Lecture 2 Conservation laws and planetary waves

• Potential vorticity

• Orographic forcing of Rossby waves

• Barotropic Instability

•Baroclinic Instability

• Fronts and wave breaking

• Midlatitude storm tracks

Planetary Balance and conservation

If the time scale τ of the motion is larger than :

$\tau > f^{-1} > N^{-1}$

$\tau > 20$hs > 10min

And for scales large such U < L/f, L<~ a (earth radius).

The flow is in *hydrostatic balance*

Buoyancy ~vertical component of pressure force.

And in *geostrophic balance*:

coriolis force ~ horizontal component of pressure force.

Hydrostatic Balance

$$\frac{\partial P}{\partial z} = -\rho g$$

$$\pi = \left(\frac{P}{P_0}\right)^{\kappa}$$

$$\frac{1}{\rho}\frac{\partial P}{\partial z} = C_p \theta \frac{\partial \pi}{\partial z}$$

$$C_p \theta_0 \frac{\partial \pi}{\partial z} = g\frac{\vartheta}{\theta_0}$$

Geostrophic Balance

If the Rossby Number is much smaller than 1 (R_0 = U/(fL) << 1)

$$-u 2\Omega \sin\phi_0 = \frac{C_p \theta_0}{a\cos\phi_0}\frac{\partial \pi}{\partial \lambda}$$

$$v 2\Omega \sin\phi_0 = \frac{C_p \theta_0}{a}\frac{\partial \pi}{\partial \phi}$$

Where *u* and *v* are zonal and meridional velocity components and ϕ_0 a particular latitude

Conservation of Potential Vorticity

Full system

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

$$Q=(f+\zeta)\theta_z+\eta\theta_y+\mu\theta_x$$

$$\zeta=\frac{\partial v}{\partial x}-\frac{\partial u}{\partial y}$$

$$\eta=\frac{\partial u}{\partial z}-\frac{\partial w}{\partial x}$$

$$\mu=\frac{\partial w}{\partial x}-\frac{\partial v}{\partial z}$$

Quasi-geostrophic

$$\frac{\partial q}{\partial t}+u_g\frac{\partial q}{\partial x}+v_g\frac{\partial q}{\partial y}=0$$

$$q=(\zeta+f_0+\beta y+\frac{f_0}{\rho_r}\frac{\partial}{\partial z}\left\{\frac{\rho_r}{N^2}\phi_z\right\}$$

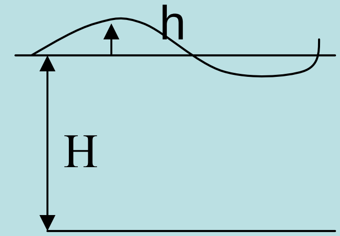
$$\zeta=\frac{\partial v}{\partial x}-\frac{\partial u}{\partial y}=\nabla^2\phi$$

$$N^2=\frac{g}{\theta_r}\frac{\partial\theta}{\partial z}$$

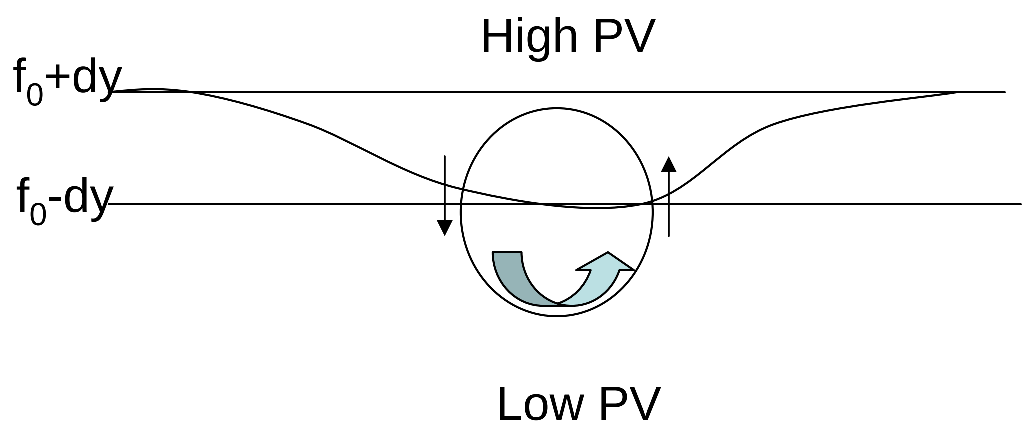
Shallow water

$$\frac{\partial q_s}{\partial t}+u_s\frac{\partial q_s}{\partial x}+v_s\frac{\partial q_s}{\partial y}=0$$

$$q_s=(\frac{\zeta+f}{H+h})$$



Rossby Waves



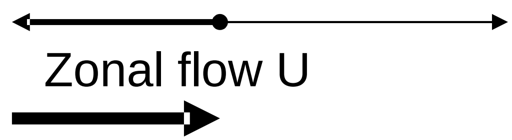
$Q_s=(\zeta_1+f_1)/H$

$Q_s=(\zeta_2+f_2)/H$

$\zeta_2 > \zeta_1$

Possibility of stationary because:

Phase group (Energy flows to the east)



Zonal flow U

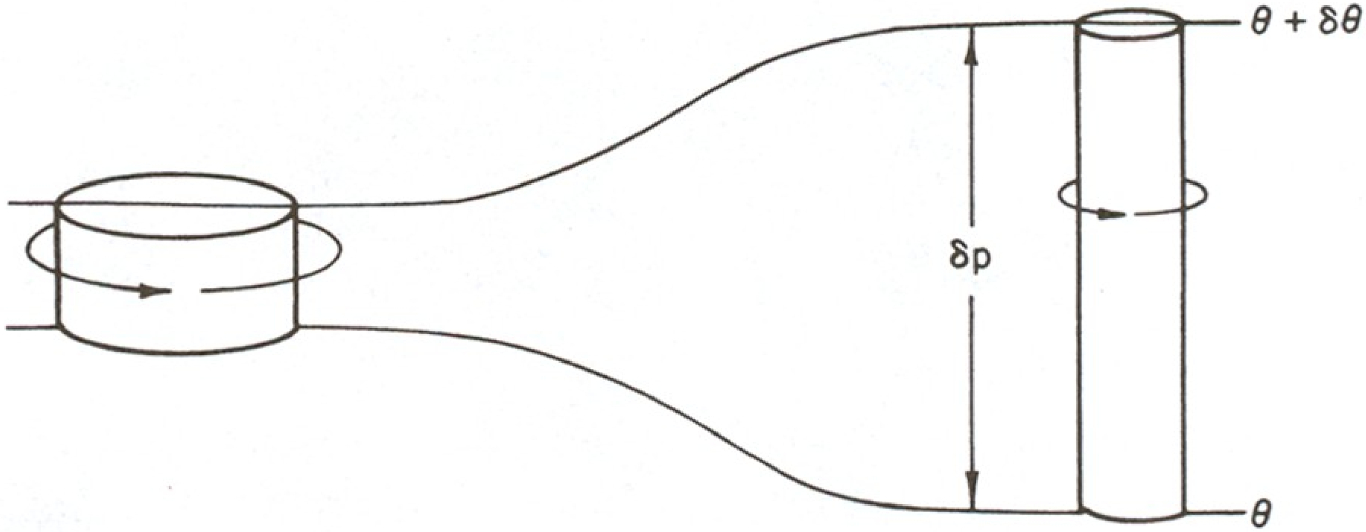
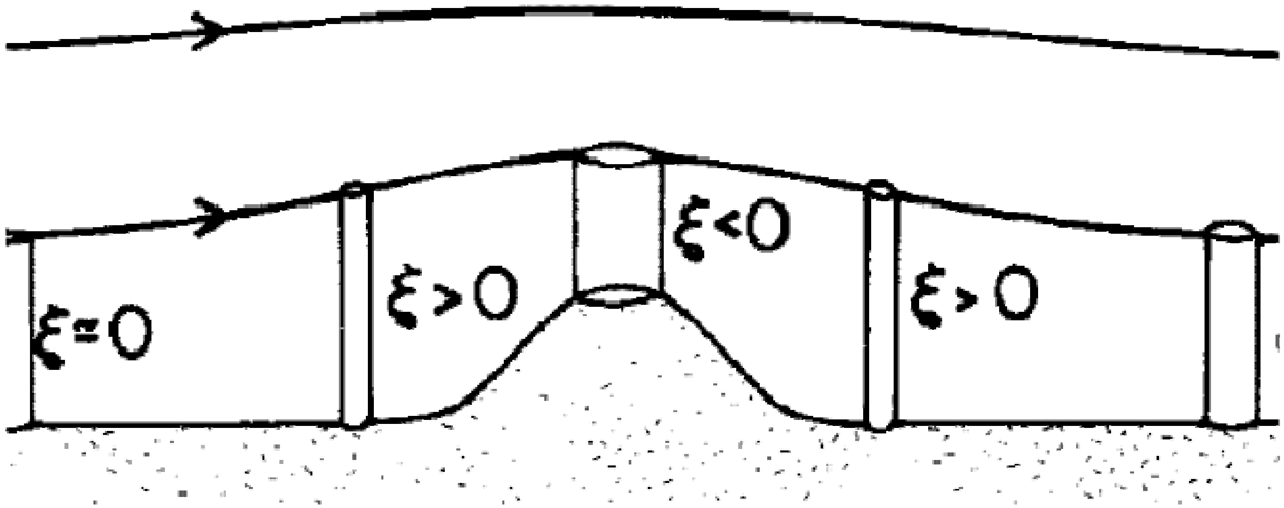
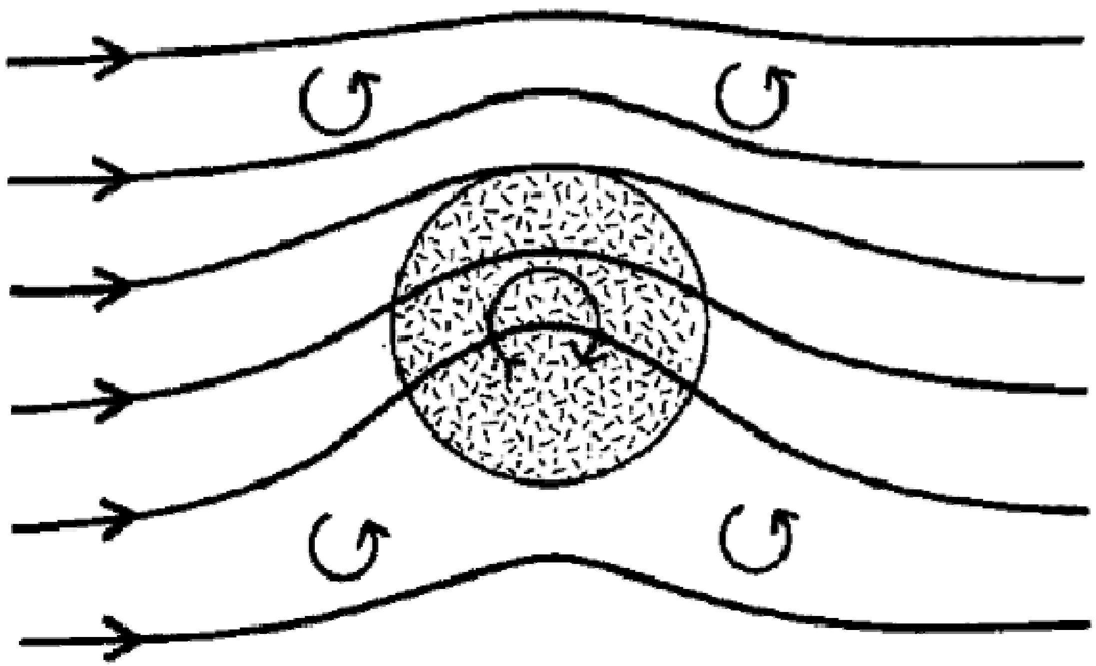


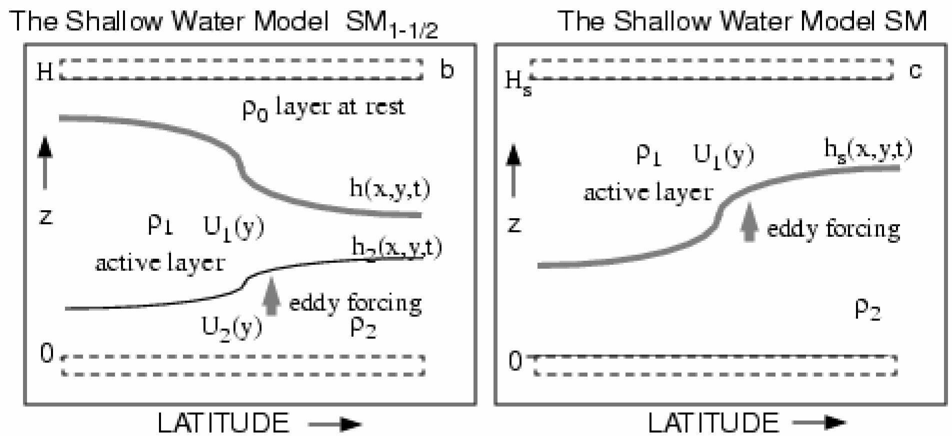
Fig. 4.7 A cylindrical column of air moving adiabatically, conserving potential vorticity.

[3] Named for the German meteorologist Hans Ertel. A more general form of Ertel's potential vorticity is discussed, for example, in Gill (1982). Potential vorticity is often expressed in the potential vorticity unit (PVU), where 1 PVU $= 10^{-6}$ K kg^{-1} m^{2} s^{-1}.



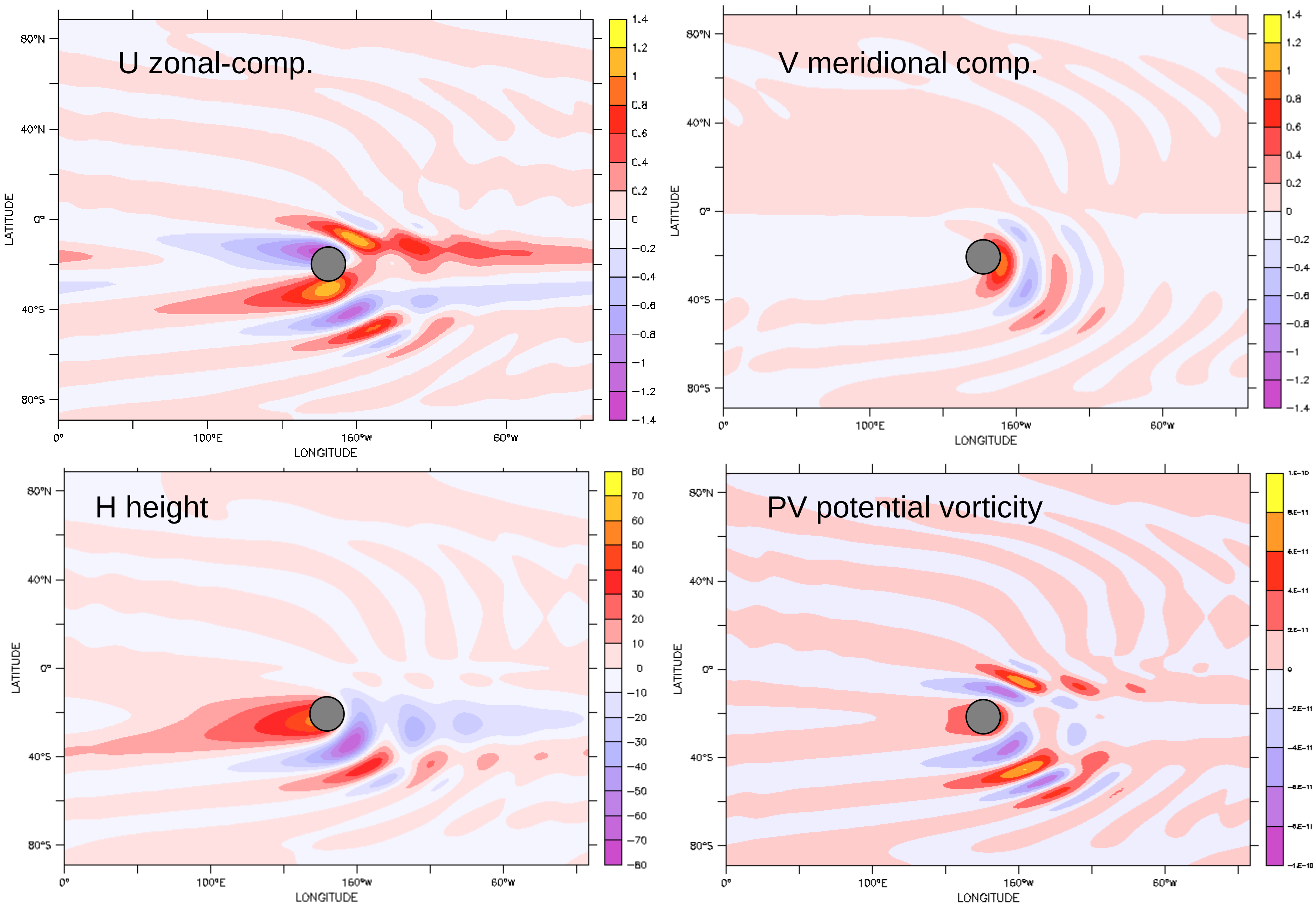
$\xi \approx 0$
$\xi > 0$
$\xi < 0$
$\xi > 0$

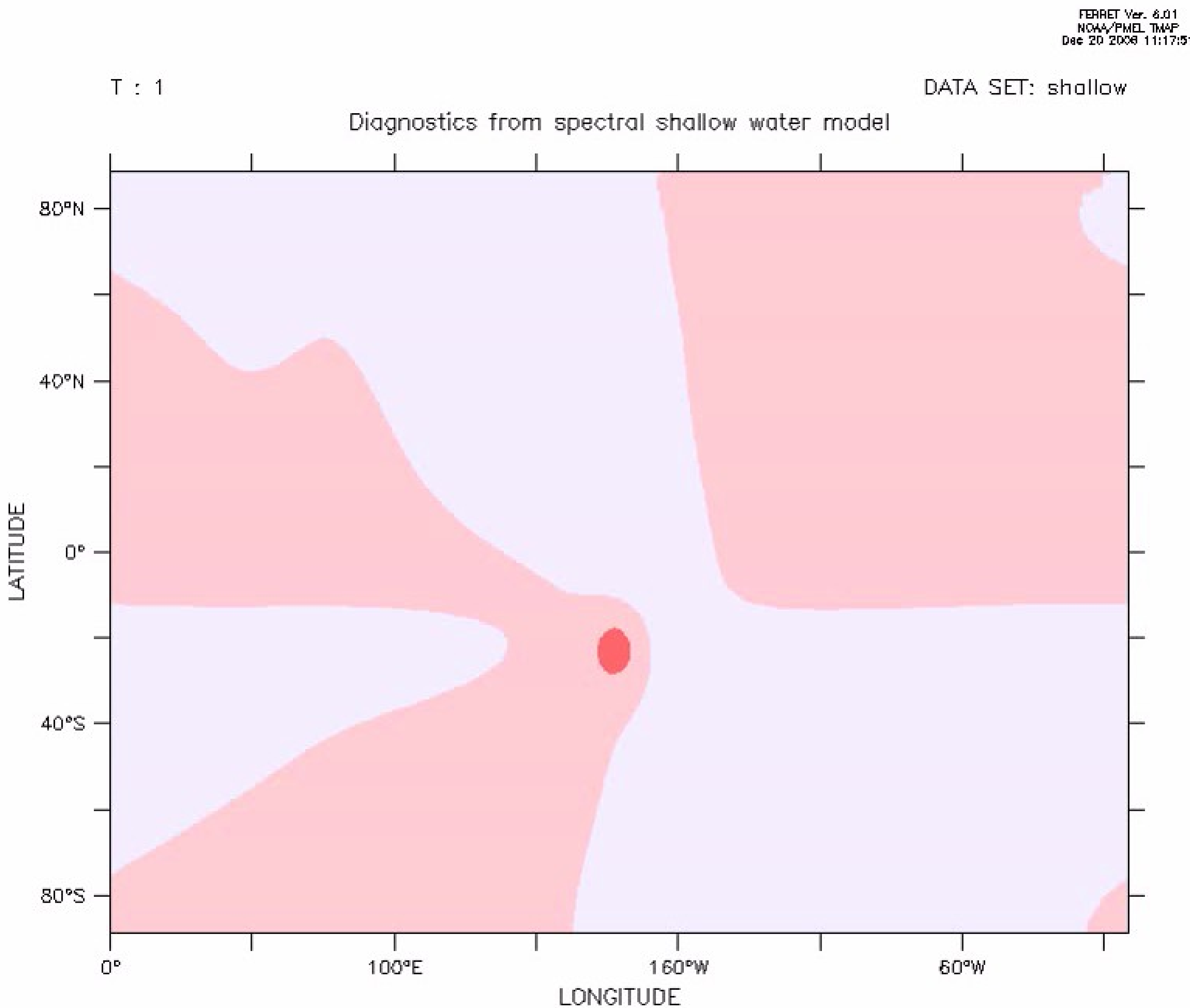




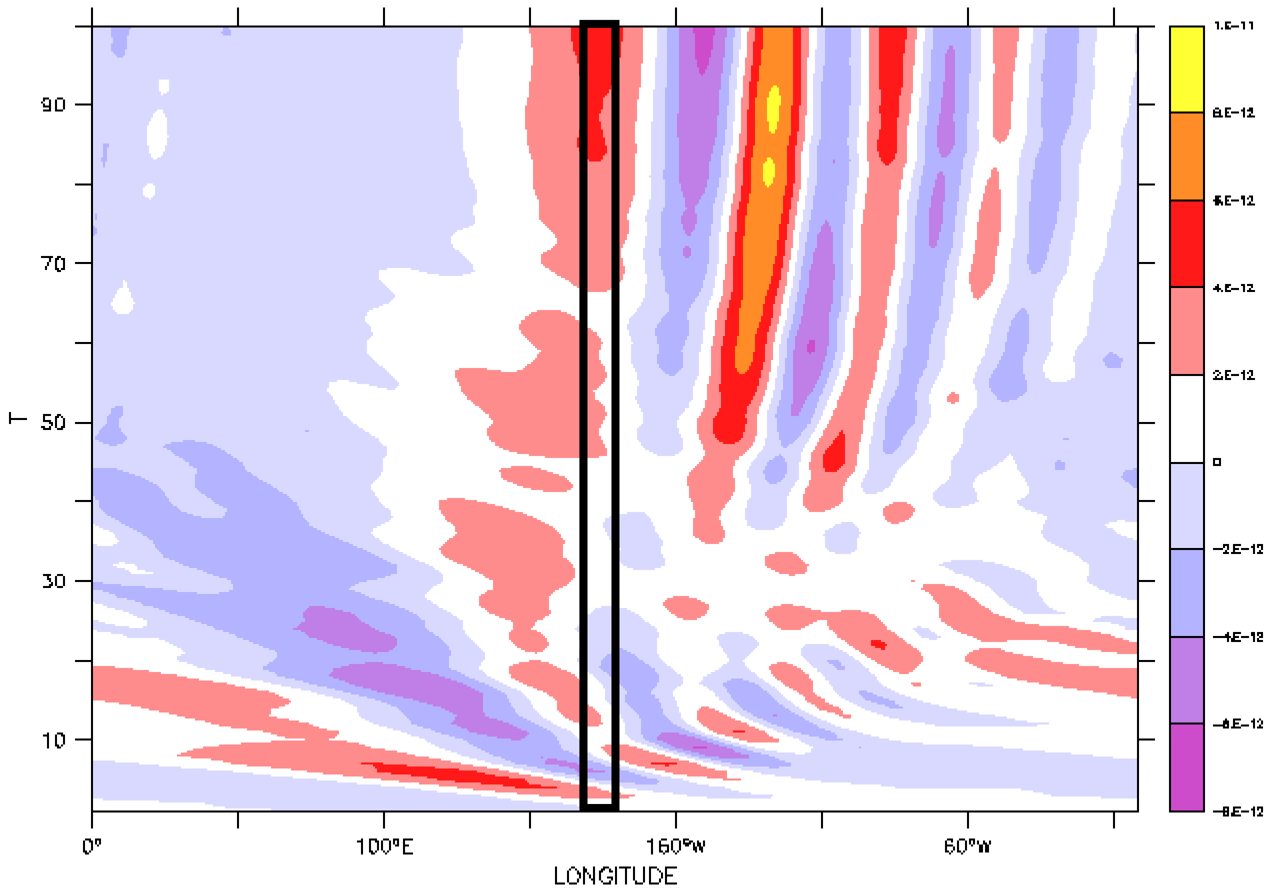
The Shallow Water Model SM1-1/2
H
b
ρ0 layer at rest
z
h(x,y,t)
ρ1
U1(y)
active layer
h2(x,y,t)
eddy forcing
U2(y)
ρ2
0
LATITUDE
The Shallow Water Model SM
Hs
c
ρ1
U1(y)
hs(x,y,t)
active layer
z
eddy forcing
ρ2
0
LATITUDE

Stationary response due to orography (shallow water global model)




FERRET Ver. 6.01
NOAA/PMEL TMAP
Dec 20 2008 11:17:51
T : 1
DATA SET: shallow
Diagnostics from spectral shallow water model
80°N
40°N
0°
40°S
80°S
LATITUDE
0°
100°E
160°W
60°W
LONGITUDE

PV−PV[X=@AVE]


90
70
50
30
10
T
0°
100°E
160°W
60°W
LONGITUDE
1.E-11
8.E-12
6.E-12
4.E-12
2.E-12
0
-2.E-12
-4.E-12
-6.E-12
-8.E-12

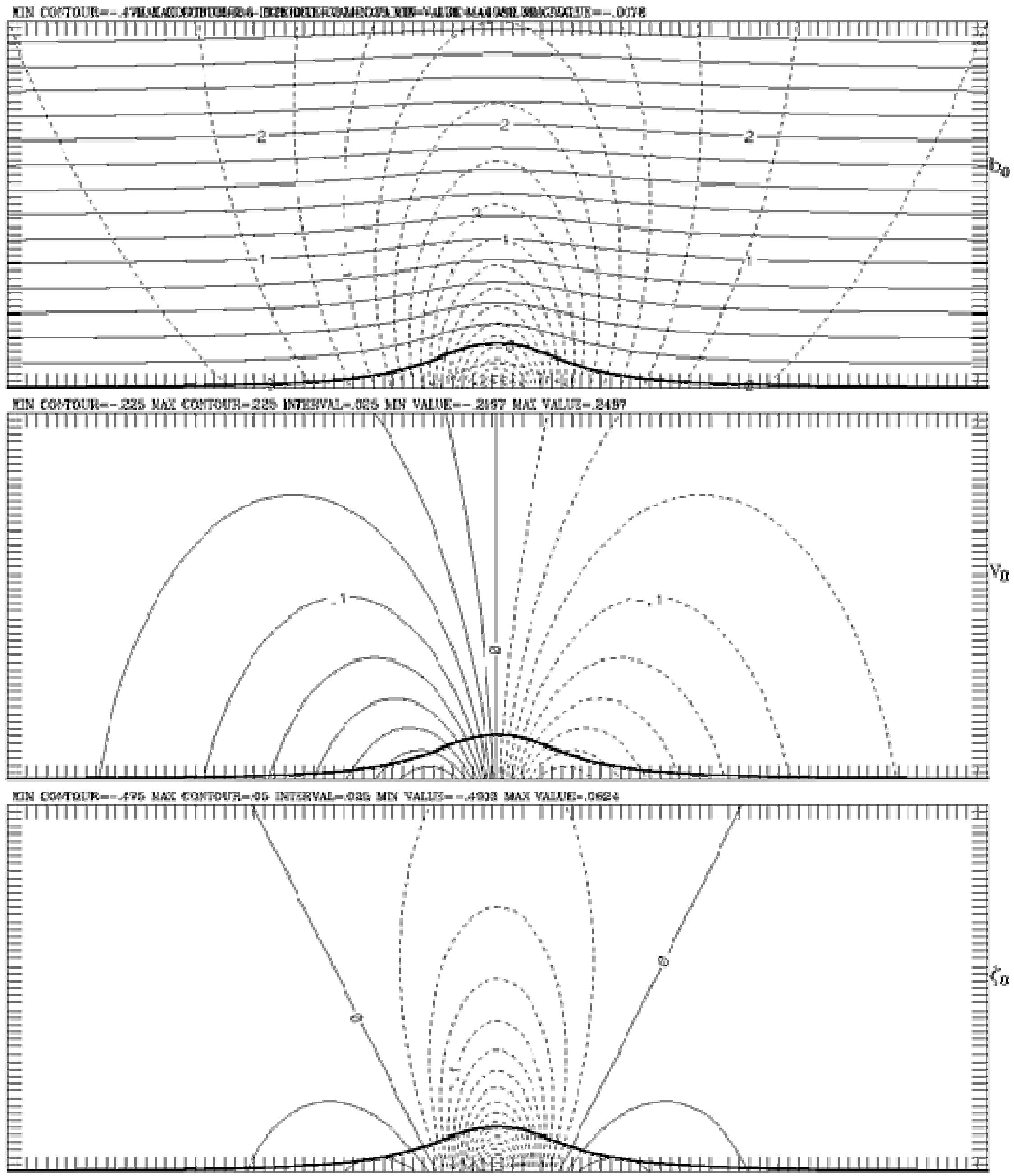
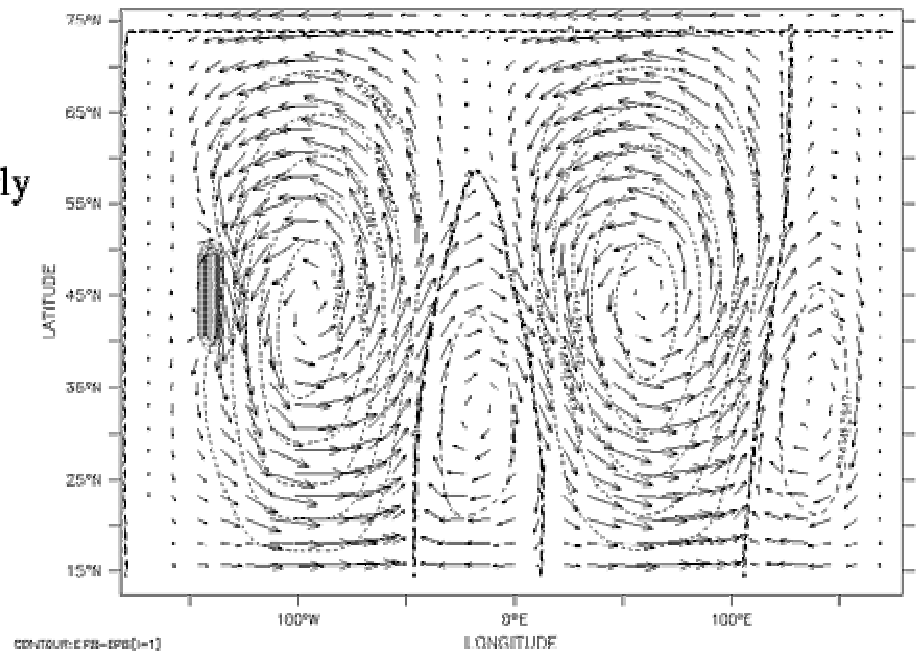
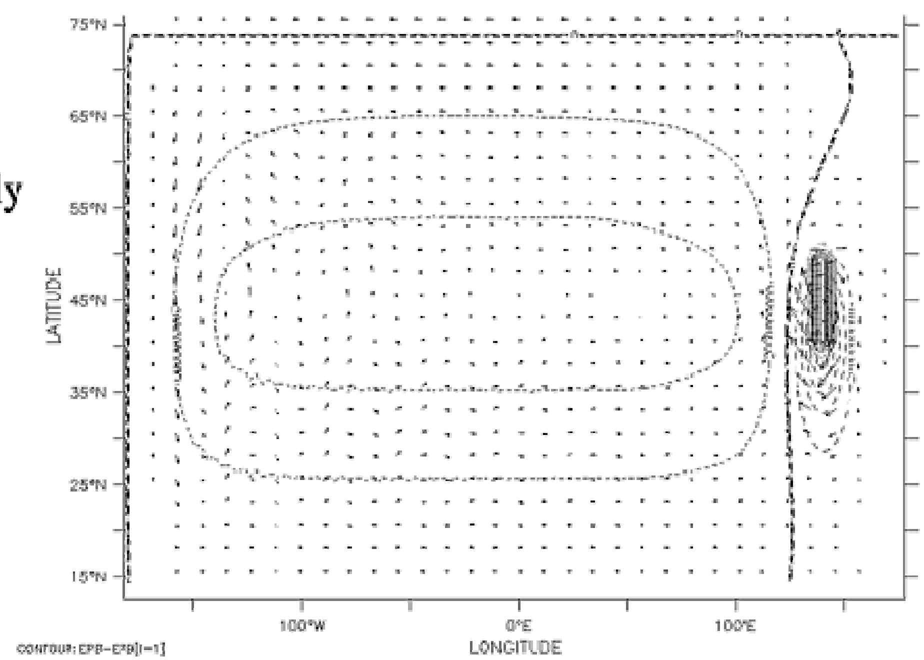


Figure 6. The quasigeostrophic solution for steady flow over a semi-infinite ridge in a semi-infinite atmosphere. The plots show (top) total buoyancy (solid) and perturbation (dashed)buoyancy, (middle) meridional velocity perturbation, and (bottom) relative vorticity perturbation. The heavy contour indicates the mountain profile.

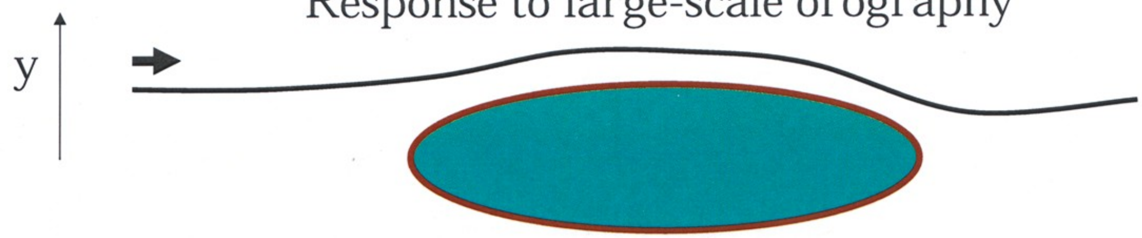
$U_0 > 0$
Westerly
flow

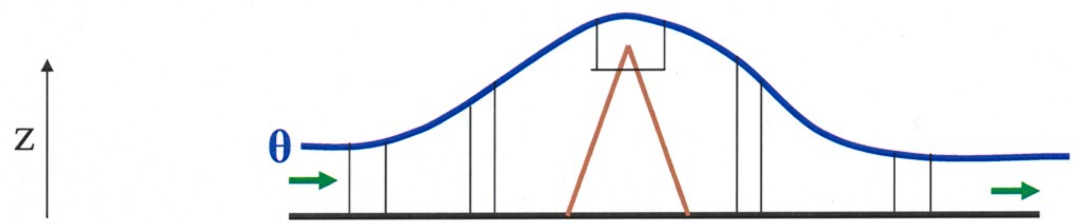


$U_0 < 0$
Easterly
flow



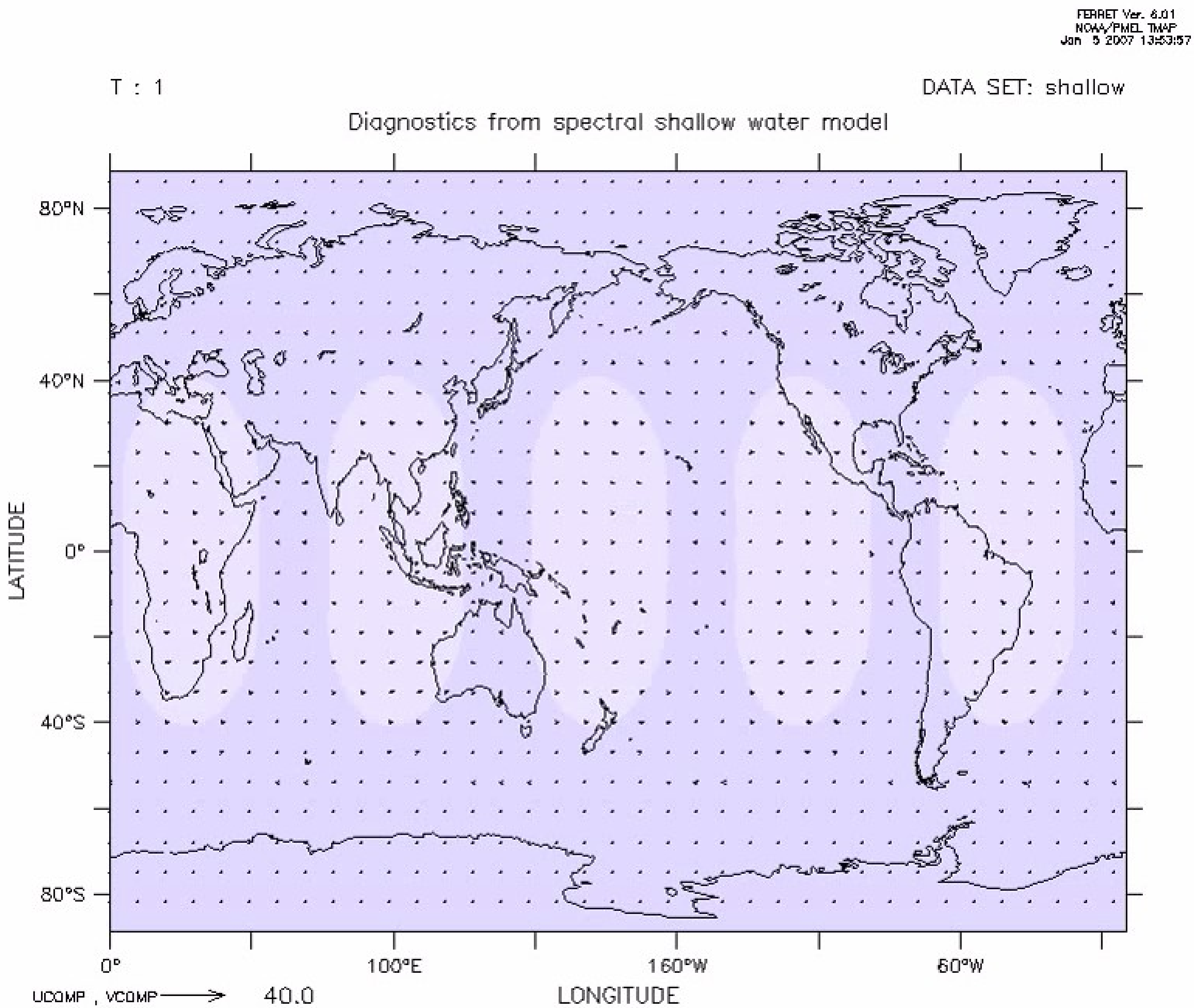
Response to large-scale orography



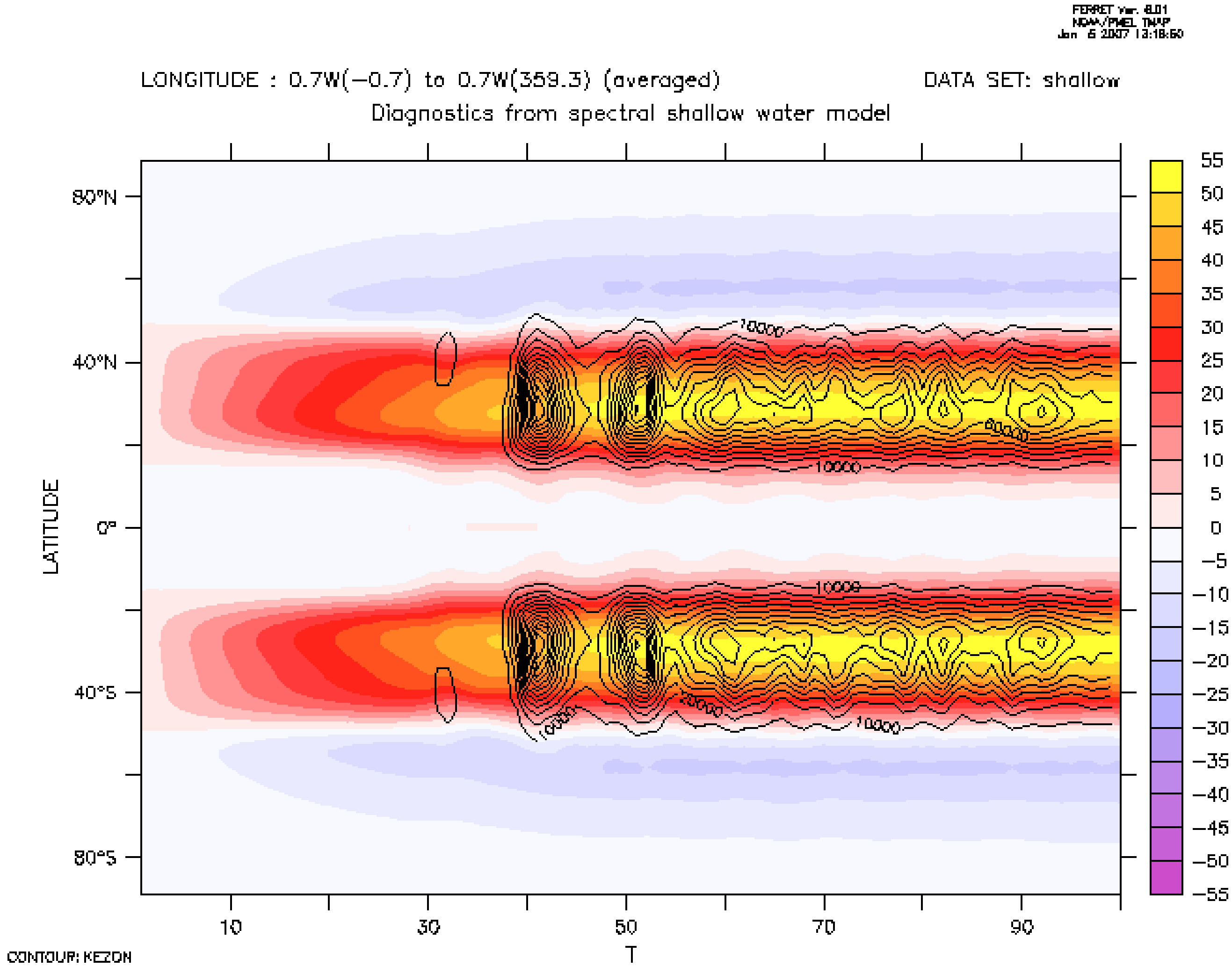


Barotropic and Baroclinic Instabilities

The major source of midlatitudes weather



geopotential (m2/s2)



FERRET Ver. 6.01
NOAA/PMEL TMAP
Jan 6 2007 13:18:50
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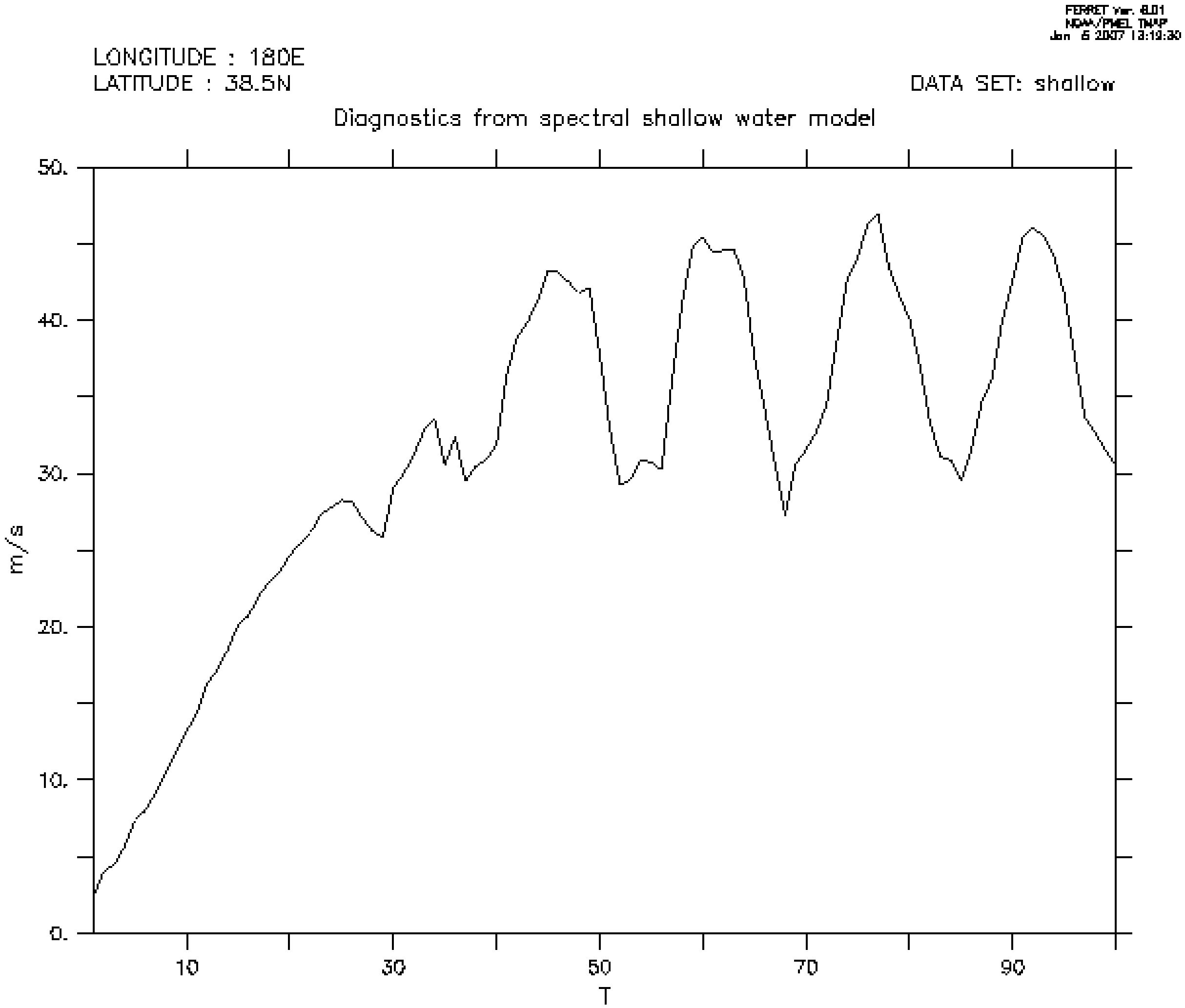
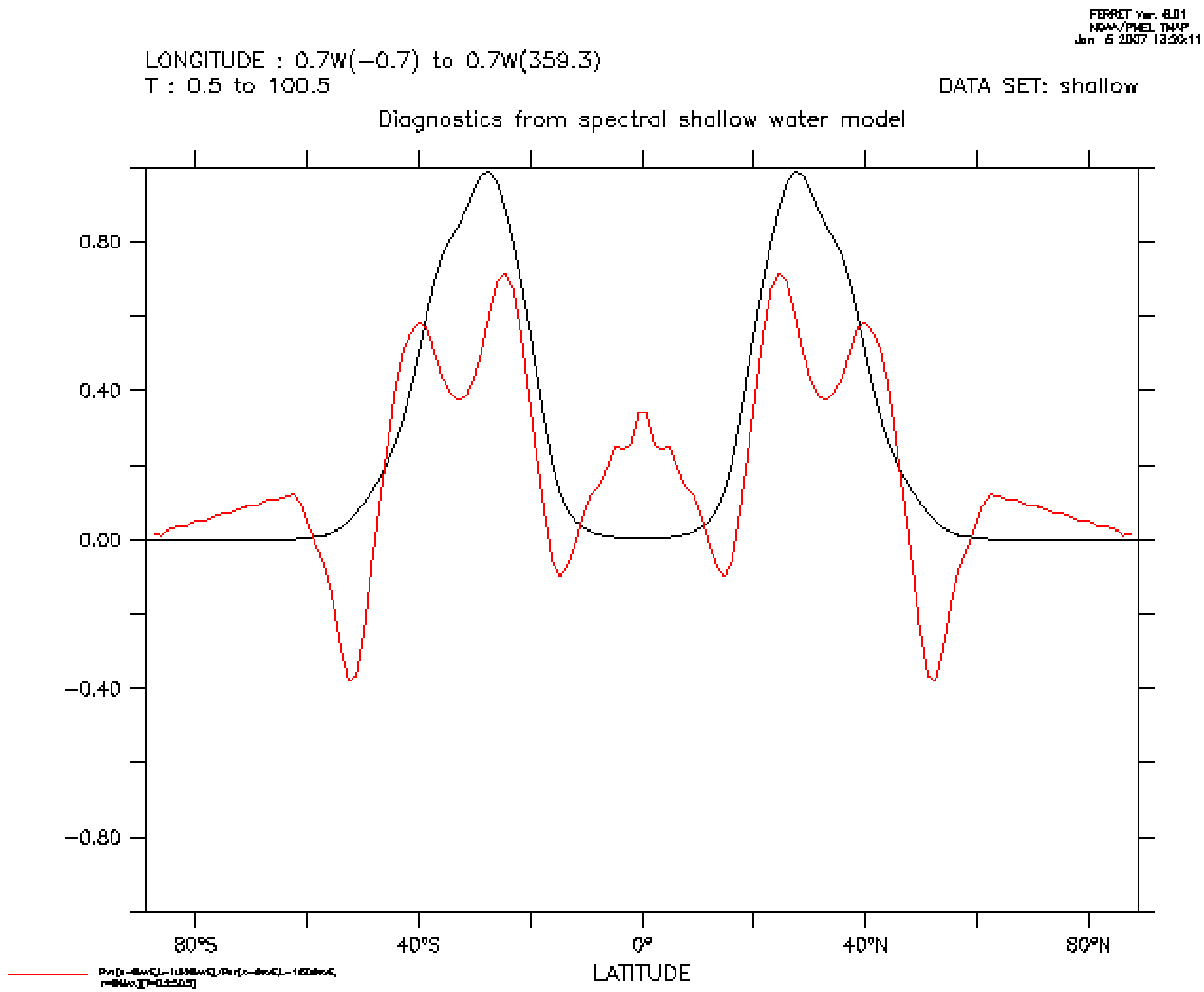
DATA SET: shallow
Diagnostics from spectral shallow water model
80°N
40°N
0°
40°S
80°S
LATITUDE
10
30
50
70
90
T
10000
CONTOUR: KEZON
55
50
45
40
35
30
25
20
15
10
5
0
-5
-10
-15
-20
-25
-30
-35
-40
-45
-50
-55

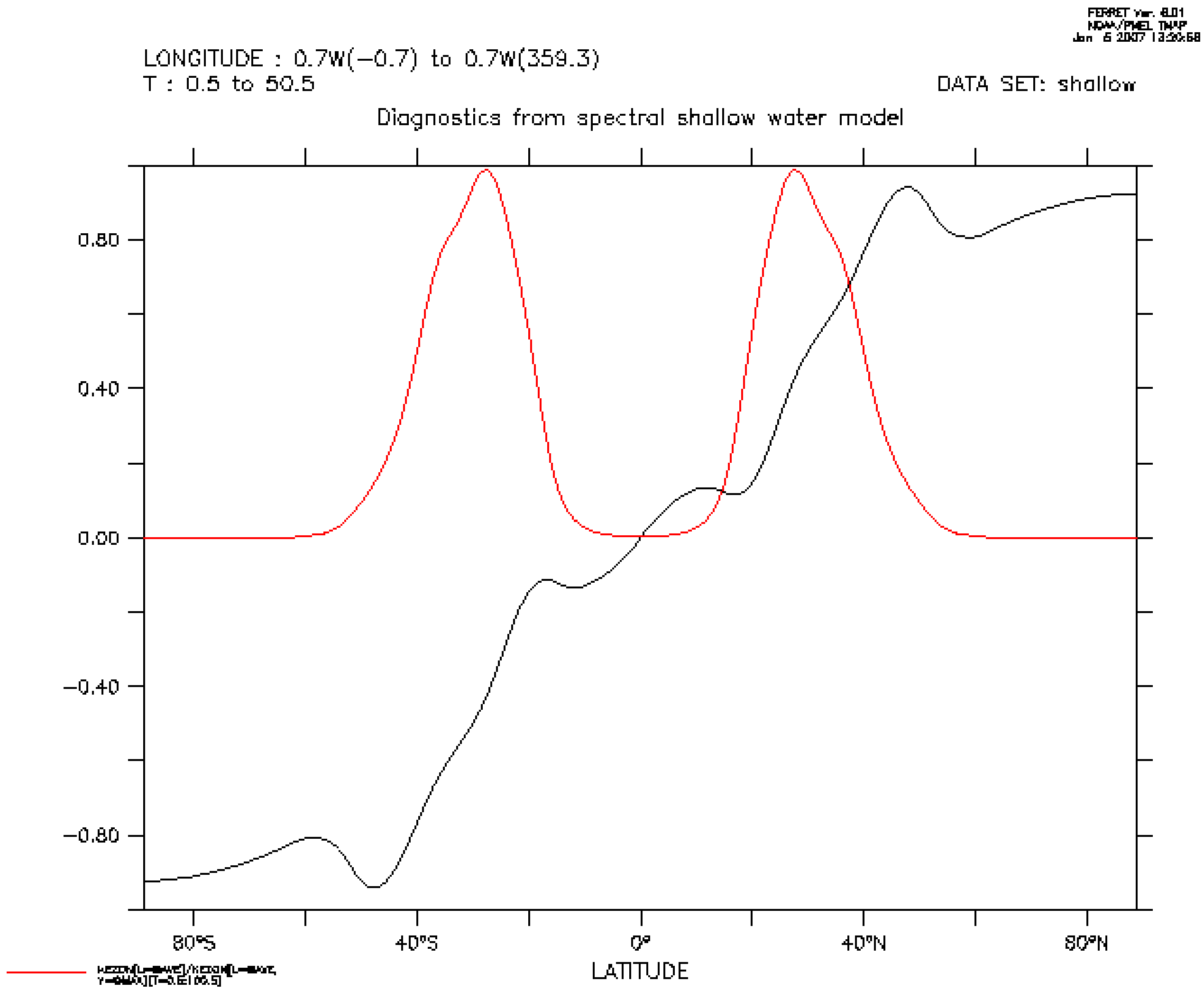
u_wind (m/s)


FERRET Ver. 6.01
NOAA/PMEL TMAP
Jan 6 2007 13:19:30
LONGITUDE : 180E
LATITUDE : 38.5N
DATA SET: shallow
Diagnostics from spectral shallow water model
m/s
0.
10.
20.
30.
40.
50.
10
30
50
70
90
T

u_wind (m/s)

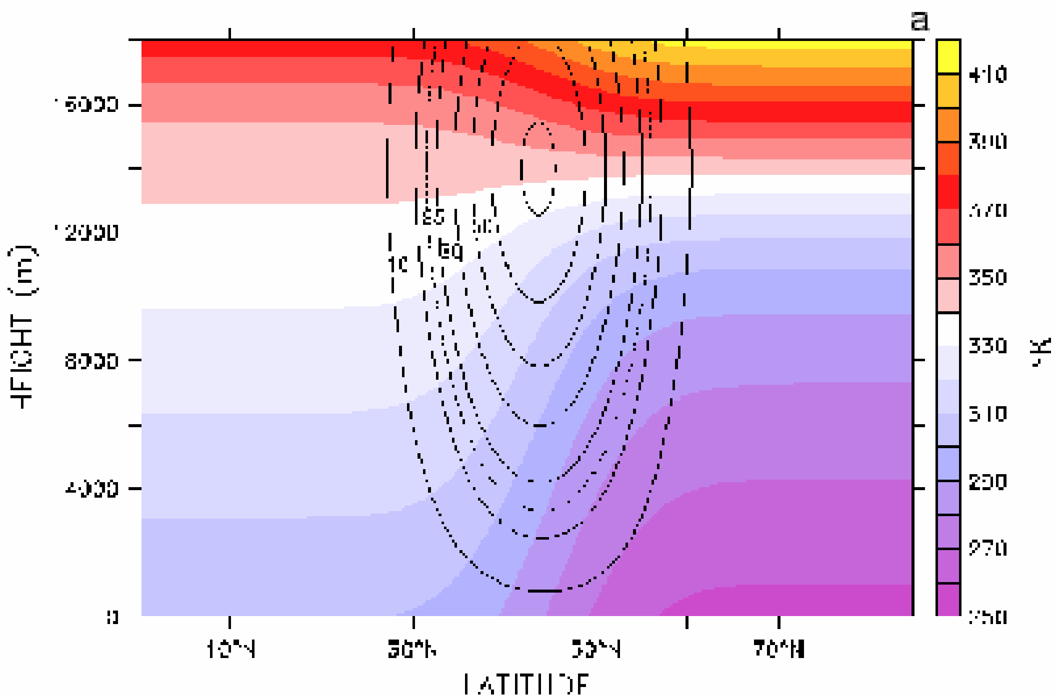


KEZON[L=@AVE]/KEZON[L=@AVE,Y=@MAX]



PV[X=@AVE,L=1:50@AVE]/PV[X=@AVE,L=1:50@AVE,Y=@MAX]

Zonal Mean Flow and Potential Temperature for the Baroclinic Simulations

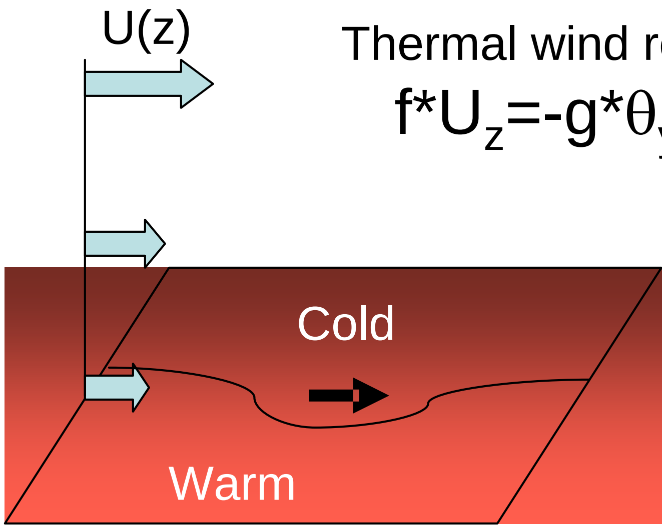


Elements of Baroclinic Instability

Lower level thermal wave

Thermal wind relation

$$f^*U_z=-g^*\theta_y/\Theta$$



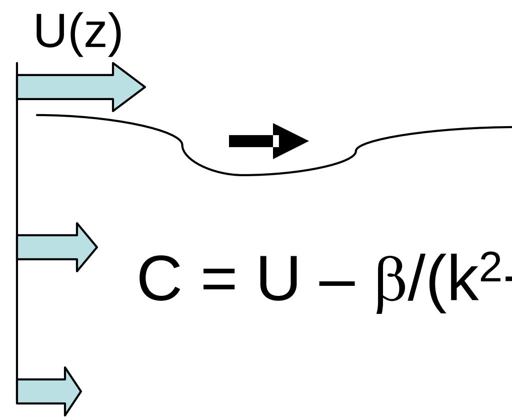
The phase velocity:

$$C=U(z=0)+U_z(z=0)/\gamma$$

The perturbation temperature

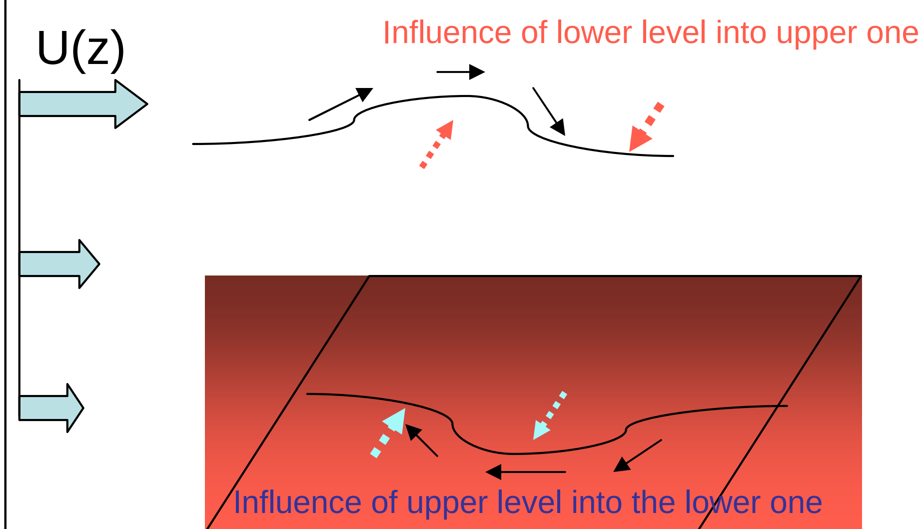
$$q = q_0{}^*\exp[\,ik(x-ct)]^*\exp(-\gamma z)$$

Upper level Rossby wave



$$C = U - \beta/(k^2+\gamma^2)$$

Baroclinic growth



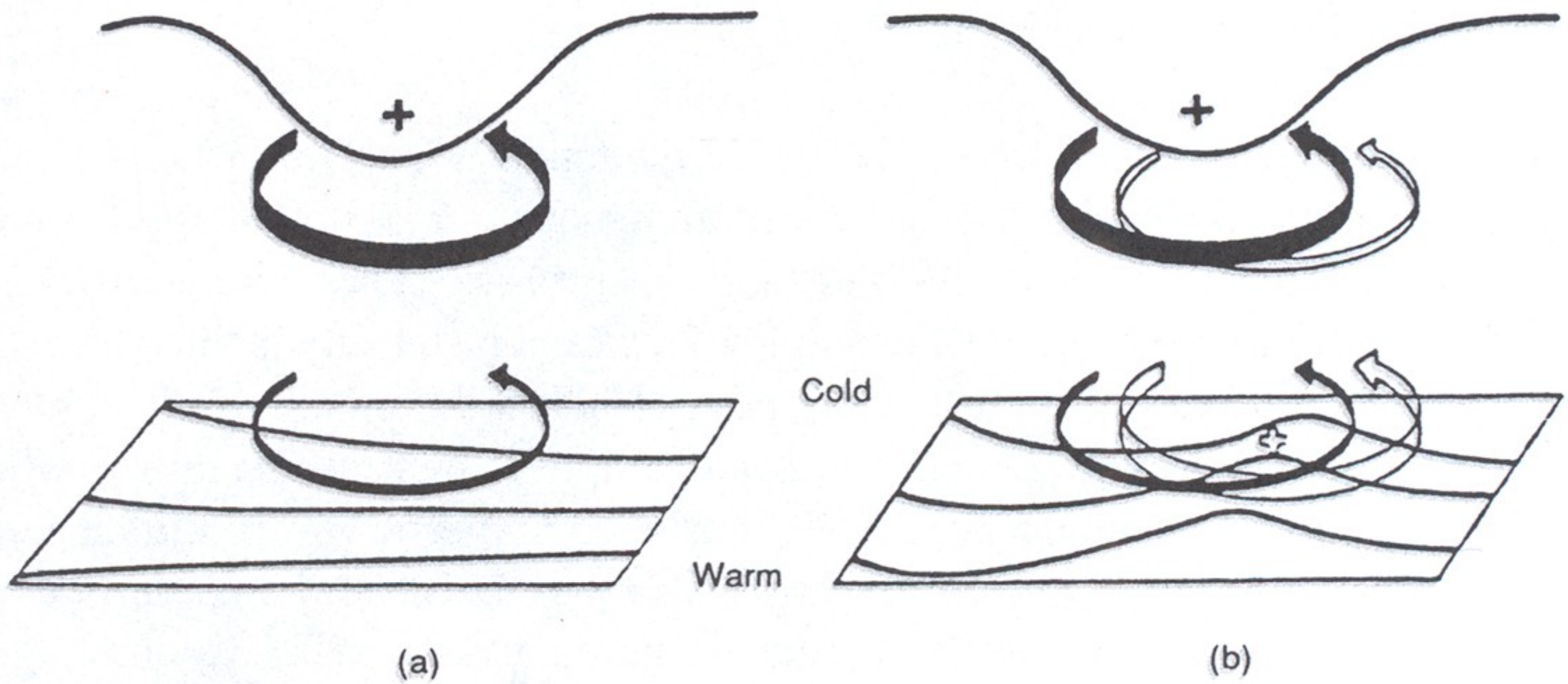
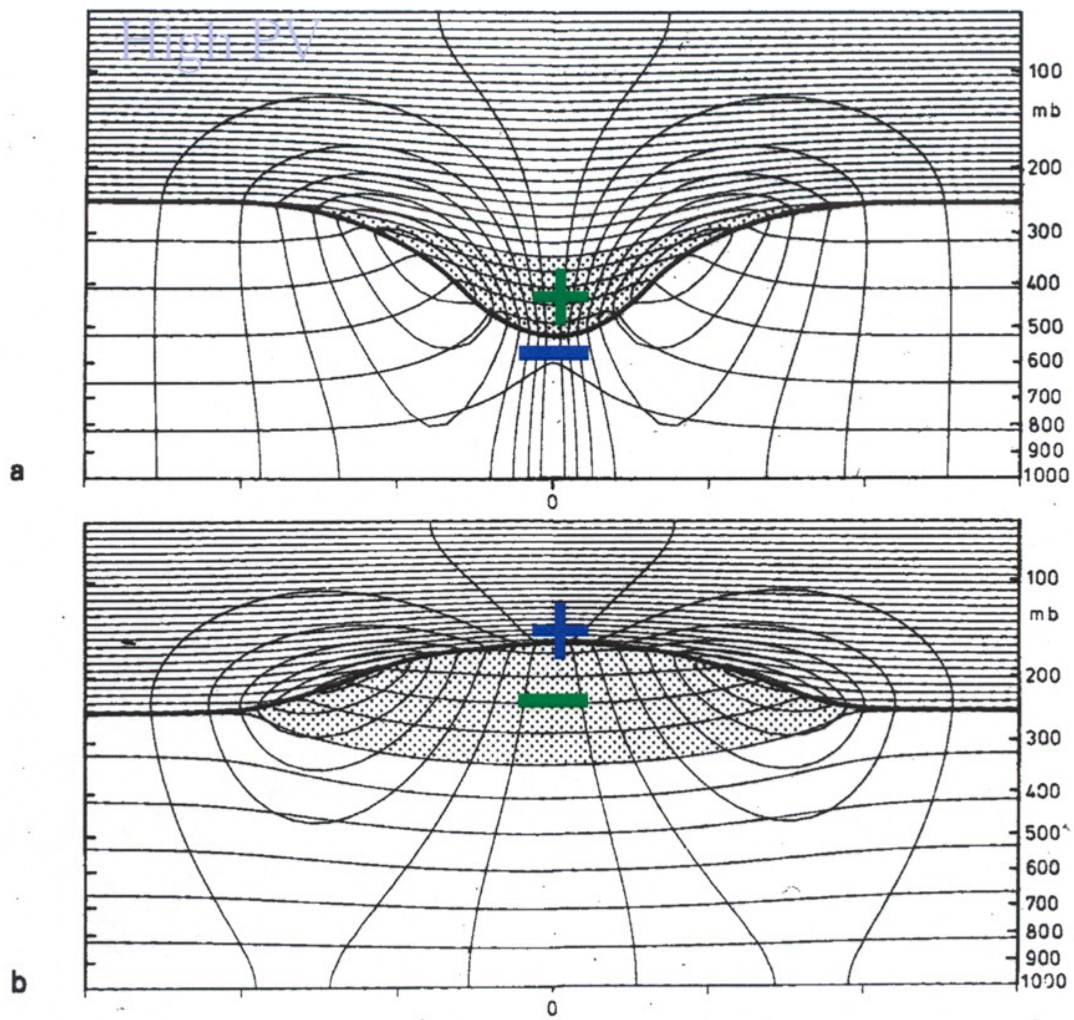


Fig. 8.1 A schematic picture of cyclogenesis associated with the arrival of an upper-level positive vorticity perturbation over a lower-level baroclinic region. (a) Lower-level cyclonic vorticity induced by the upper-level vorticity anomaly. The circulation induced by the vorticity anomaly is shown by the solid arrows, and potential temperature contours are shown at the lower boundary. The advection of potential temperature by the induced lower-level circulation leads to a warm anomaly slightly east of the upper-level vorticity anomaly. This in turn will induce a cyclonic circulation as shown by the open arrows in (b). The induced upper-level circulation will reinforce the original upper-level anomaly and can lead to amplification of the disturbance. (After Hoskins et al., 1985.)

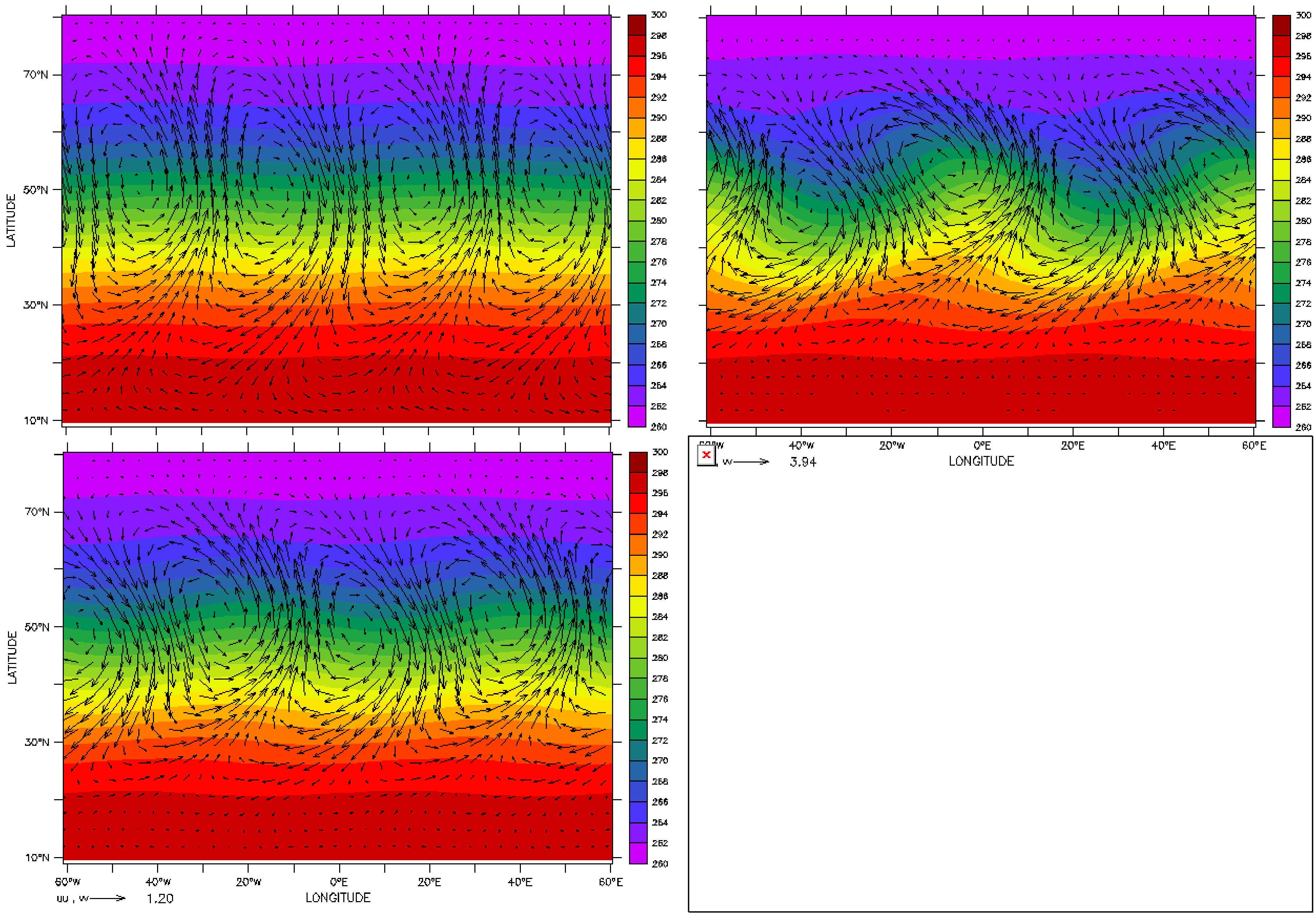


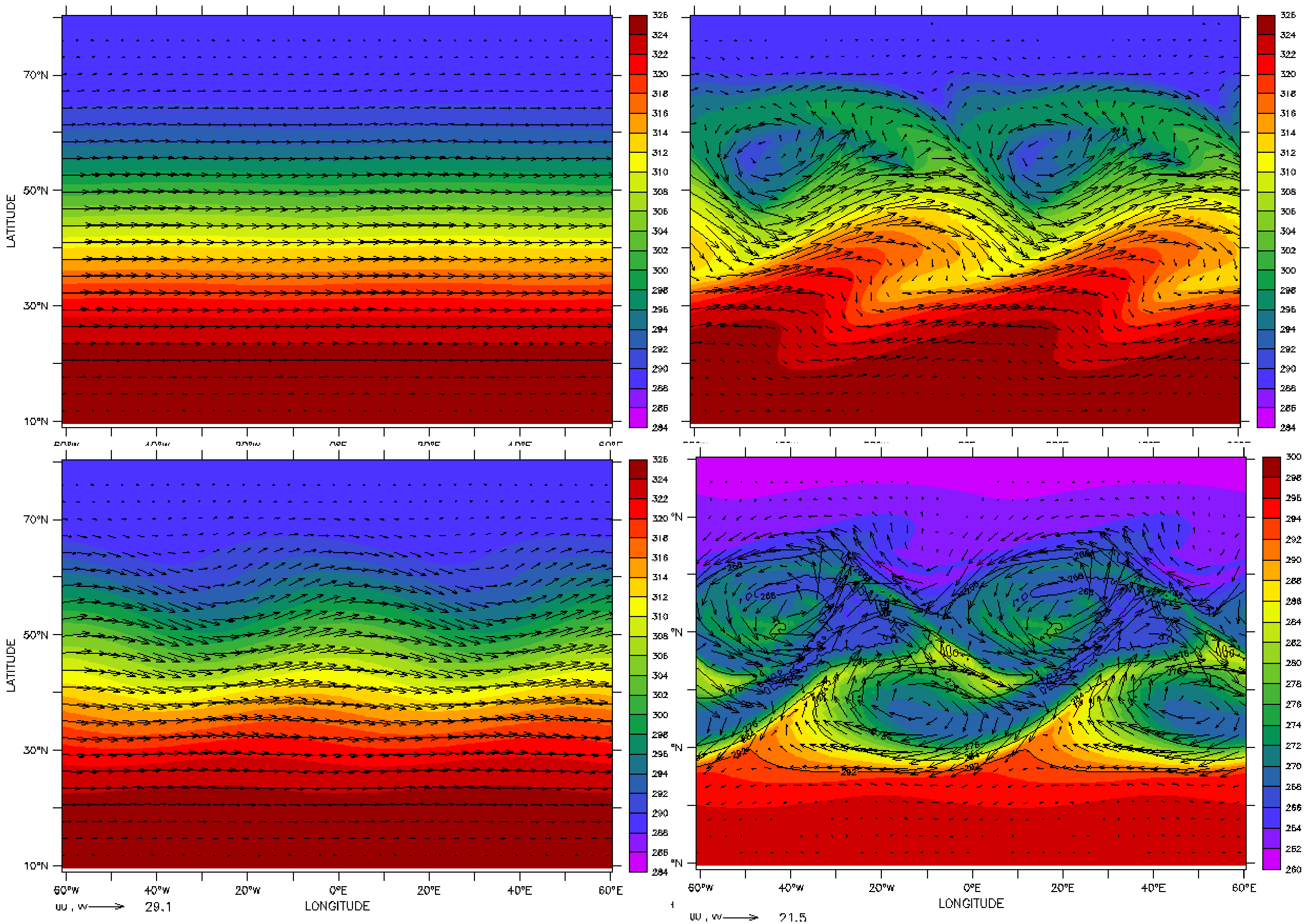
Cut-off cyclone

PV anomaly

PT anomaly

Cut-off anticyclone


LATITUDE
LONGITUDE
70°N
50°N
30°N
10°N
60°W
40°W
20°W
0°E
20°E
40°E
60°E
300
298
296
294
292
290
288
286
284
282
280
278
276
274
272
270
268
266
264
262
260
uu , w 1.20
3.94


70°N
50°N
30°N
10°N
LATITUDE
60°W
40°W
20°W
0°E
20°E
40°E
60°E
LONGITUDE
29.1
21.5
326
324
322
320
318
316
314
312
310
308
306
304
302
300
298
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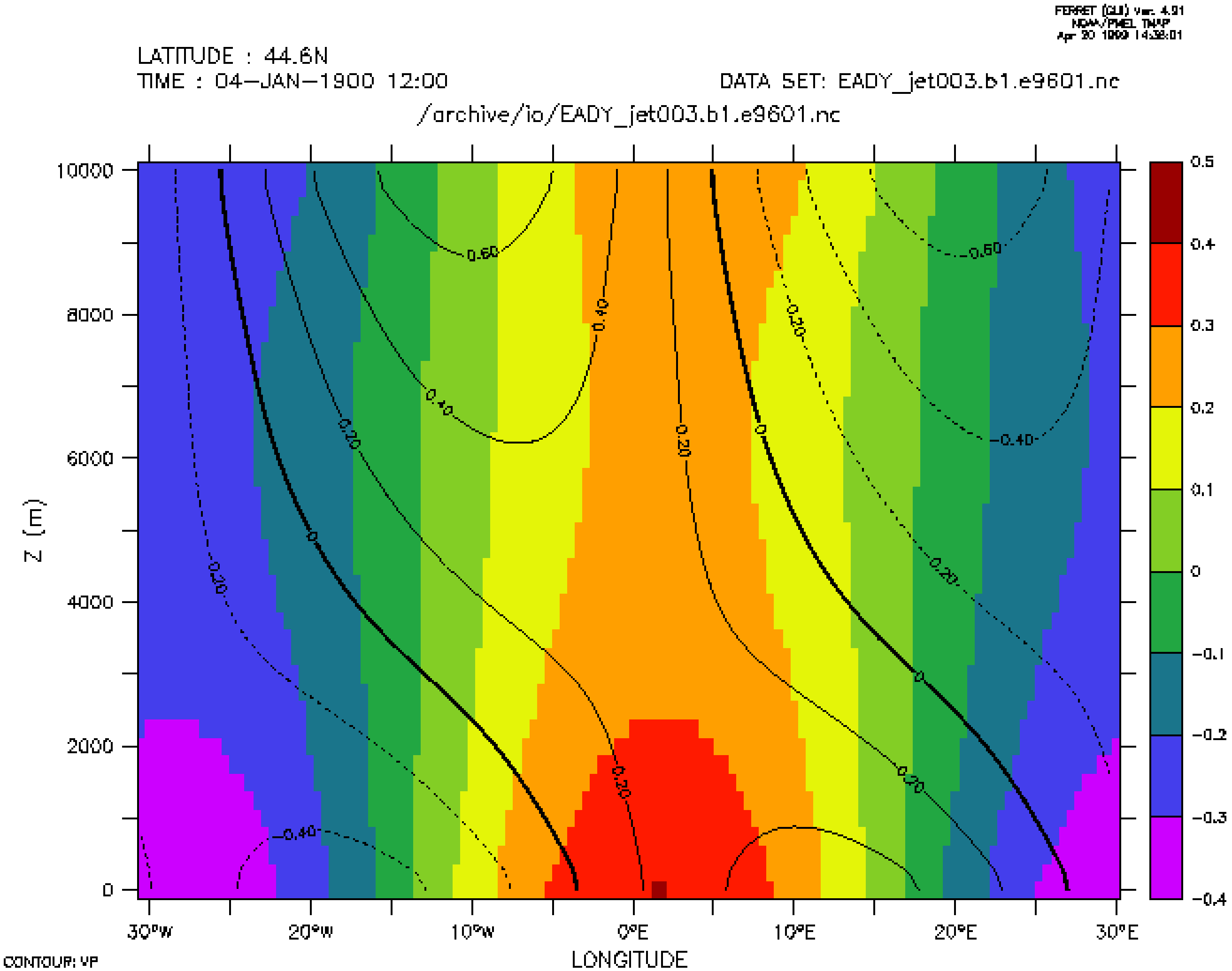
260



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Apr 20 1999 14:38:01
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TIME : 04-JAN-1900 12:00
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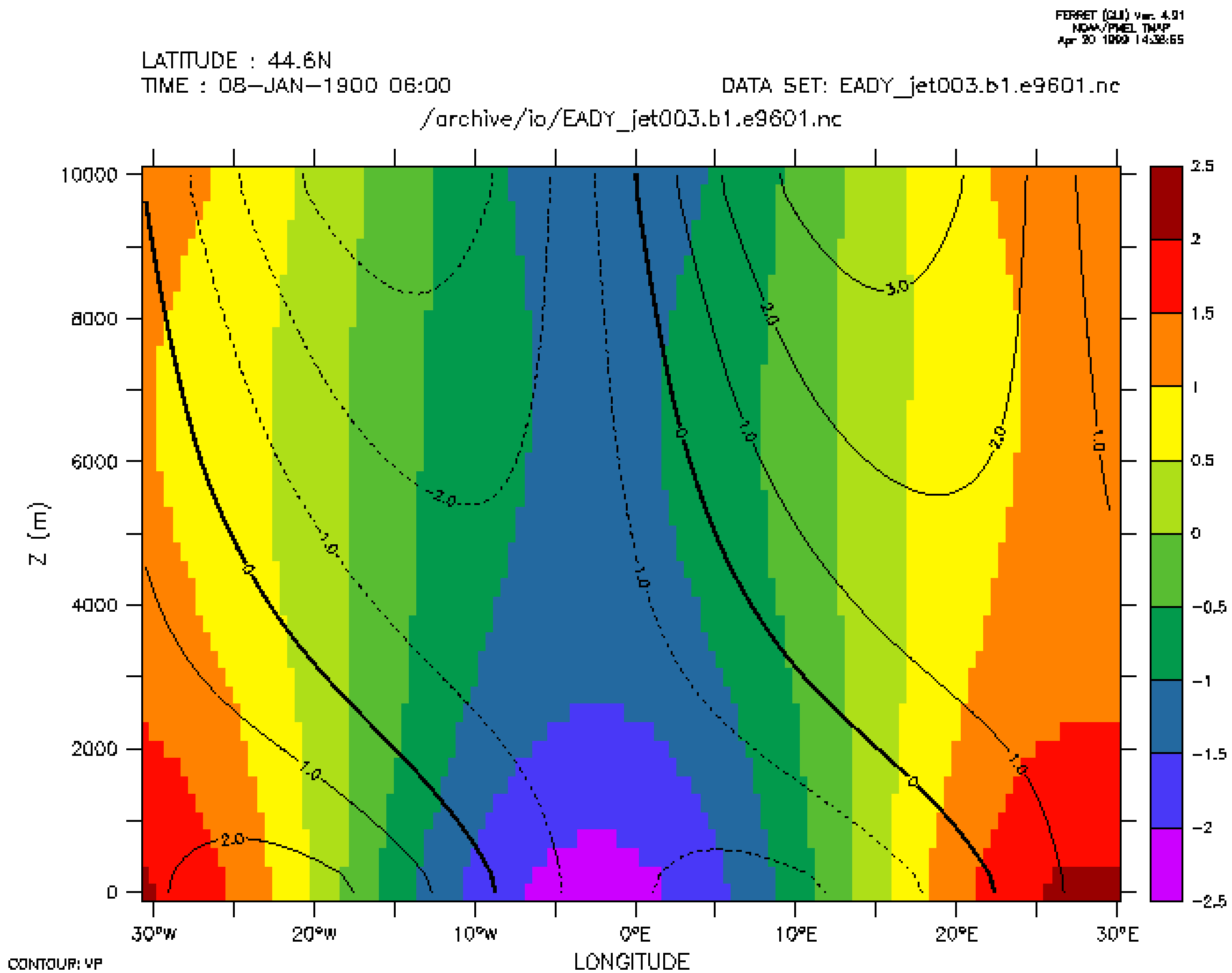
Z (m)
LONGITUDE
CONTOUR: VP

TH−TH[T=0]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NOAA/PMEL TMAP
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TIME : 08-JAN-1900 06:00
DATA SET: EADY_jet003.b1.e9601.nc
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Z (m)
10000
8000
6000
4000
2000
0
30°W
20°W
10°W
0°E
10°E
20°E
30°E
LONGITUDE
CONTOUR: VP
2.5
2
1.5
1
0.5
0
-0.5
-1
-1.5
-2
-2.5

TH−TH[T=0]

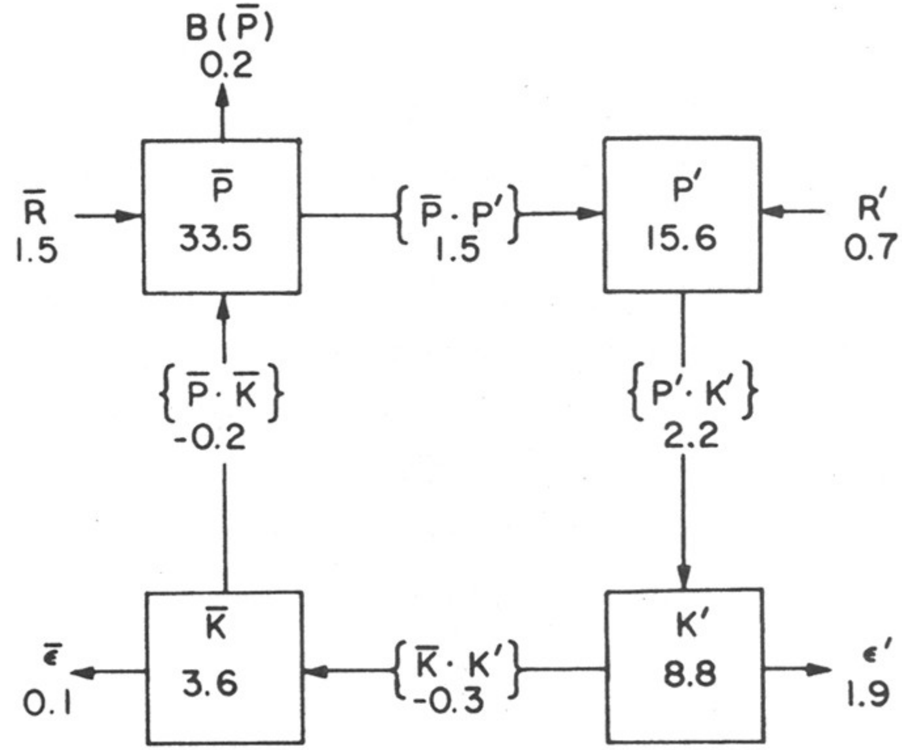


Fig. 10.13 The observed mean energy cycle for the Northern Hemisphere. Numbers in squares are energy amounts in units of $10^5\ \mathrm{J\,m^{-2}}$. Numbers next to arrows are energy transformation rates in units of $\mathrm{W\,m^{-2}}$. $B(\overline{p})$ represents a net energy flux into the Southern Hemisphere. Other symbols are defined in the text. (Adapted from Oort and Peixoto, 1974.)

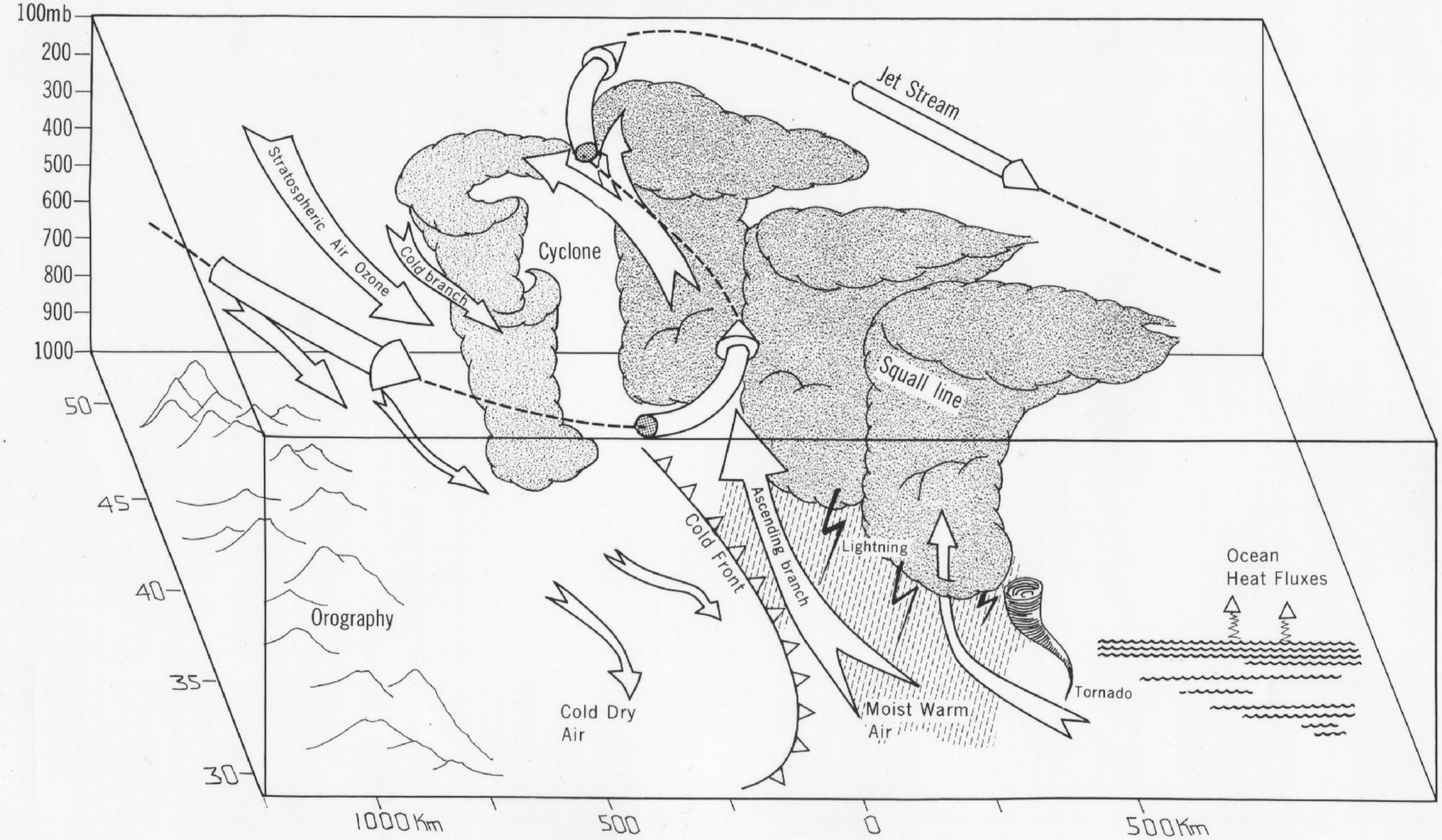
Baroclinic effects:

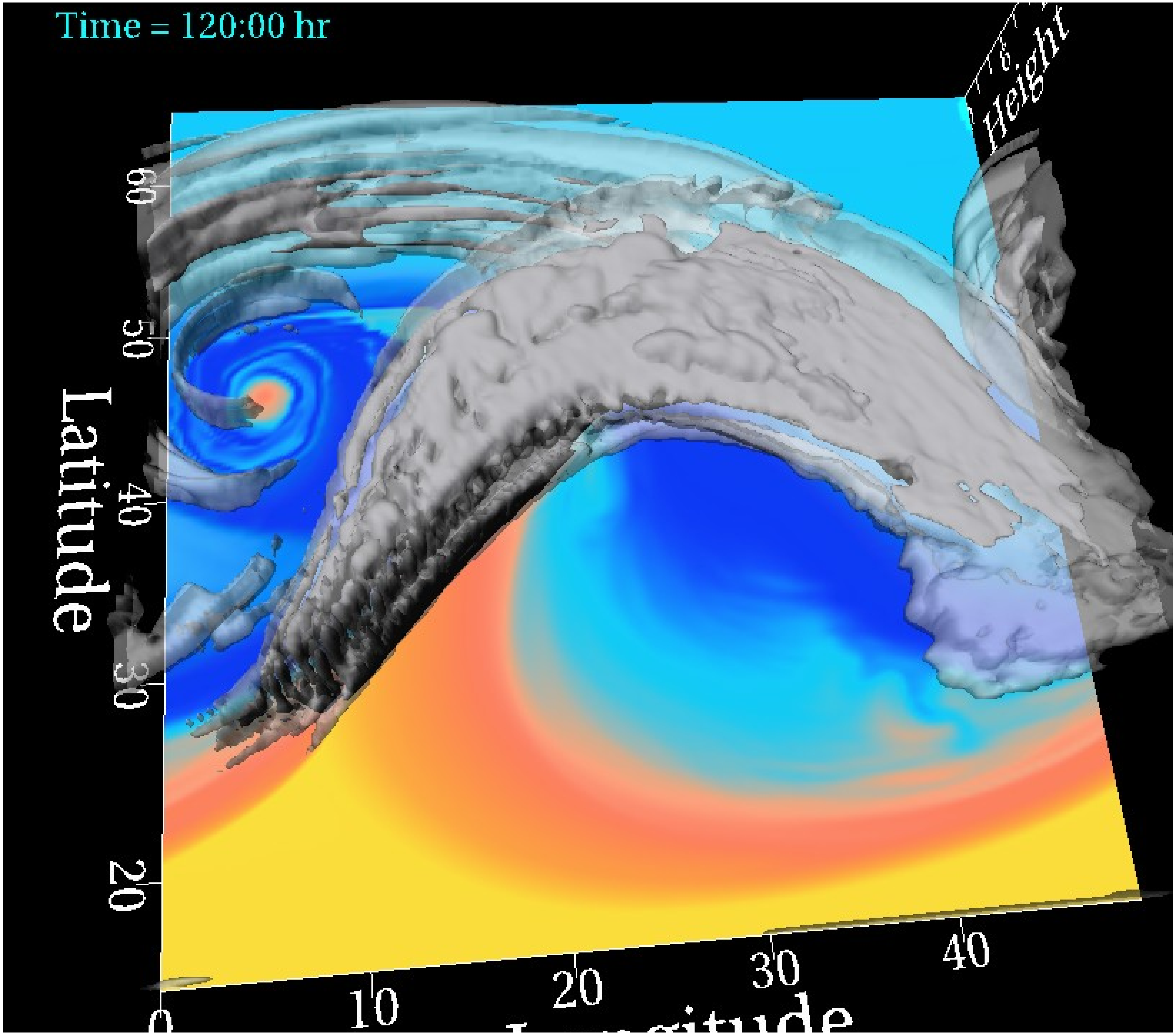
Atmospheric Fronts

Transfer of momentum and heat poleward

Breaking upper and lower waves

EXTRA TROPIC STORMS

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Height
Latitude
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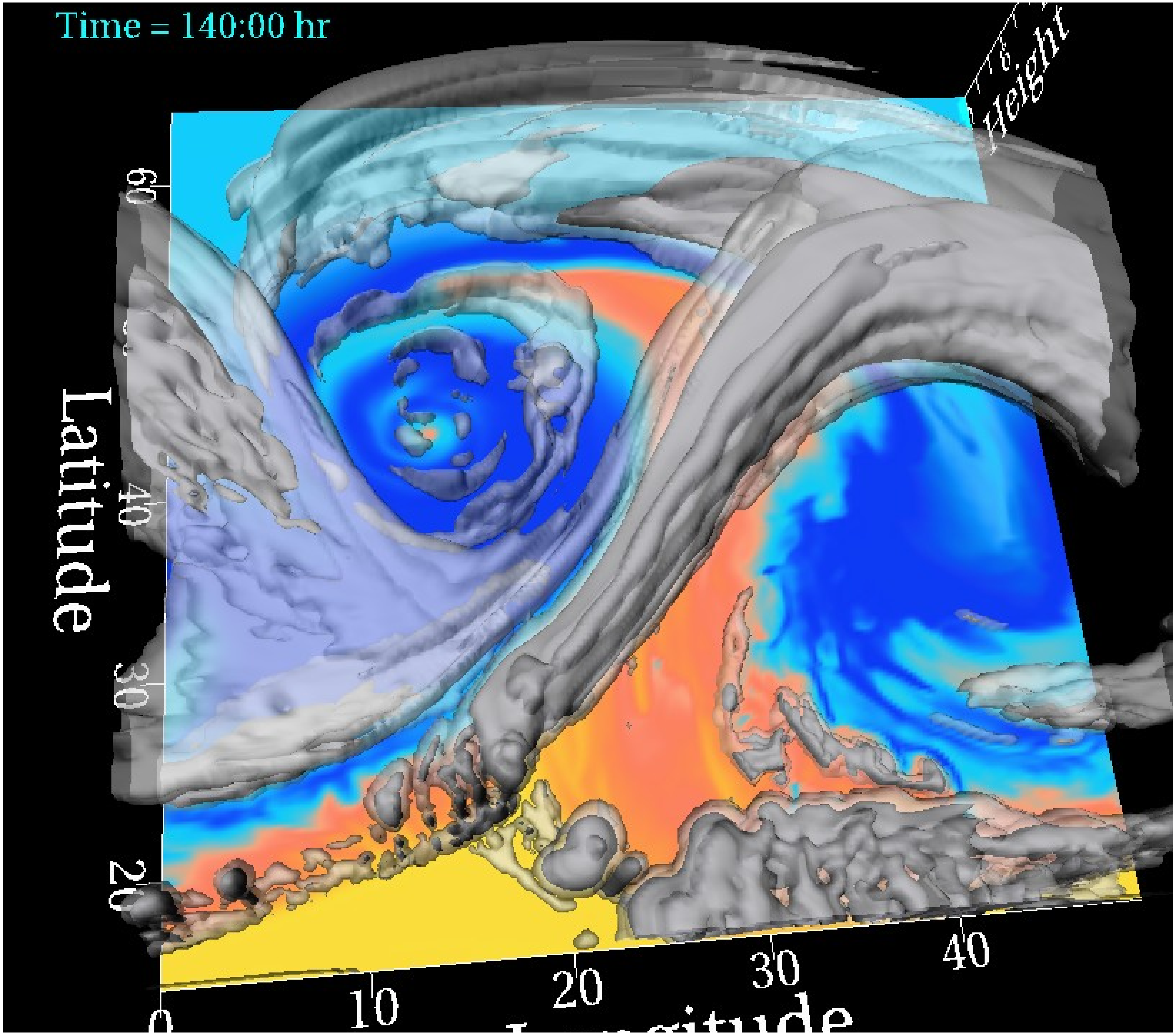
50
40
30
20
0
10
20
30
40


Time = 140:00 hr
Height
Latitude
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40
30
20
0
10
20
30
40
Longitude

Potential Vorticity of the Shallow Water Global Model
for three different amplitude of the forcing

$$F(\lambda, \phi, t) = W\frac{(U(\phi) - c)}{U_{max}}G(t)\cos(m\lambda - ct)$$

Where U(f) is the initial jet, G(t) is a gaussian in time multiplied by t2 it grows and decays in 8 days. C= 0.6 Umax Umax= 40m/s. m the zonal wave number this case is m=7.

Weak Anticyclonic Wavebreaking

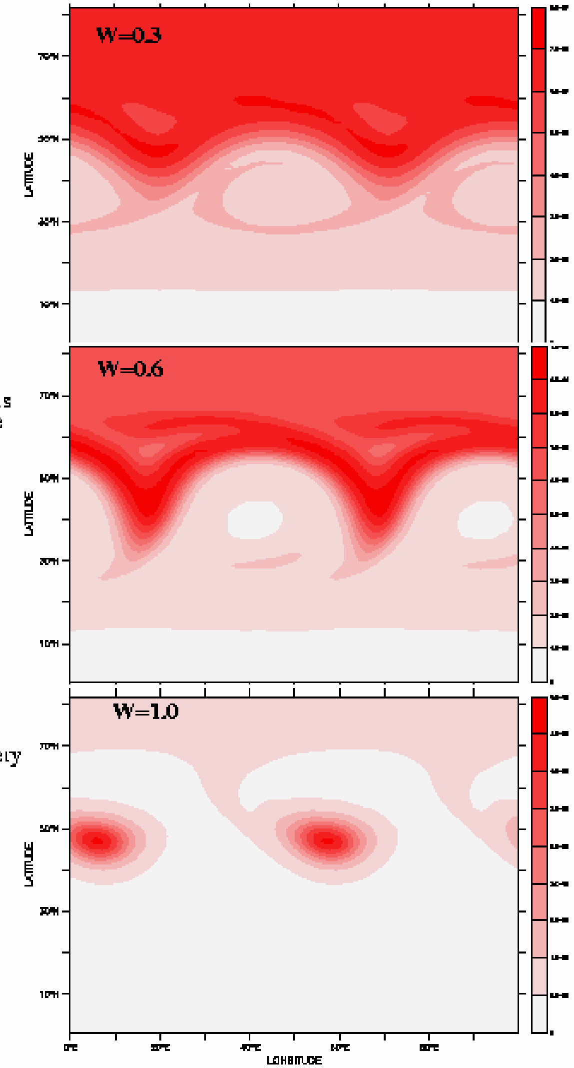
Equatorial rossby wave propagation absorbed by the critical layer due to the mean flow.

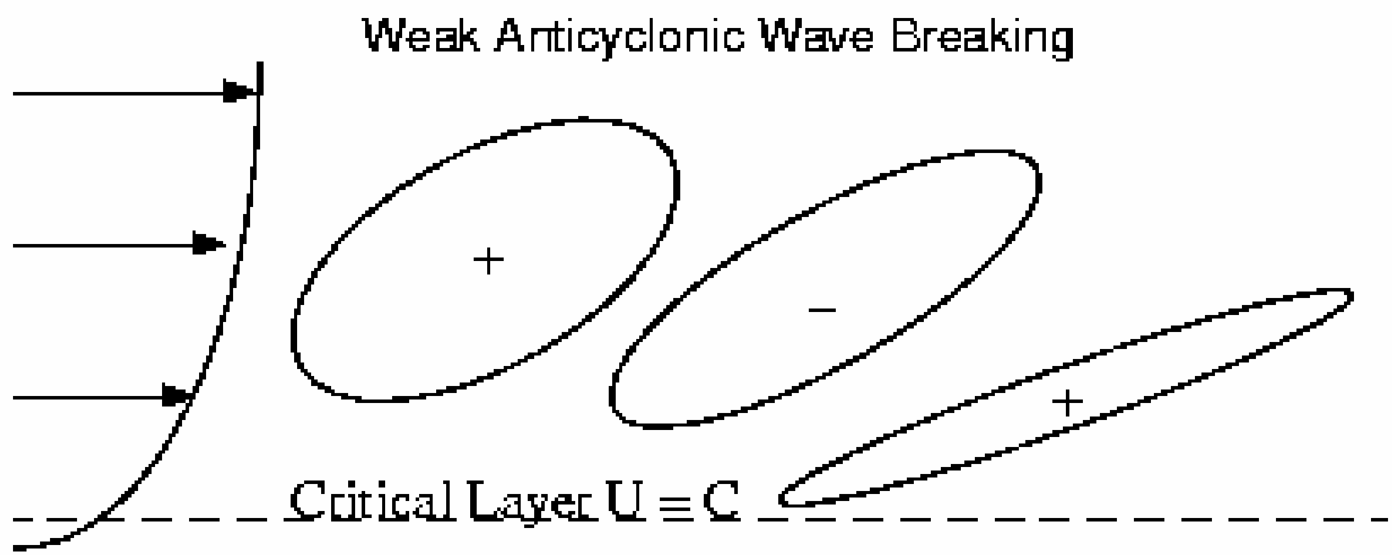
Strong Anticyclonic Wavebreaking

For intense vortices, if the anticyclone is strong enough it can pulls and shred the cyclonic centers.

Strong Cyclonic Wavebreaking

However, if the cyclone centers are very strong the they can pull and shred the the anticyclonic vorticity.

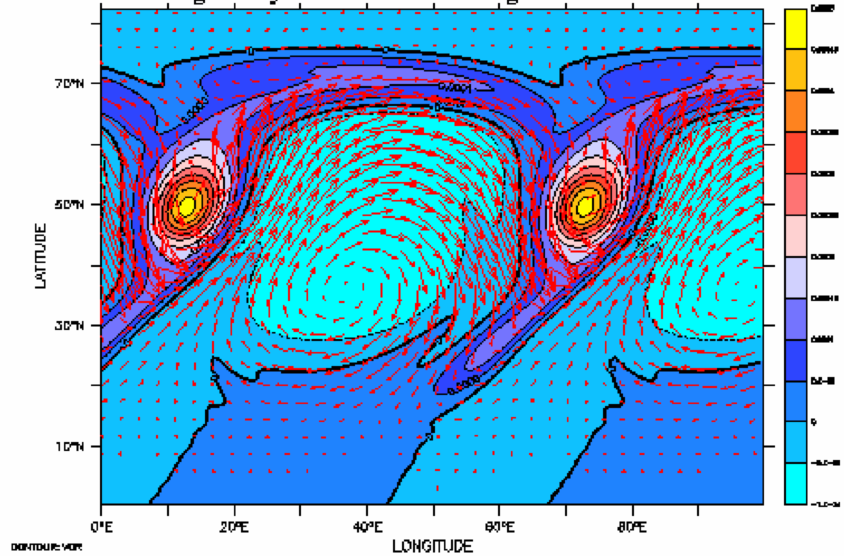




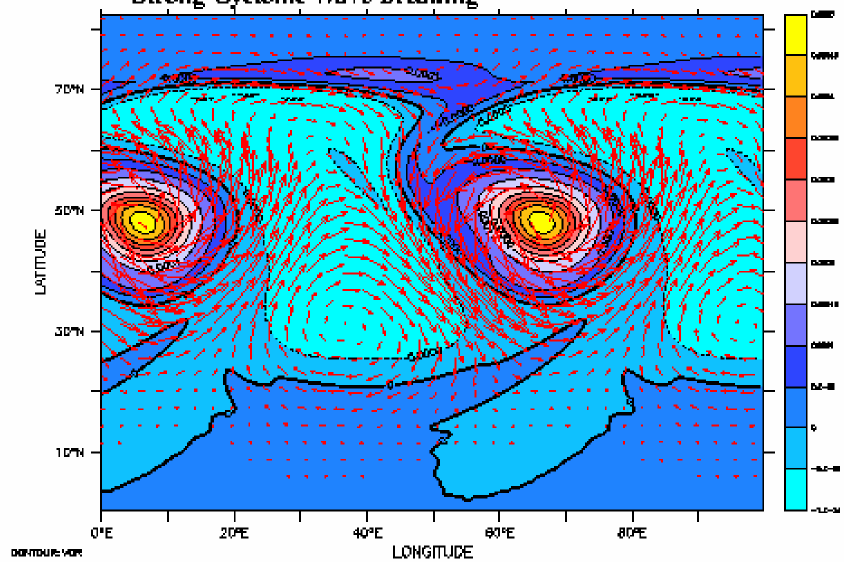
Weak Anticyclonic Wave Breaking
+
−
+
Critical Layer U = C

Vortex interactions in the life-cycle of upper level eddies

Strong Anticyclonic Wave Breaking



Strong Cyclonic Wave Breaking

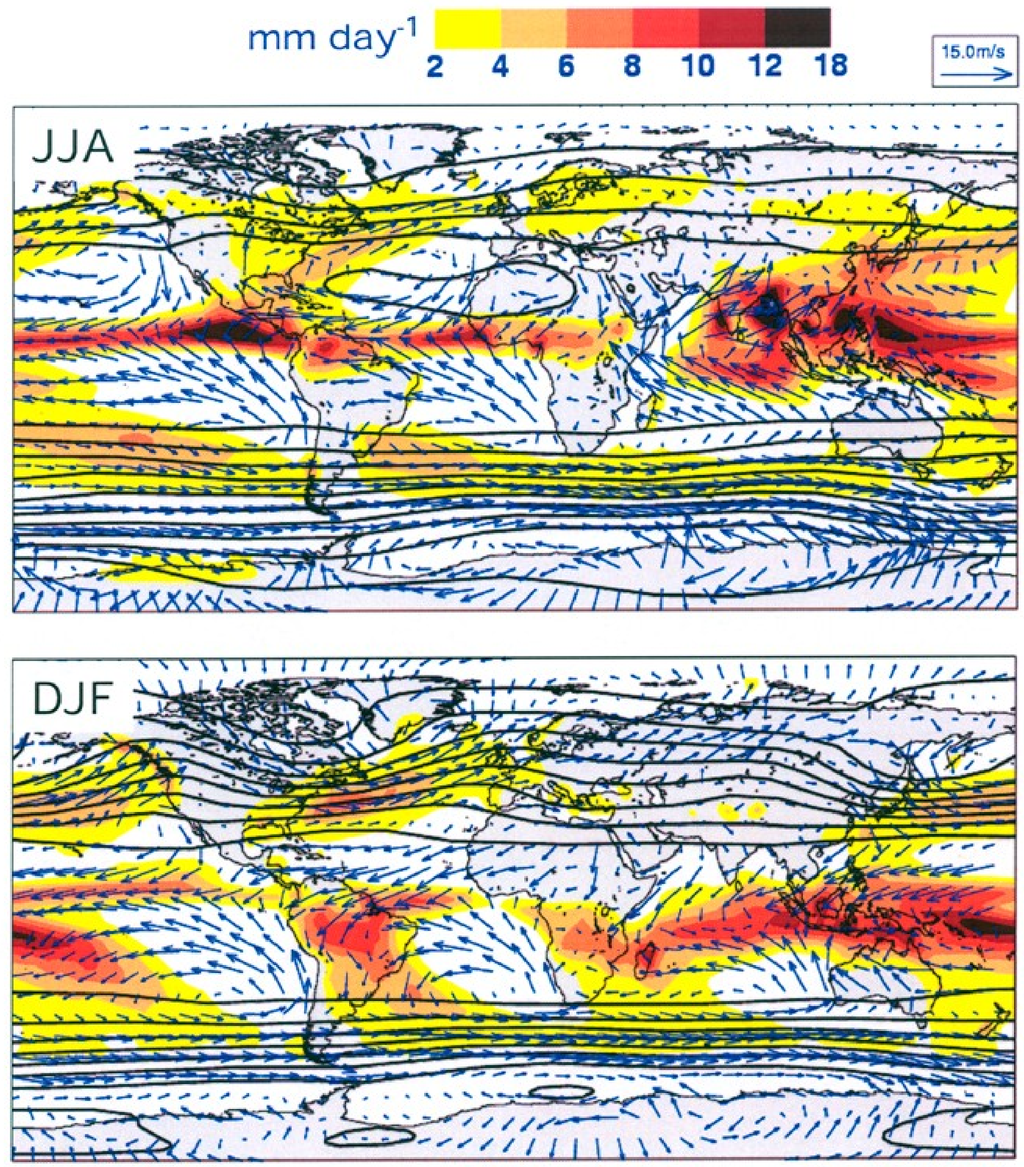


Collective effects of baroclinic waves (daily weather)

Produce

Storm tracks and quasi stationary patterns affecting the basic flow (westerlies)

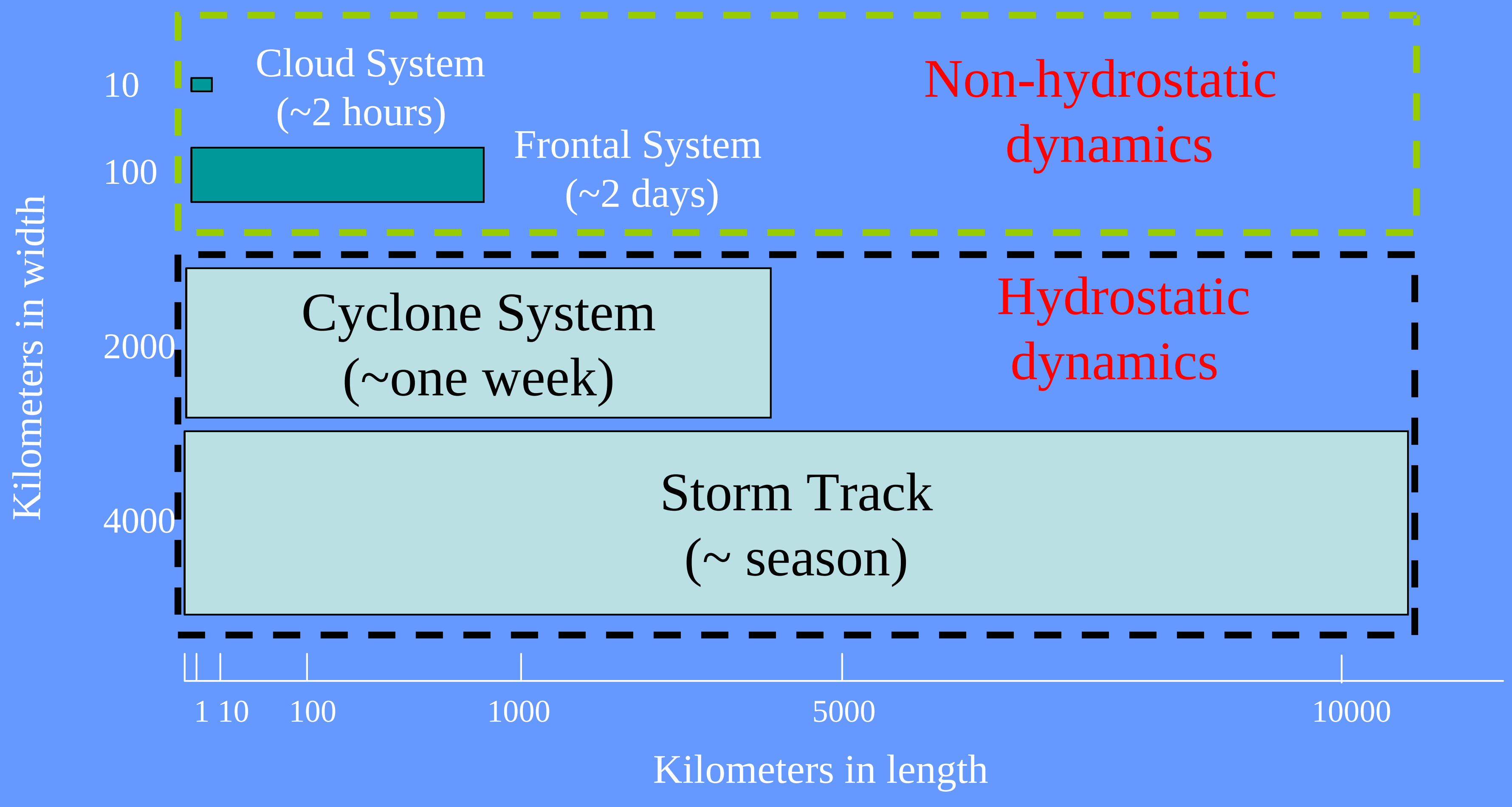
Observed Precipitation $\underline{V}_{925}$ and Z_{500}

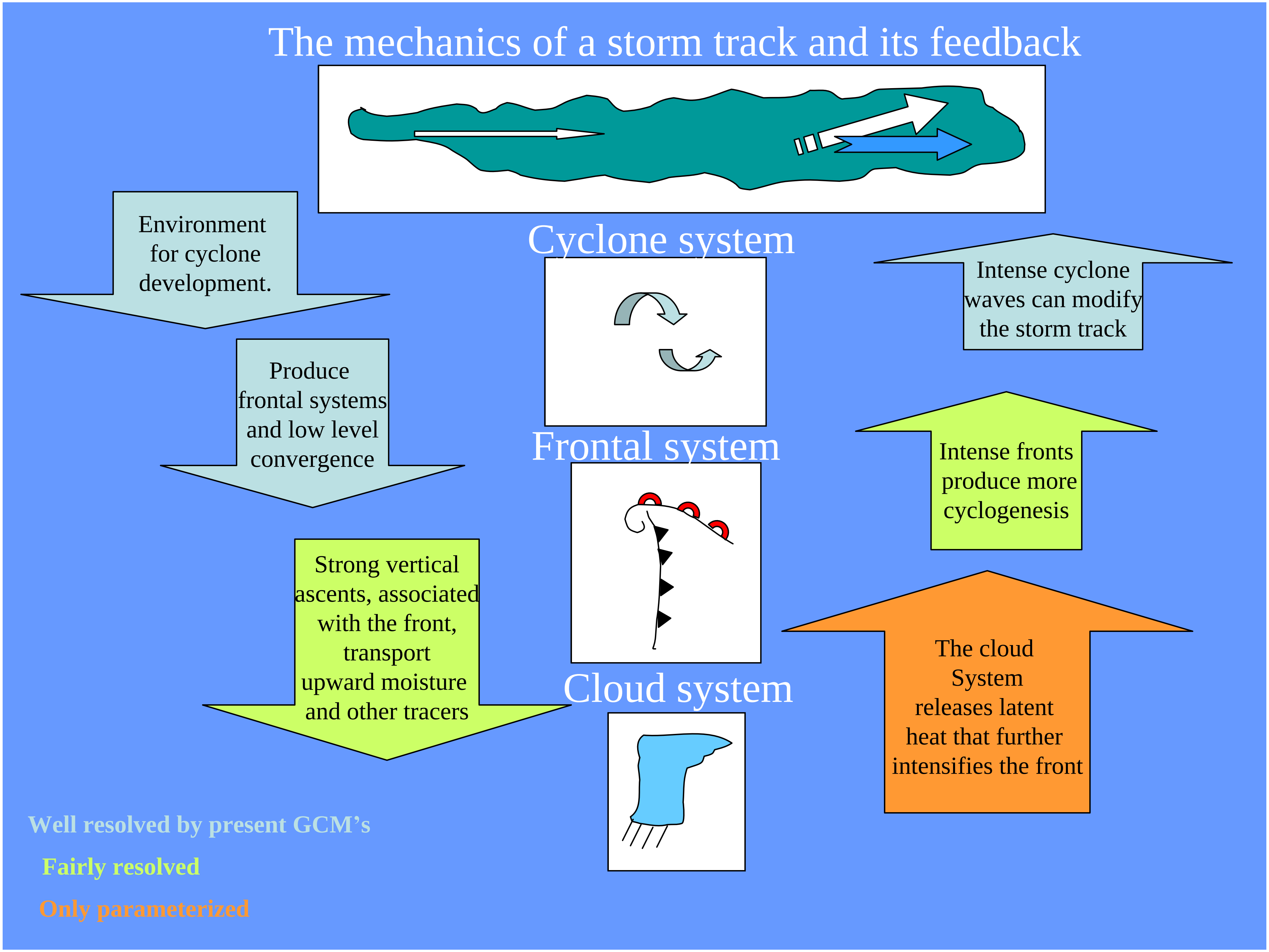


Precipitation: Xie-Arkin 1979/80 – 1998/99, $\underline{V}_{925}$: ERA40 1962-01, Z_{500}: ERA40 1962-01, CI=10 dam

Mark Rodwell

Time-space scale of atmospheric systems




The mechanics of a storm track and its feedback
Cyclone system
Frontal system
Cloud system
Environment for cyclone development.
Produce frontal systems and low level convergence
Strong vertical ascents, associated with the front, transport upward moisture and other tracers
Intense cyclone waves can modify the storm track
Intense fronts produce more cyclogenesis
The cloud System releases latent heat that further intensifies the front
Well resolved by present GCM's
Fairly resolved
Only parameterized

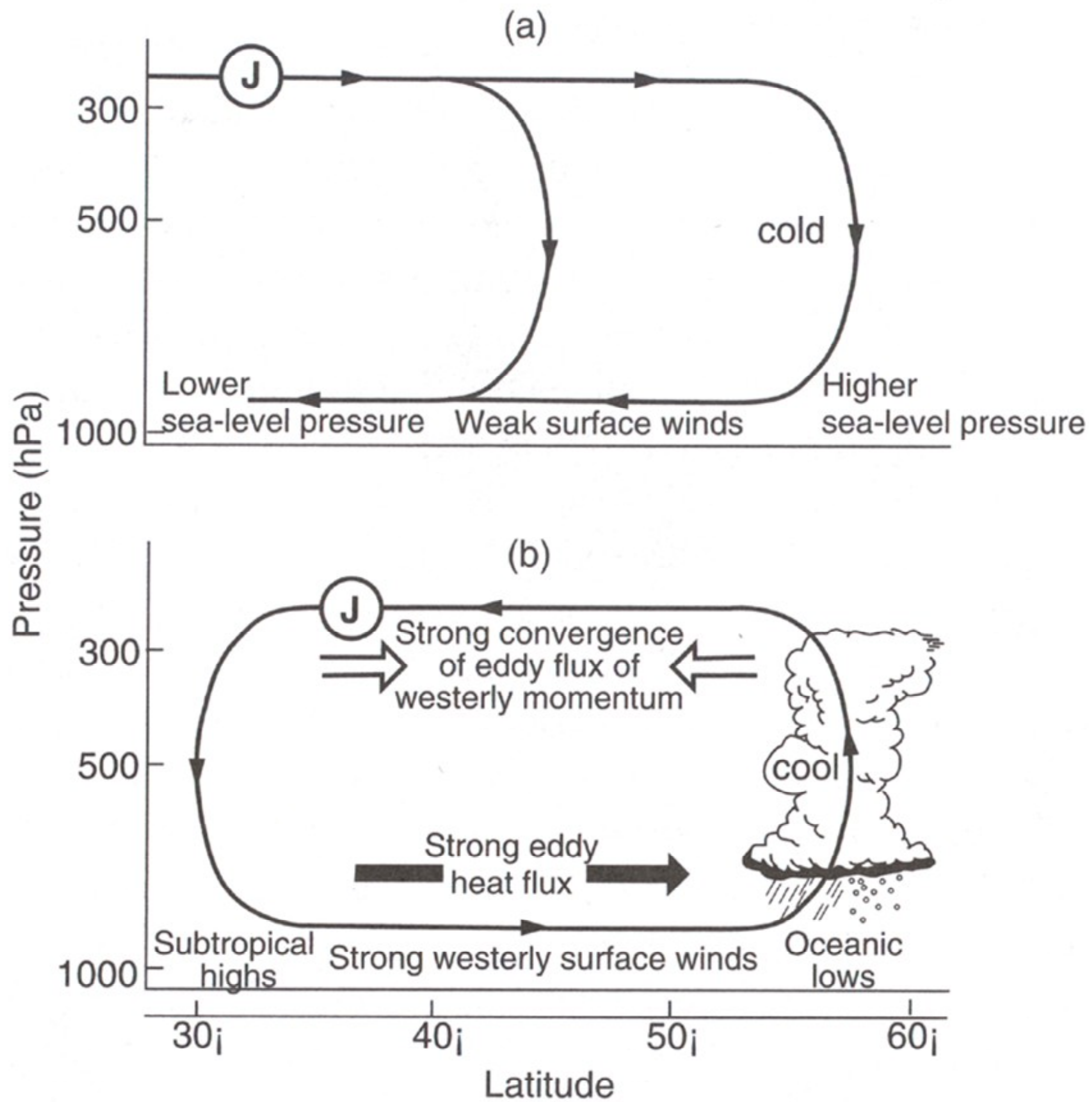
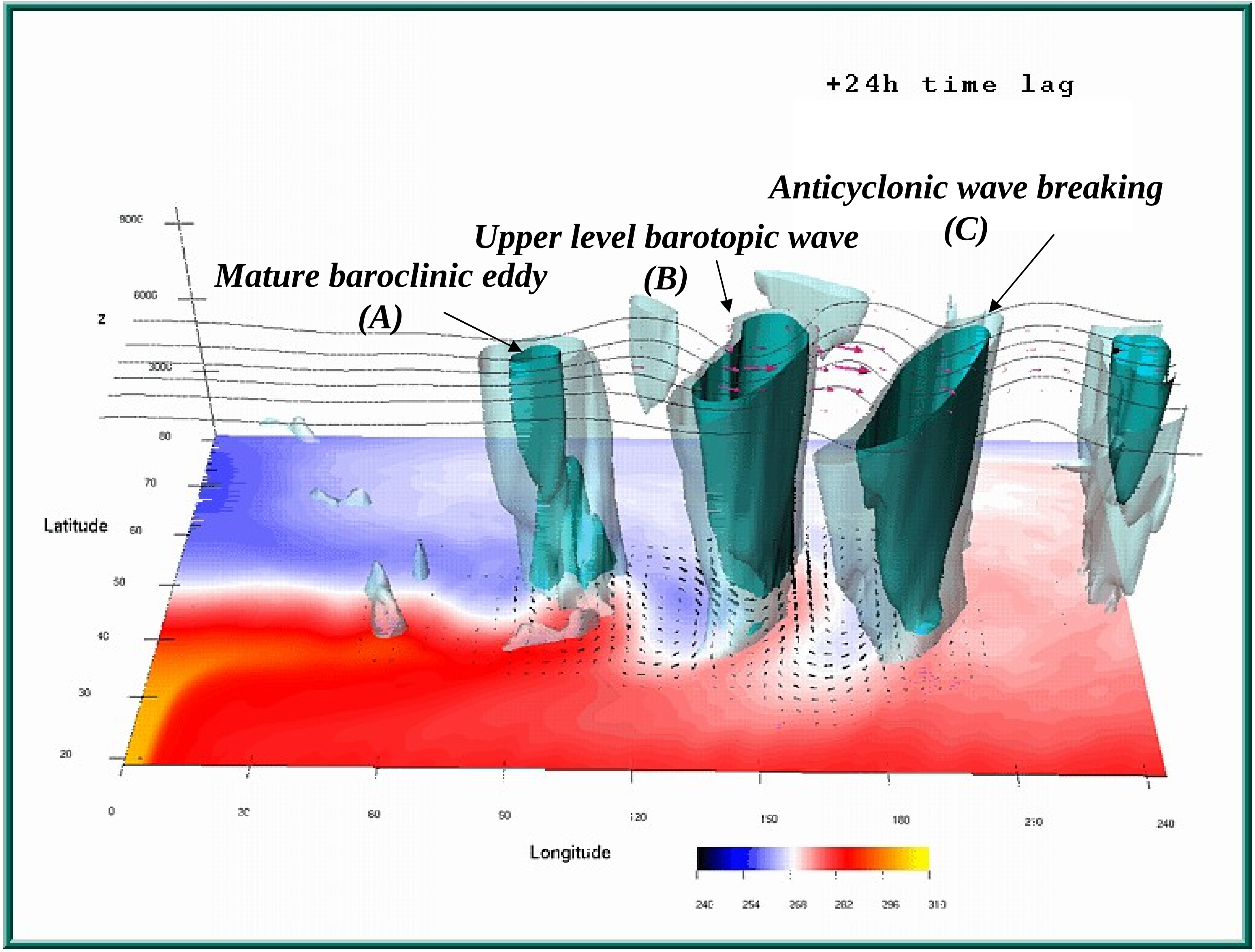


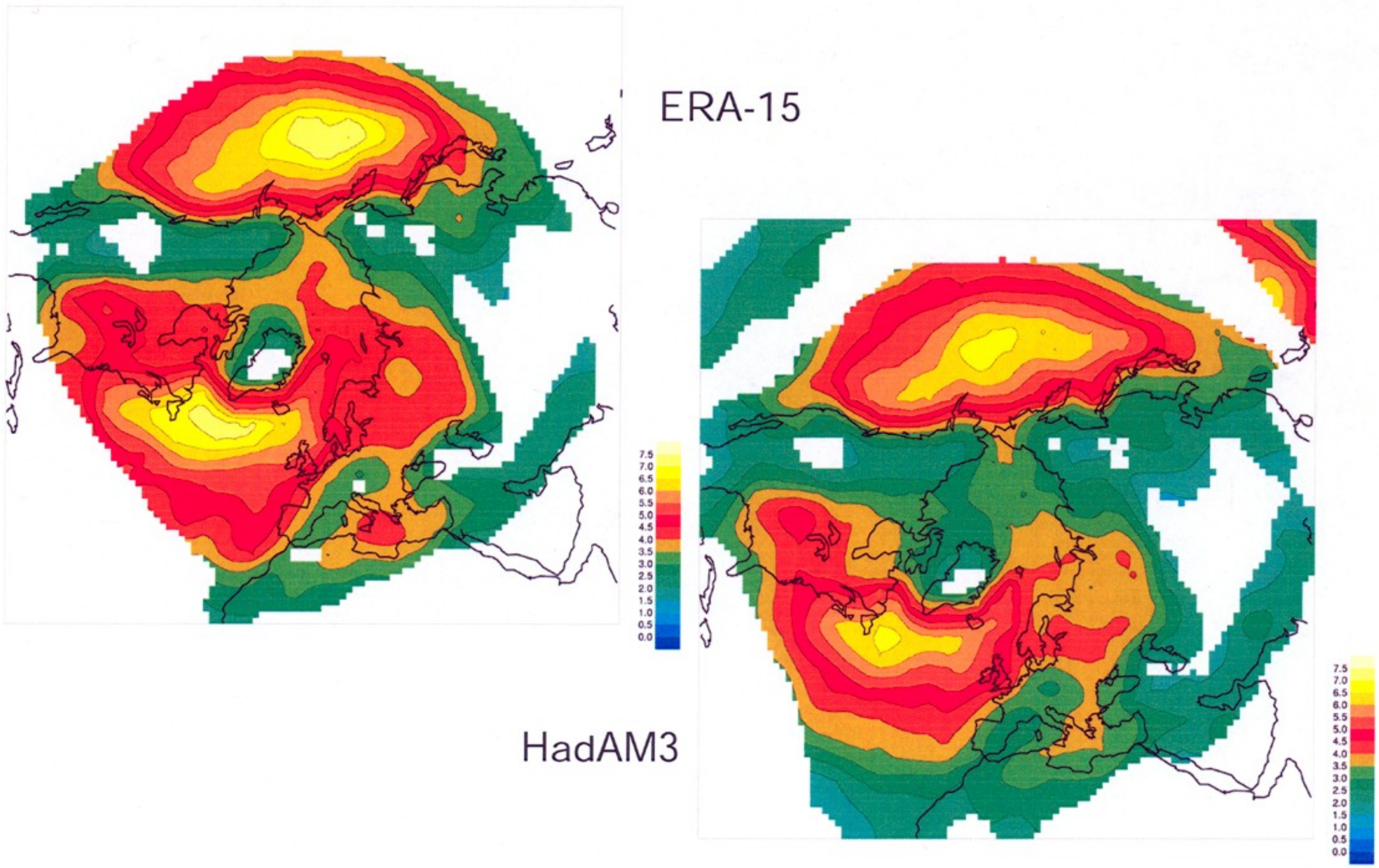
Fig. 10.16 Meridional cross sections showing the relationship between the time mean secondary meridional circulation (continuous thin lines with arrows) and the jet stream (denoted by J) at locations (a) upstream and (b) downstream from the jet stream cores. (After Blackmon et al., 1977. Reproduced with permission of the American Meteorological Society.)

Surfaces of relative vorticity as inferred from regression analysis (-24h, 0h, +24h)



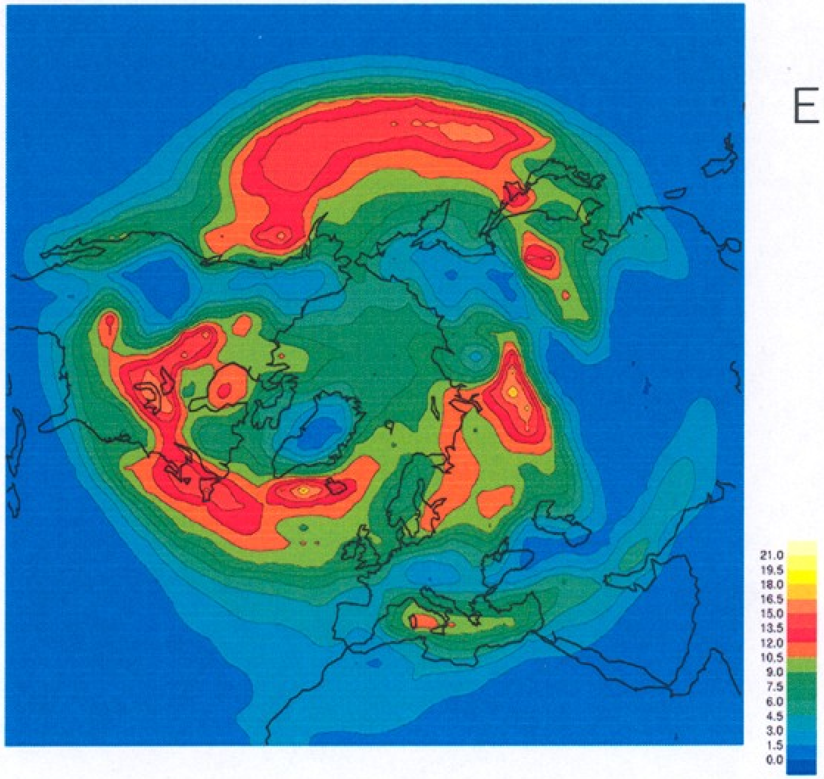
I. Orlanski and B. Gross: Baroclinic lifecycles in a storm track environment. JAS 2000

Mean intensity of the cyclones for DJF: 1979-95

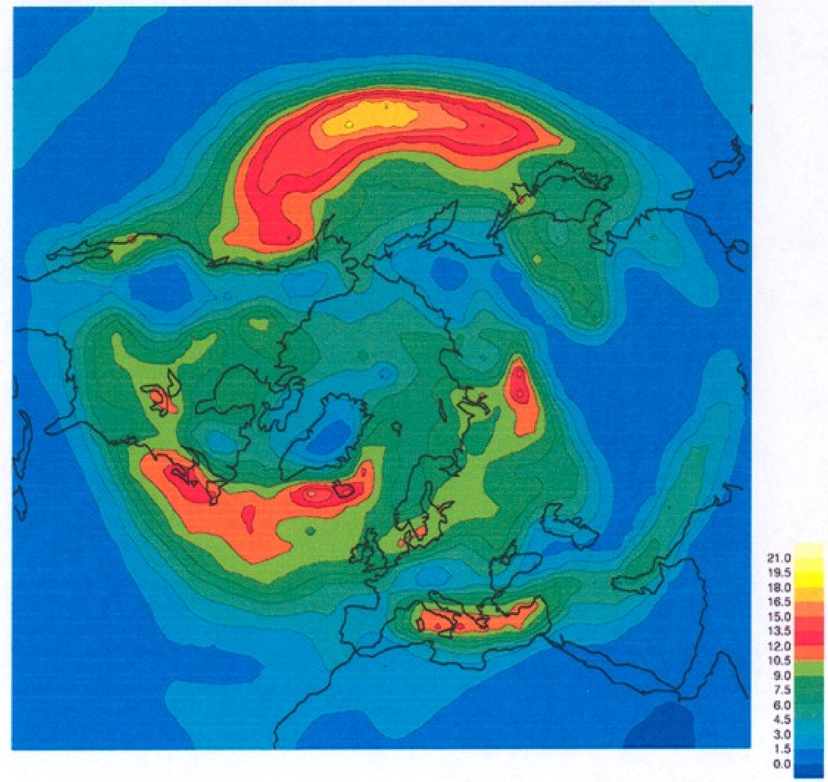


Patterns quite well simulated, although peak intensity generally underestimated in HadAM3. Resolution?

Mean track density for DJF: 1979-95



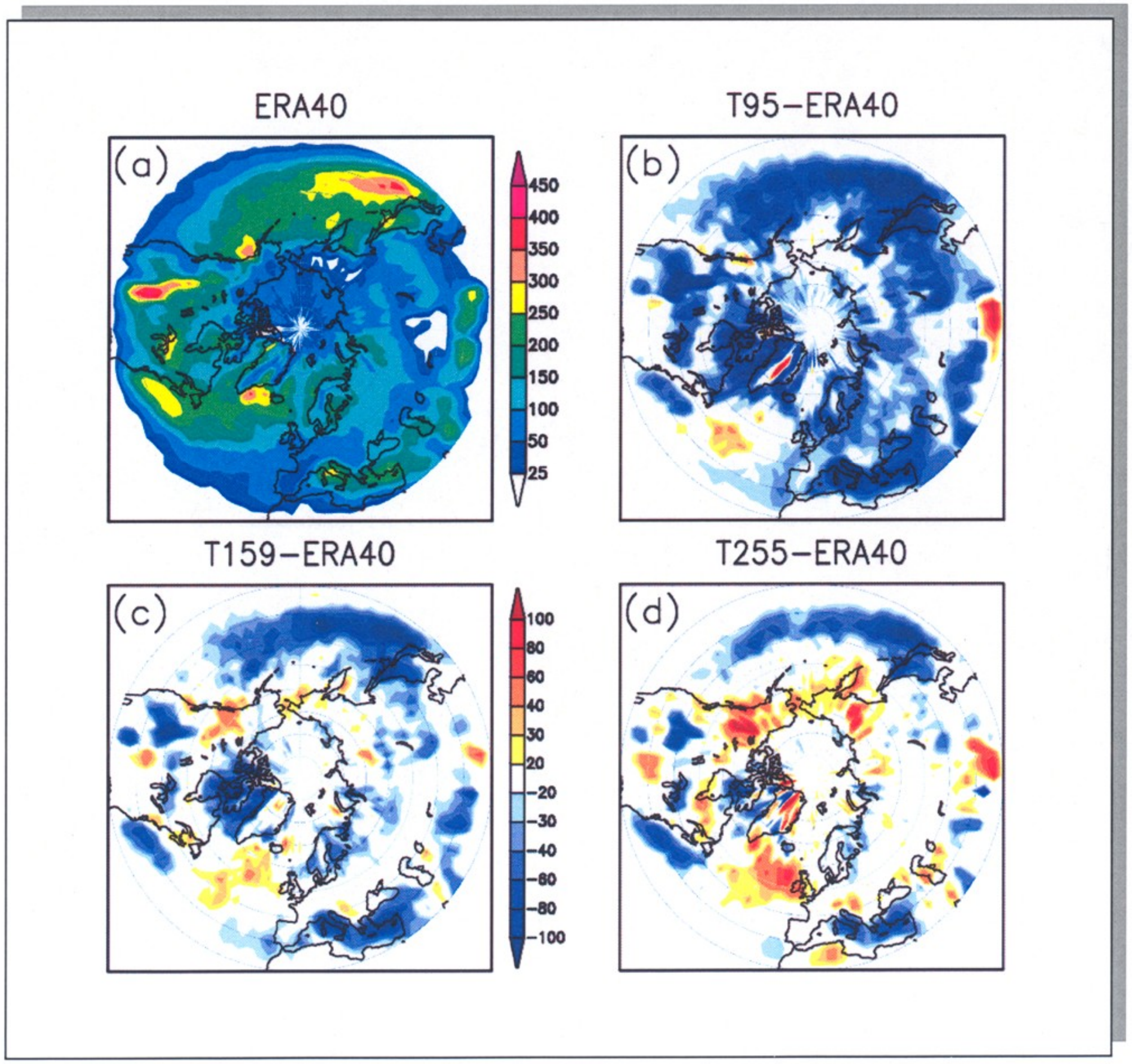
ERA-15



HadAM3

Pacific storm track is too strong. Atlantic storm track too weak in HadAM3, possibly related to lack of systems coming off the eastern side of the Rockies.

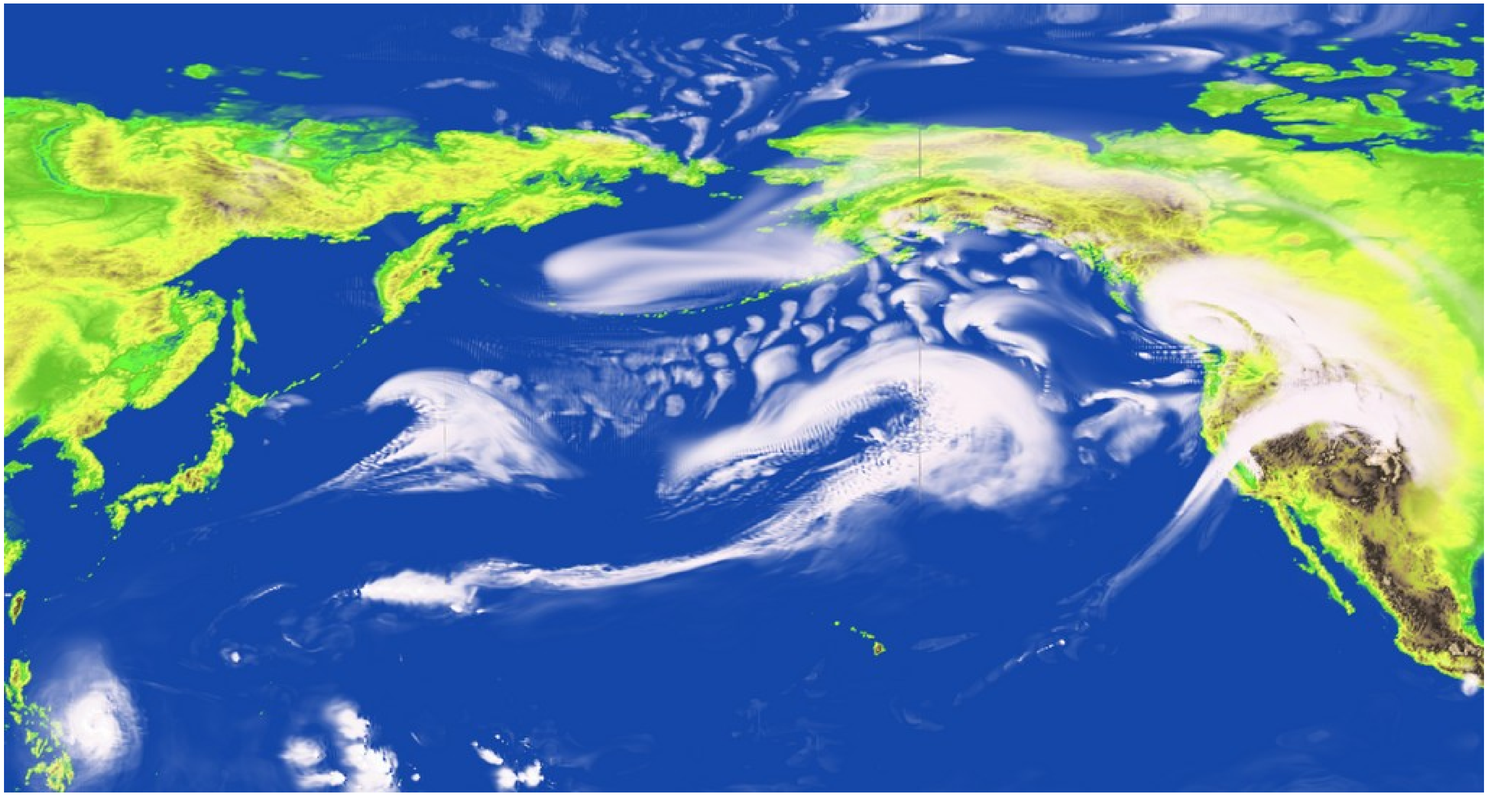
Number of Cyclones DJFM 1982-2001



Experiments on Storm Track variability

A high resolution (9km and 18km) cloud resolving non-hydrostatic model (ZETANC*) was used to perform several idealized storm track simulations. The solution run for 220days and sensitivity to imposed SST were performed. The animation shows the column liquid water content.

The frontal rain band ahead of cold fronts displays a variety of cloud systems, from deep convective clouds in sub tropical latitudes to more stable stratiform clouds in the middle latitudes.



Things to remember from Lecture 2.

- Planetary waves are an essential part of the daily weather and climate.
- Land-sea contrast, topographic features etc can force planetary waves, Rossby waves.
- Its characteristics are; they propagate to the west and radiate energy to the east.
- Also these waves could be unstable, they can grow to large amplitudes from very small perturbations.
- The major planetary wave generation are due to barotropic and baroclnic instabilities. They are responsible for producing the daily weather. Most the time they transport heat and momentum to higher latitudes. Warming the polar regions and removing heat from the subtropics.
- The assemble of all these waves tend to produce wave active regions which we called the storm tracks.