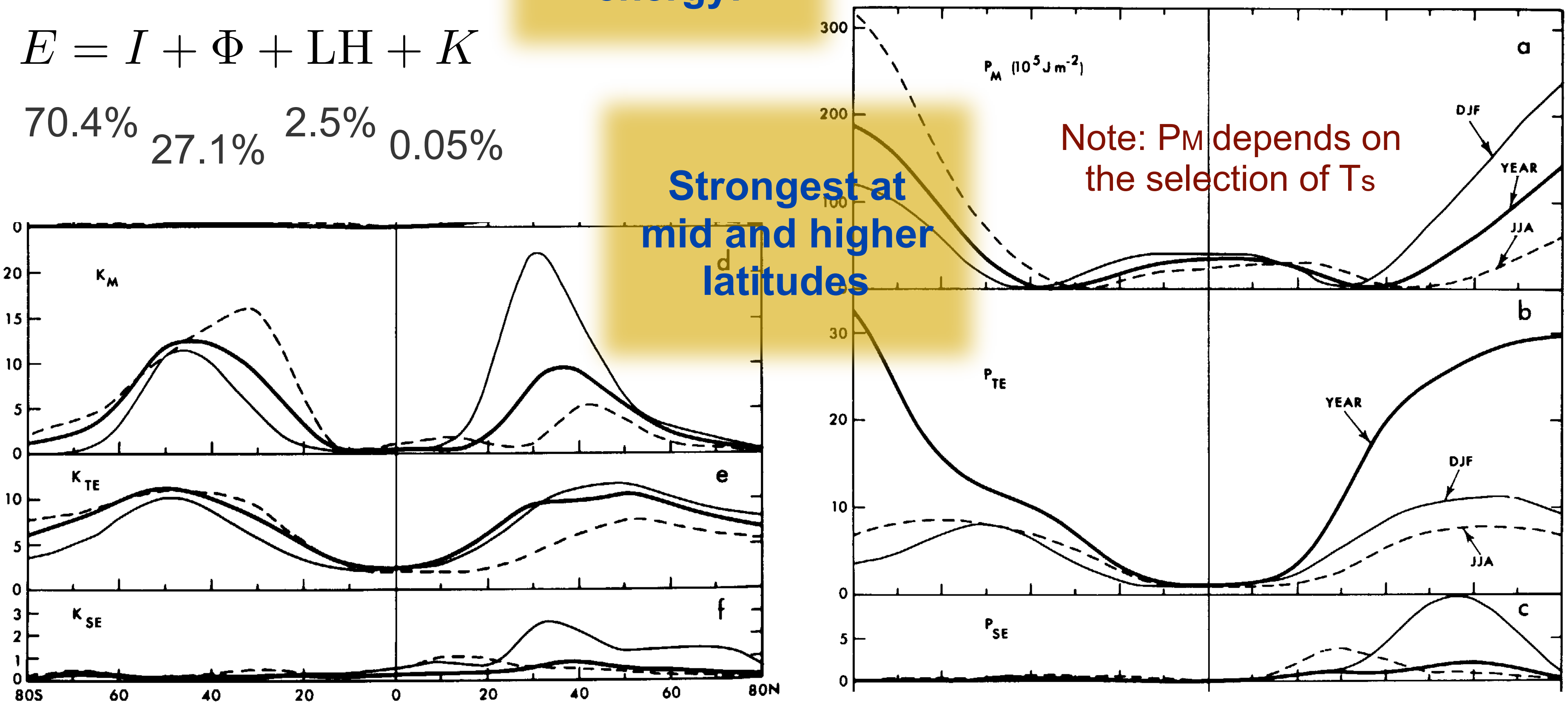
Distribution of each component

- Total energy:

$$E = I + \Phi + \mathrm{LH} + K$$

70.4% 27.1% 2.5% 0.05%

The available energy:



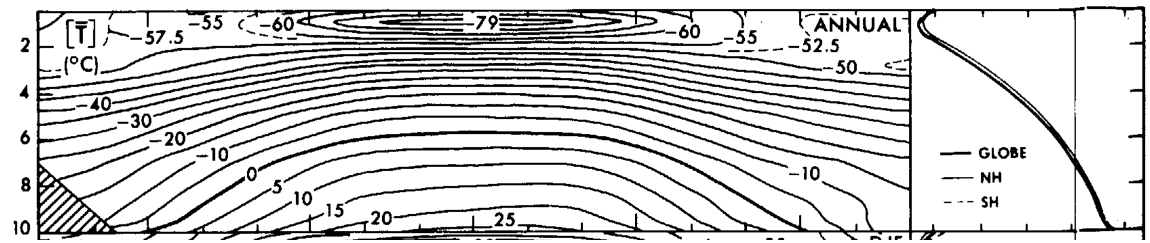
Strongest at mid and higher latitudes

Note: P_M depends on the selection of T_s

Distribution of each component

- Total energy:

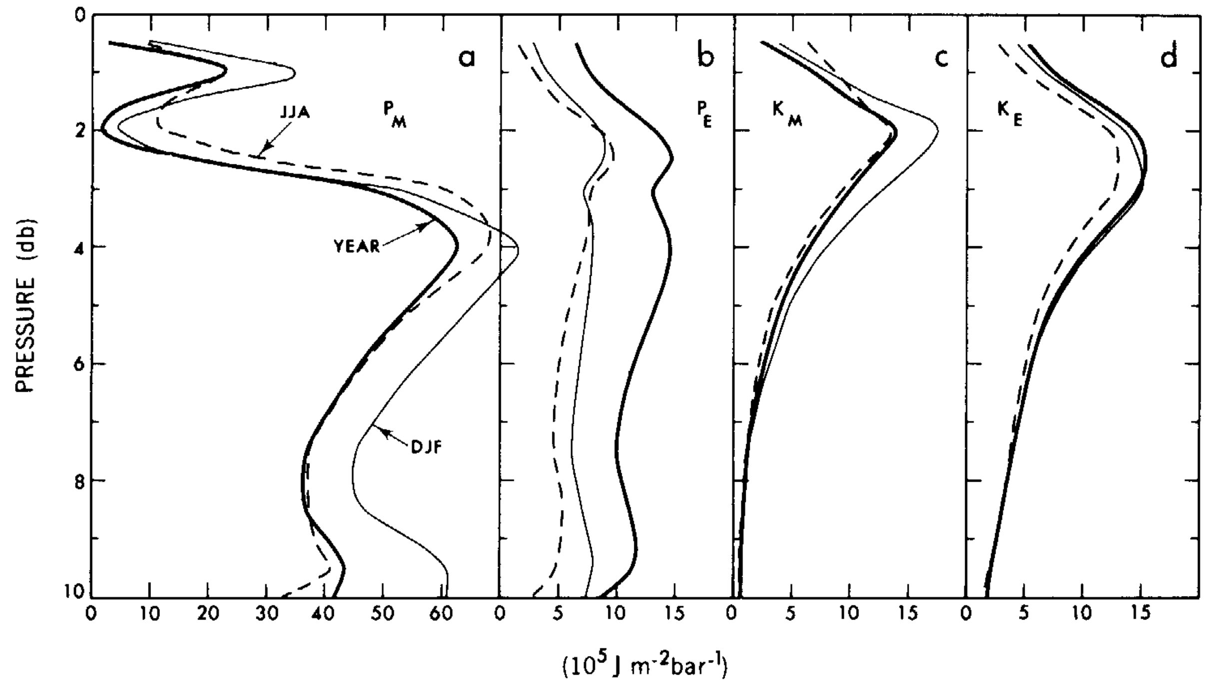
$$E = I + \Phi + \mathrm{LH} + K$$



- Vertical distribution:

Total energy:
decreases with height.

The available energy:
PE: peaks near tropopause and surface;
KE: strongest at tropopause.

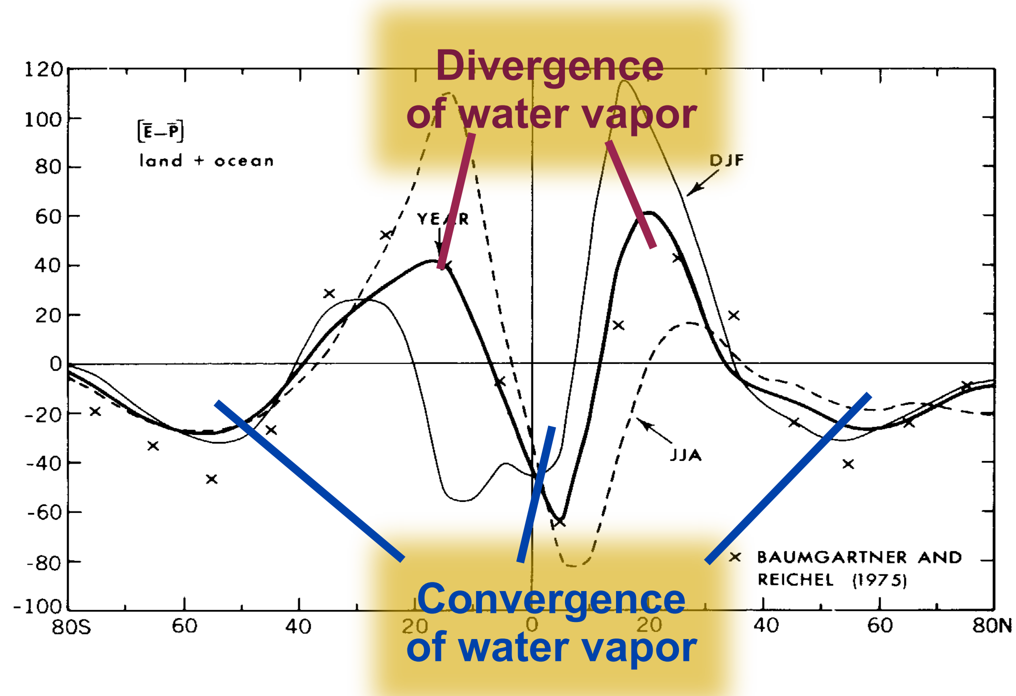


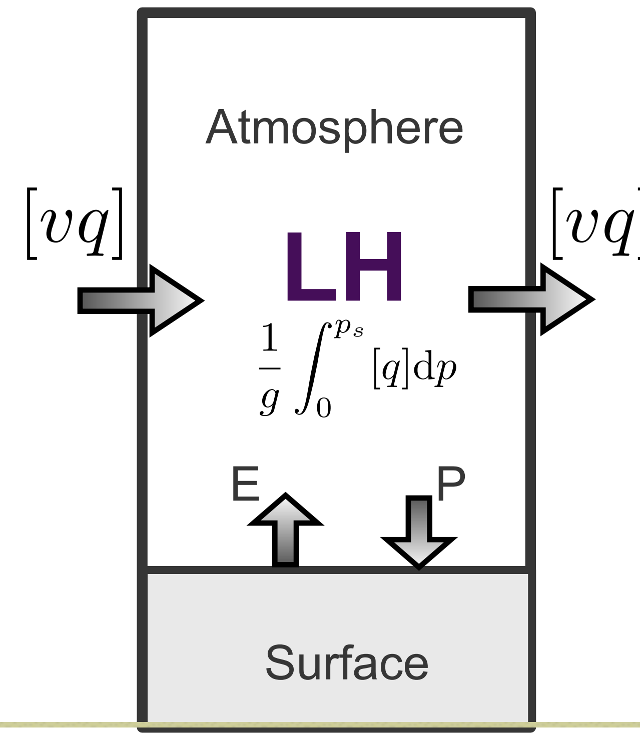
The budget equation of water vapor

$$\left(\frac{dq}{dt}\right)_p = s(q) + D \qquad s(q) = e - c$$

Integrate above equation vertically and over a latitudinal belt:

$$\frac{\partial}{\partial t}\int_0^{p_s}[q]\frac{\mathrm{d}p}{g} = -\frac{\partial}{\partial y}\int_0^{p_s}[vq]\frac{\mathrm{d}p}{g} + [E-P]$$



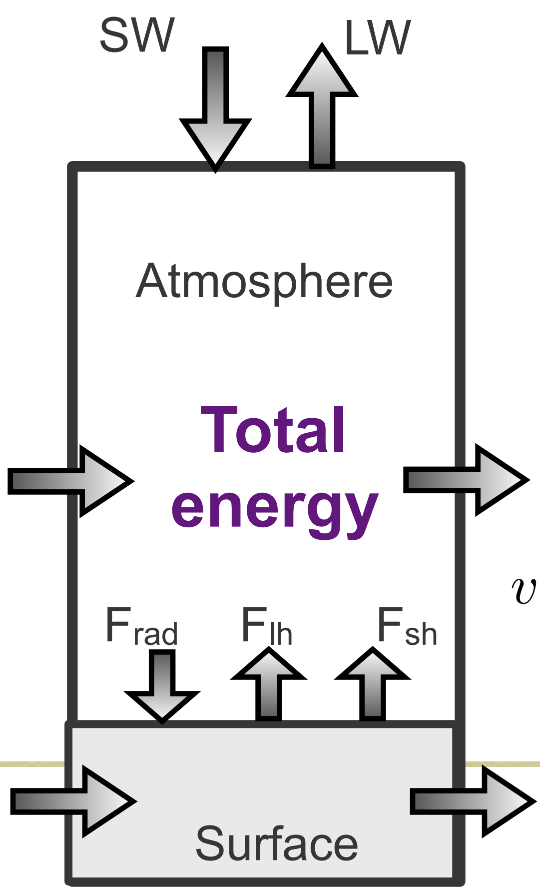


The energy budget

From the momentum, thermodynamic equation and the water vapor budget:

$$Q_{\mathrm{RAD}} + Q_{\mathrm{B}} \approx -g\frac{\partial}{\partial p}(F_{\mathrm{rad}} + F_{\mathrm{sh}})$$

$$\frac{\partial}{\partial t}\int_0^{p_s}(c_v T + gz + Lq + K)\frac{\mathrm{d}p}{g} = -\int_0^{p_s}\nabla\cdot\mathbf{v}\,(c_p T + gz + Lq + K)\frac{\mathrm{d}p}{g} + \int_0^{p_s}(Q + uF_x + vF_y)\frac{\mathrm{d}p}{g}$$



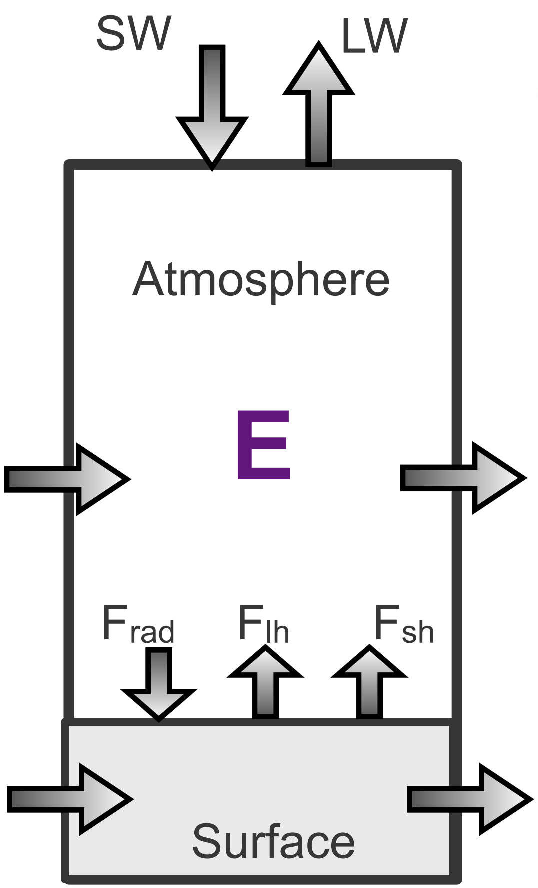
After simplification and zonal average:

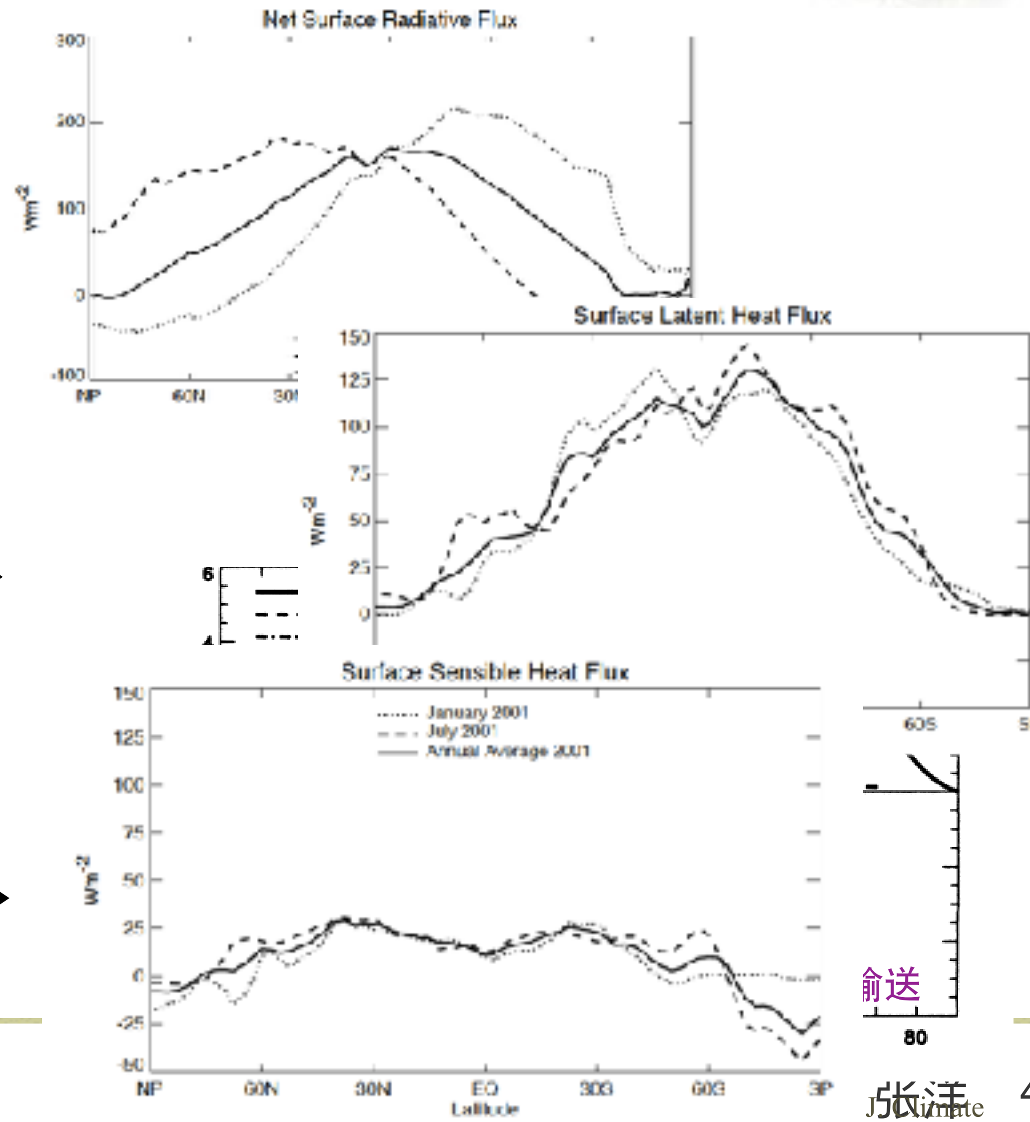
$$\frac{\partial}{\partial t}\int_0^{p_s}[c_p T + Lq + K]\frac{\mathrm{d}p}{g} = -\frac{\partial}{\partial y}\int_0^{p_s}[v(c_p T + gz + Lq + K)] + [F_{\mathrm{rad}}]^{\mathrm{top}} - [F_{\mathrm{rad}} + F_{\mathrm{sh}} + F_{lh}]^{\mathrm{surf}}$$

Energy transport:

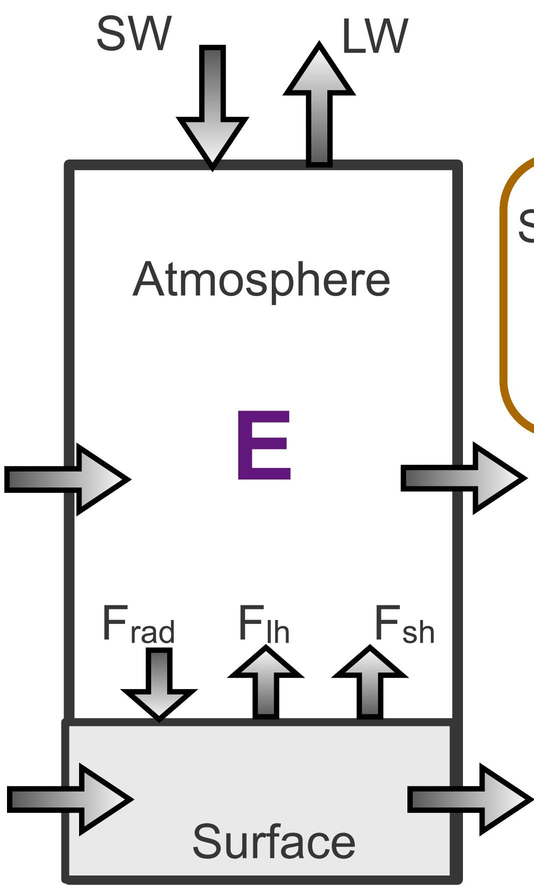
$$v(c_p T + gz + Lq + K)$$

Summary:
distribution, budget and transport





Summary:
distribution, budget and transport



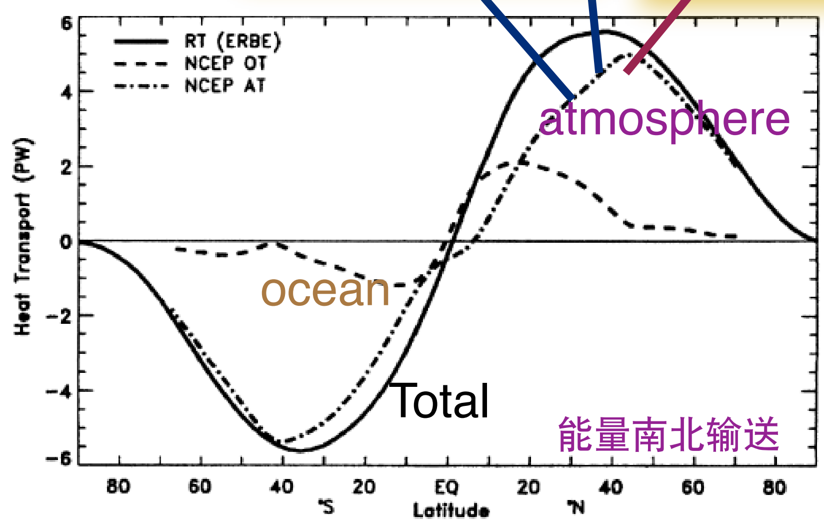
Energy transport:

$$v(c_pT + gz + Lq + K)$$

Similar meridional distribution; different vertical distribution

Dominant component

Transient eddy dominant



Wunsch (2005), J. Climate

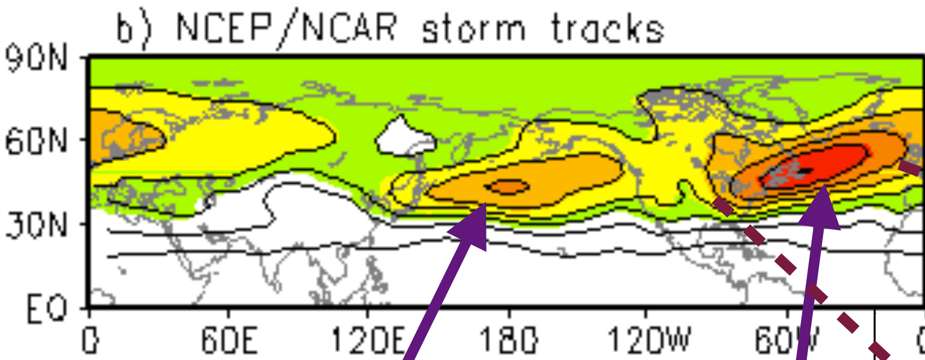
第五章:

大气环流中的纬向环流系统

5.1 Storm Tracks

授课教师：张洋

Observed features



Shaded: standard deviation of 24-h filtered 500-hPa geopotential height (contour interval 20 m) computed from the Januaries of 1982-1994 (NCEP/NCAR reanalysis)

Two storm track zones in N.H.

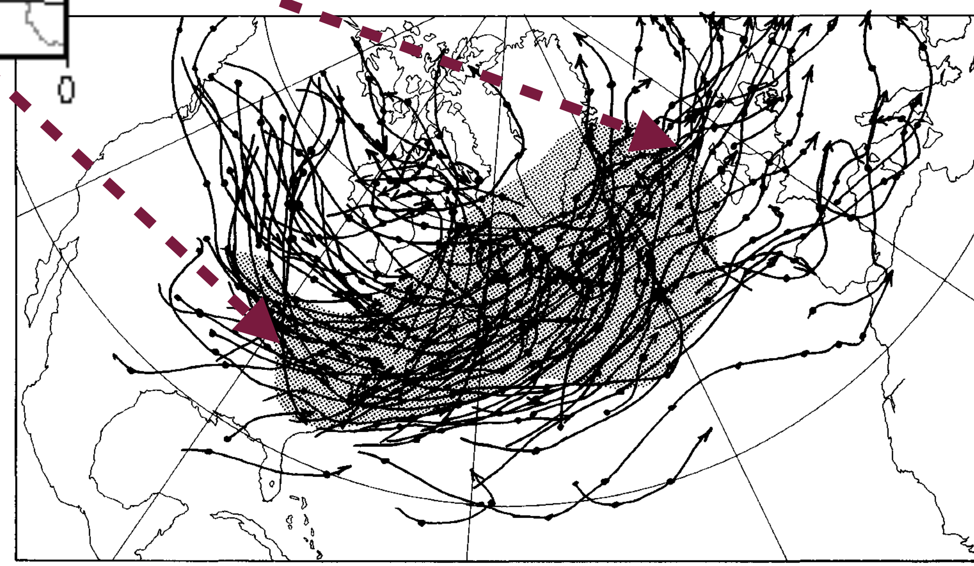


Fig. 7.9. The tracks of low pressure centres over the North Atlantic for the period December 1985 to February 1986. The shading indicates the region where the high frequency $\overline{Z'^2}^{1/2}$ exceeded 90 m in the ECMWF analyses for the same period.

Observed features

- Summary:
  - Structure: zonally located in the north Pacific and Atlantic, with the mean flow baroclinicity, jet, eddy activity, eddy heat and momentum flux in different zonal distribution.
  - Seasonal variation: different variations between the Pacific and Atlantic storm tracks; for the Pacific storm zone, mid-winter minimum observed.
  - Inter-annual variation: Pacific storm track shifts equatorward and downstream during El Nino years.
  - Decadal variation: variations in intensity occur in both storm zones, with the storm tracks in the 1990s stronger than in the 1960s.

Storm track dynamics
- from the baroclinic eddy life cycle

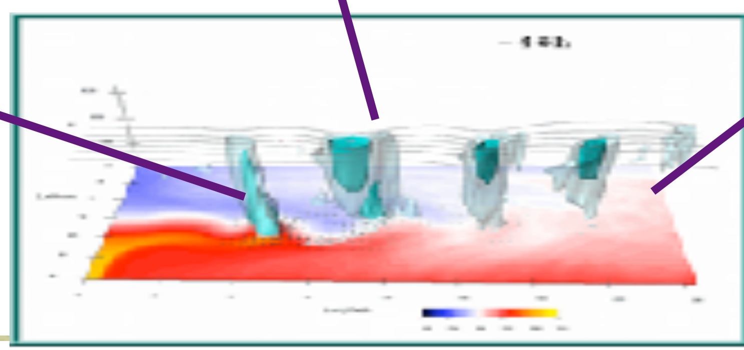
Baroclinic eddy life cycle in **time**:

Relatively small amplitude pert. → Baroclinic growth → Finite amplitude perturbation → Barotropic decay

Storm track structure can heuristically equate with an eddy life cycle in **space**:

Upstream end: perturbations are introduced and begin develop. (entrance region)

develop in space and time



Downstream end: decay stage of the eddy life cycle. (exit region)

Storm track dynamics
- Transient eddy energy budget

For storm tracks, define a **total transient eddy energy**:

$$E = K_{\mathrm{TE}} + P_{\mathrm{TE}} = \frac{1}{2}\overline{(u'^2 + v'^2)} + \frac{c_p}{2}\Gamma\overline{(T'^2)} = \frac{1}{2}\overline{(u'^2 + v'^2)} - \frac{\alpha_m}{2\theta_m}\frac{\overline{\theta'^2}}{\partial\theta_s/\partial p}$$

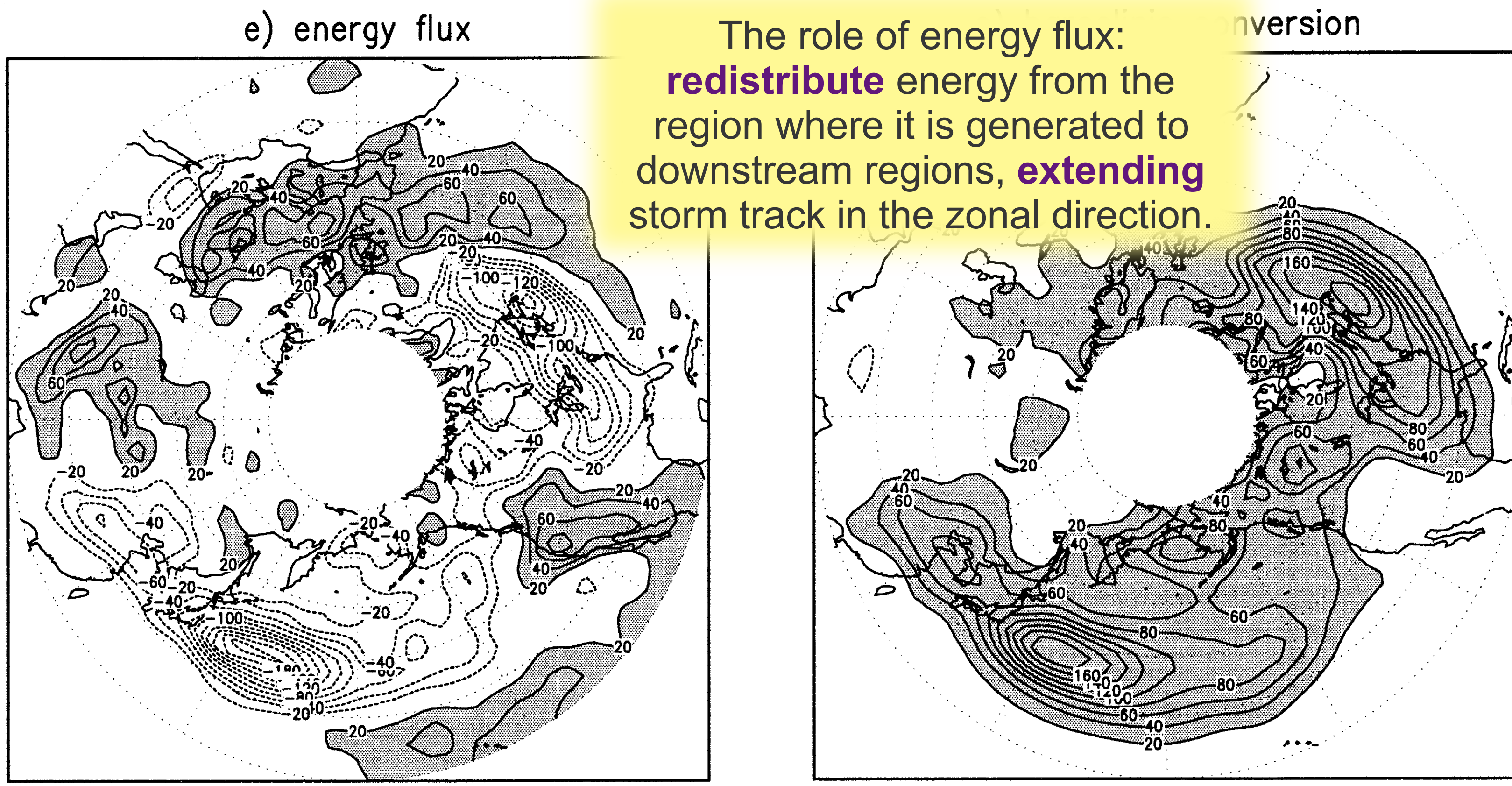
Transient eddy energy budget:

$$\frac{\partial E}{\partial t} = \nabla\cdot\overline{(\mathbf{v}E + \mathbf{v}'_{\mathbf{a}}\phi')} + \frac{\alpha_m}{\theta_m}\frac{\overline{\mathbf{v}'\theta'}}{\partial\theta_s/\partial p}\cdot\nabla\theta - \overline{\mathbf{v}'\cdot(\mathbf{v}'\cdot\nabla)V_m} - \mathrm{diss} + \mathrm{diab}$$

advective energy flux	baroclinic generation	barotropic conversion	$D(K_E)$	$G(P_E)$

Storm track dynamics

- Transient eddy energy budget

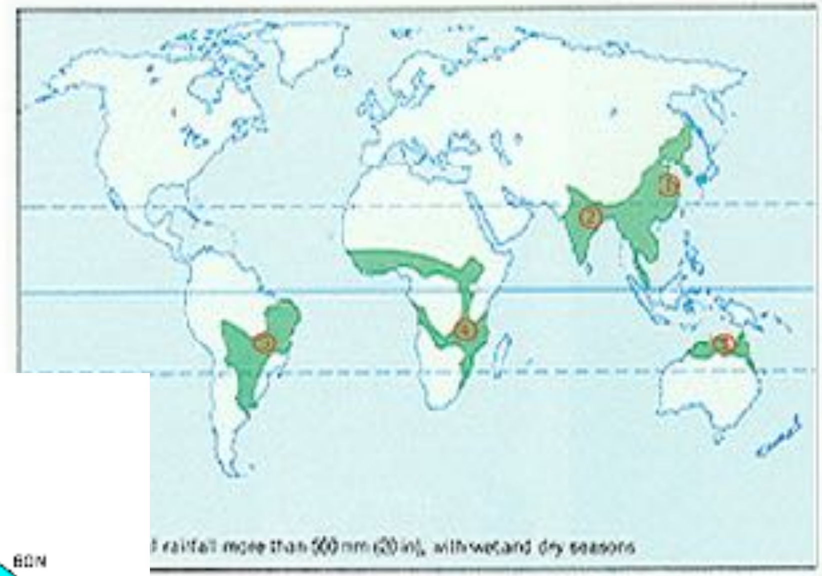


The role of energy flux: **redistribute** energy from the region where it is generated to downstream regions, **extending** storm track in the zonal direction.

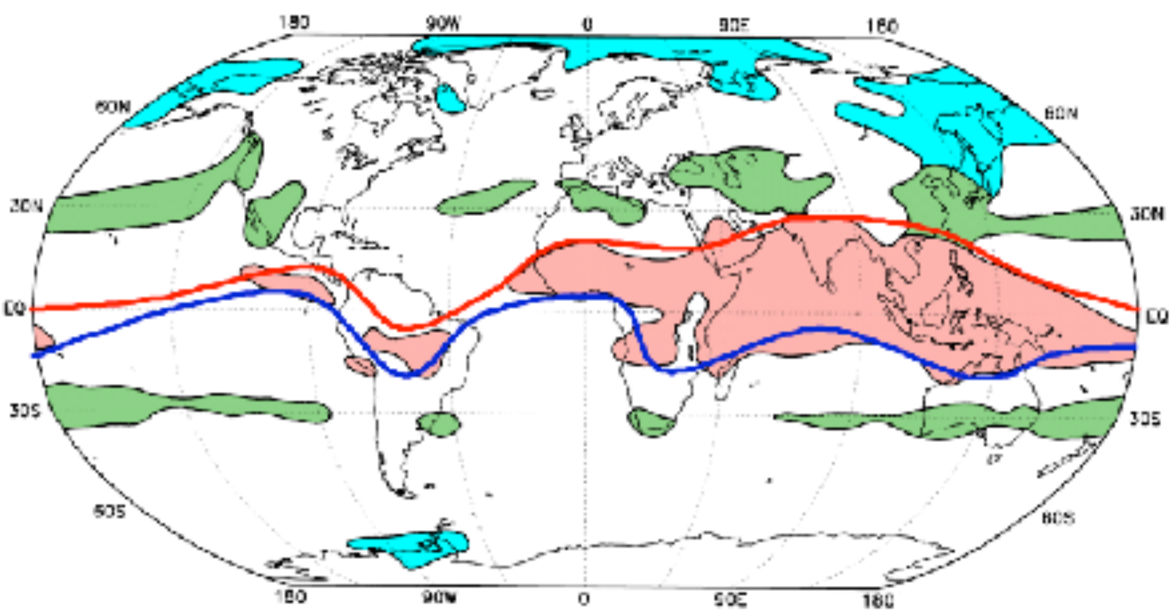
Strongly compensate the baroclinic conversion term in the entrance region.

from Chang et al, JC, 2002

Introduction



Geographical Extent of the Global Surface Monsoons



The red, green, and blue areas indicate the tropical, subtropical, and temperate-frigid monsoons, respectively. The red and blue thick lines represent the ITCZ in summer and winter, respectively. (Li, J., and Q. Zeng, 2005)

- 65% of world's population lives within monsoon;
- Monsoon precipitation is crucial to the life, food production, economy et al in these regions;
- Proper forecasting of location and quantity of precipitation is crucial to theses regions.

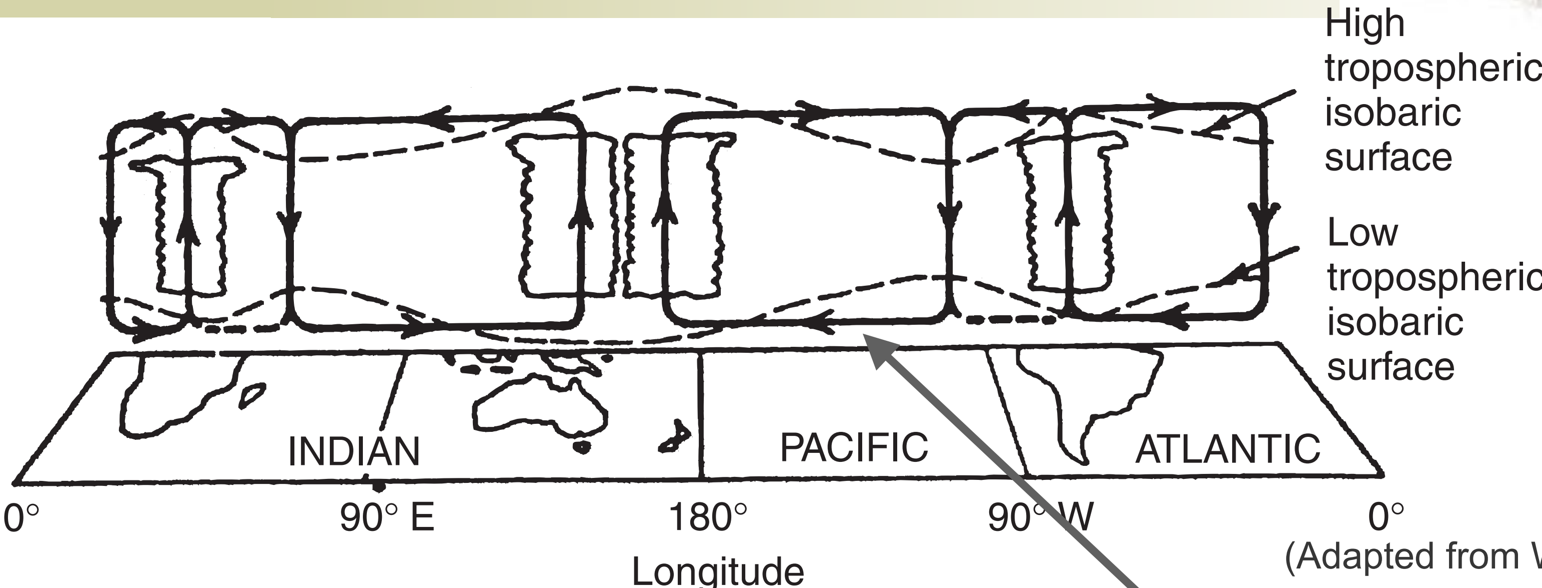
Observed features

- Summary:
  - A monsoon climate is characterized by the obvious **seasonal reverse** of wind, precipitation and atmospheric circulation.
  - From **a global view**: south asian monsoon is associated with the seasonal migration of ITCZ and Hadley circulation, which also plays an important role in the global meridional moisture and latent energy transport.
  - South asian monsoon exhibits obvious **sudden onset,** with the low-level winds and the whole monsoonal circulation built in two weeks.
  - **Intra-seasonal variation:** show periods in 4-5 days, 10-20 days and 40-50 days.
  - **Inter-annual variation:** Relatively weaker precipitation occurs during El Nino years.

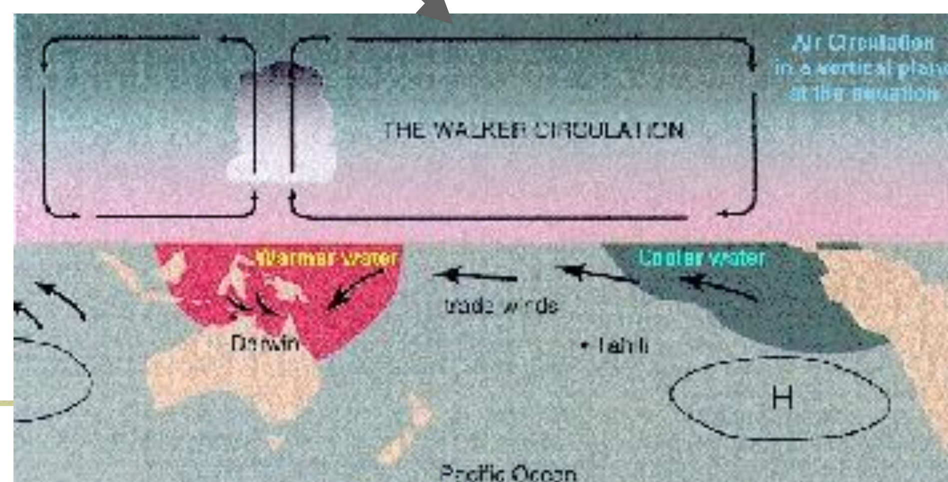
Monsoon dynamics

- Land-sea contrast
  - thermal contrast: strongest heating over subtropical land
  - moisture advection: provide precipitation water
- Orography
  - Thermal forcing as an upper level heat source
  - Mechanical forcing:
    - a local precipitation enhancement
    - a widespread barrier of cold, dry air
- GCM results
  - strong seasonal heating due to the small heat capacity of the underlying surface seems to be crucial to the formation of monsoonal circulation; monsoonal circulation is associated with the eddy activity transition;
  - the special topography of south asian reinforces the monsoon, especially by protecting warm and moist tropical air from the cold and dry extratropics
  - thermal heating from the south slope of TP suggested strengthen the monsoon
- Monsoon variation in timescales as intra-seasonal, inter-annual scales needs further studies

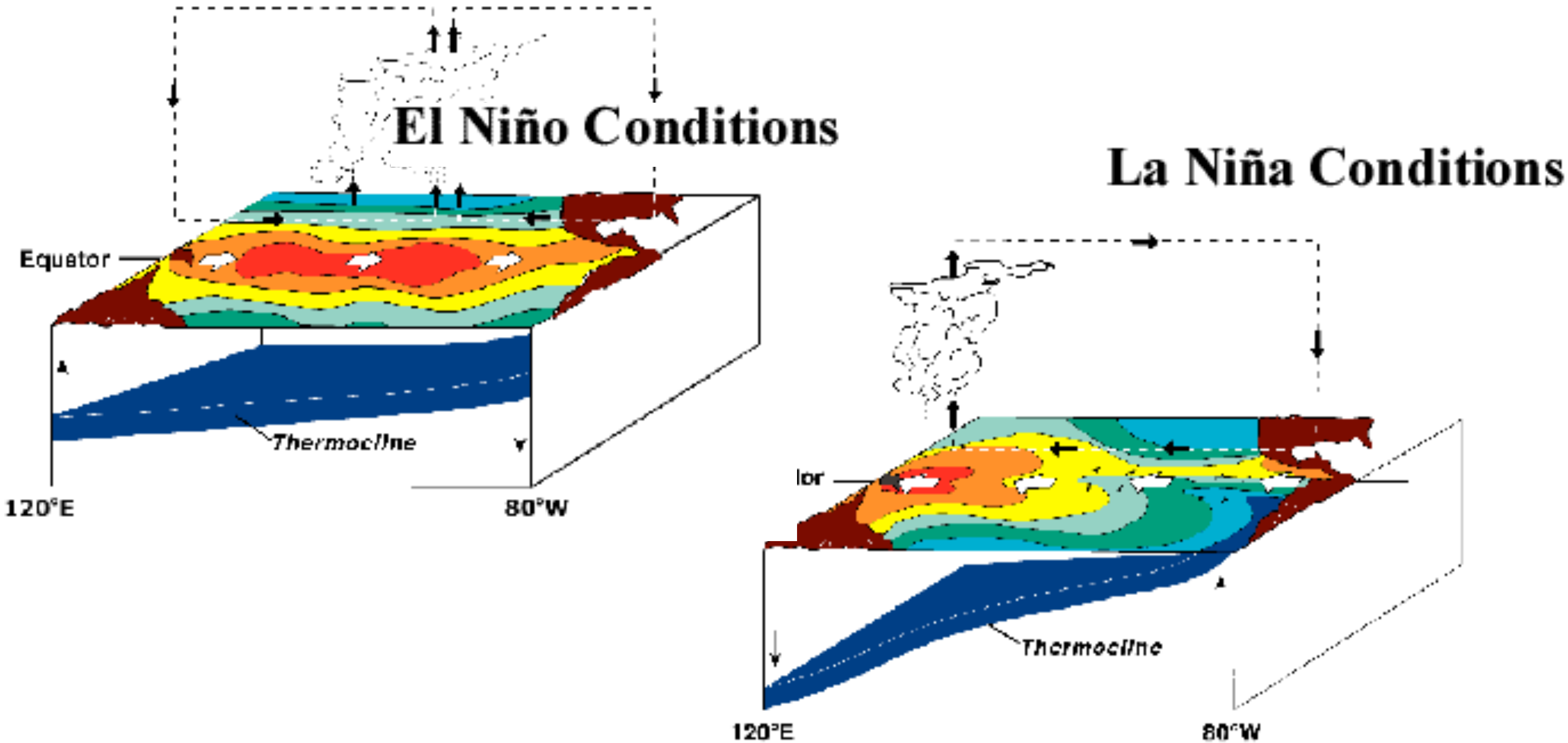
Introduction



(Adapted from Webster 1983)

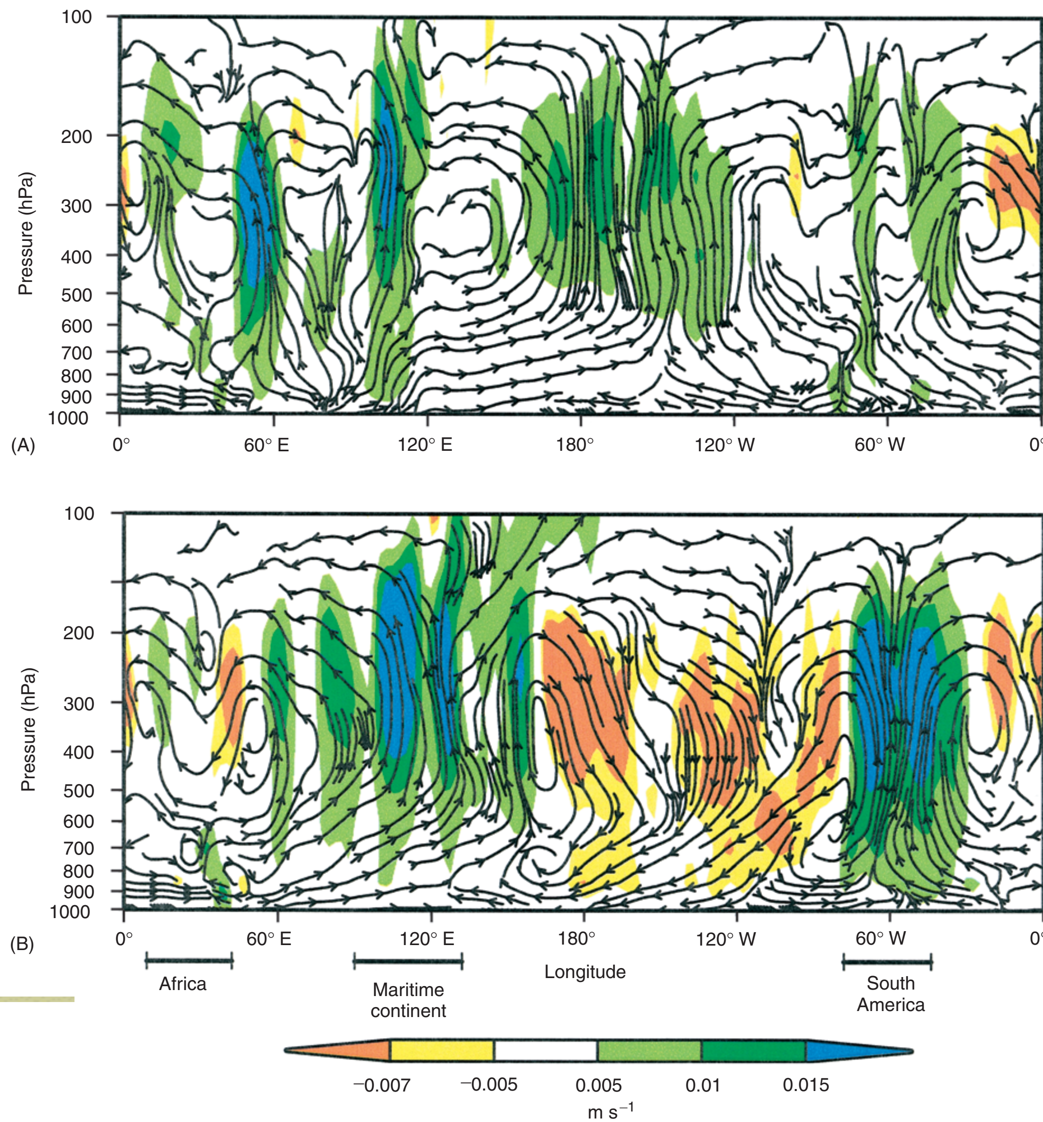


ENSO and Walker Circulation



Adapted from NOAA

ENSO and Walker Circulation



El Nino years:

Raising motion prevailed at almost all longitudes with a peak in central pacific.

La Nina years:

An enhanced Walker Circulation.

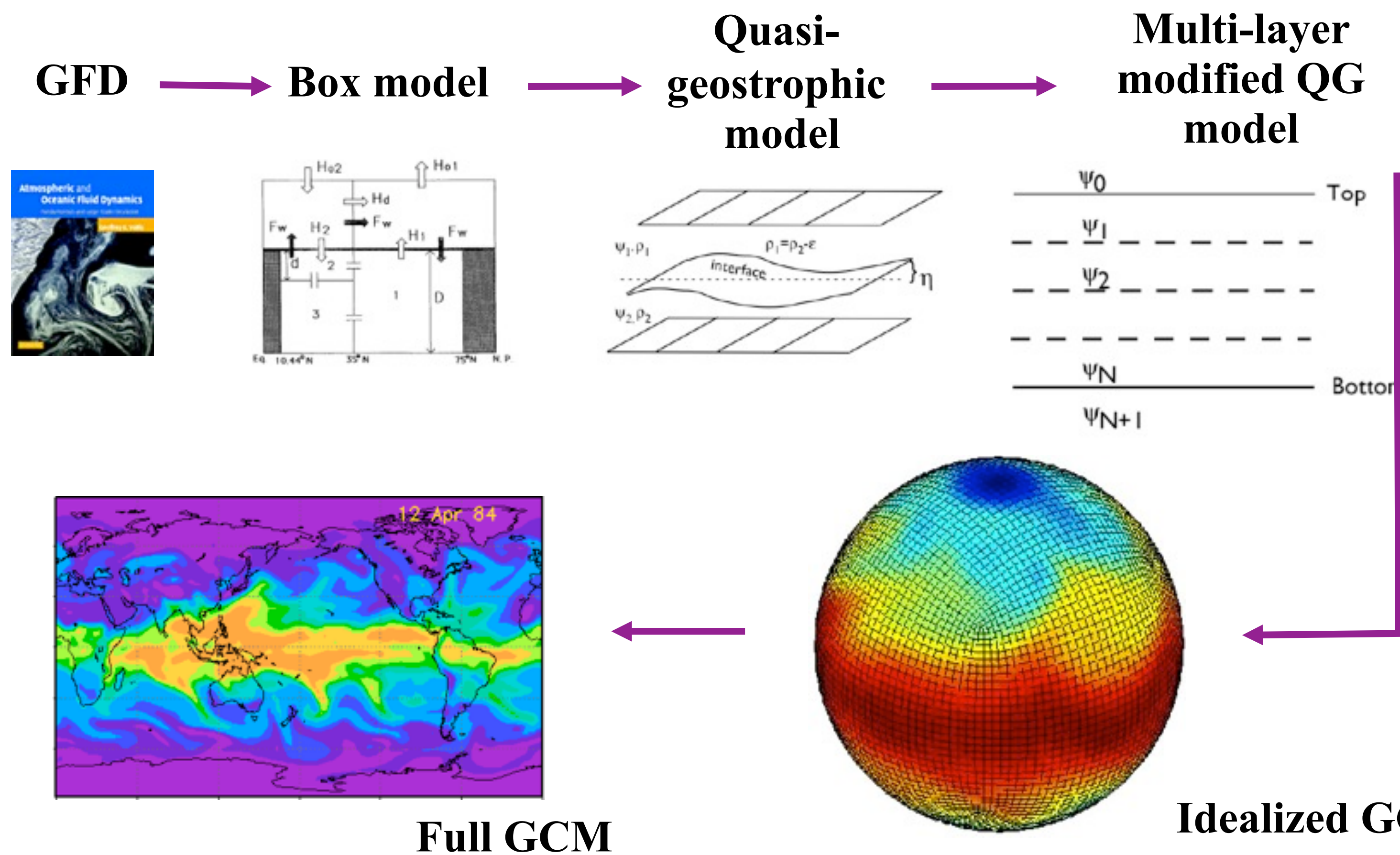
Adapted from Lau et al, 2002

Uncertainties of full GCM

- A summary from the AMIP I results:
  - Ensemble mean shows that the average large-scale seasonal distributions of pressure, temperature, and circulation are reasonably close to what are believed to be the best observational estimates available;
  - The average large-scale distributions of pressure, temperature and circulation shows relatively large intermodel differences in high/polar latitudes compared to low/mid latitudes.
  - The large-scale structure of the ensemble mean precipitation also resembles the observed estimates but show particularly large intermodel differences in low latitudes.
  - The total cloudiness, on the other hand, is rather poorly simulated.

A hierarchy of GCMs: From idealized model to full GCM

- An example for using hierarchy of models to study the role of eddies



期末考试

- 考前答疑时间：12月22日下午1：00 - 3：30，大气楼B410
- 考试时间：12月23日（周四），下午1：00- 4：00
- 地点：仙I-212
- 要求：
    - 闭卷，但是可以携带30条公式
    - 携带计算器
    - 试卷以6-7道大题的形式，认真回答每一个小问题