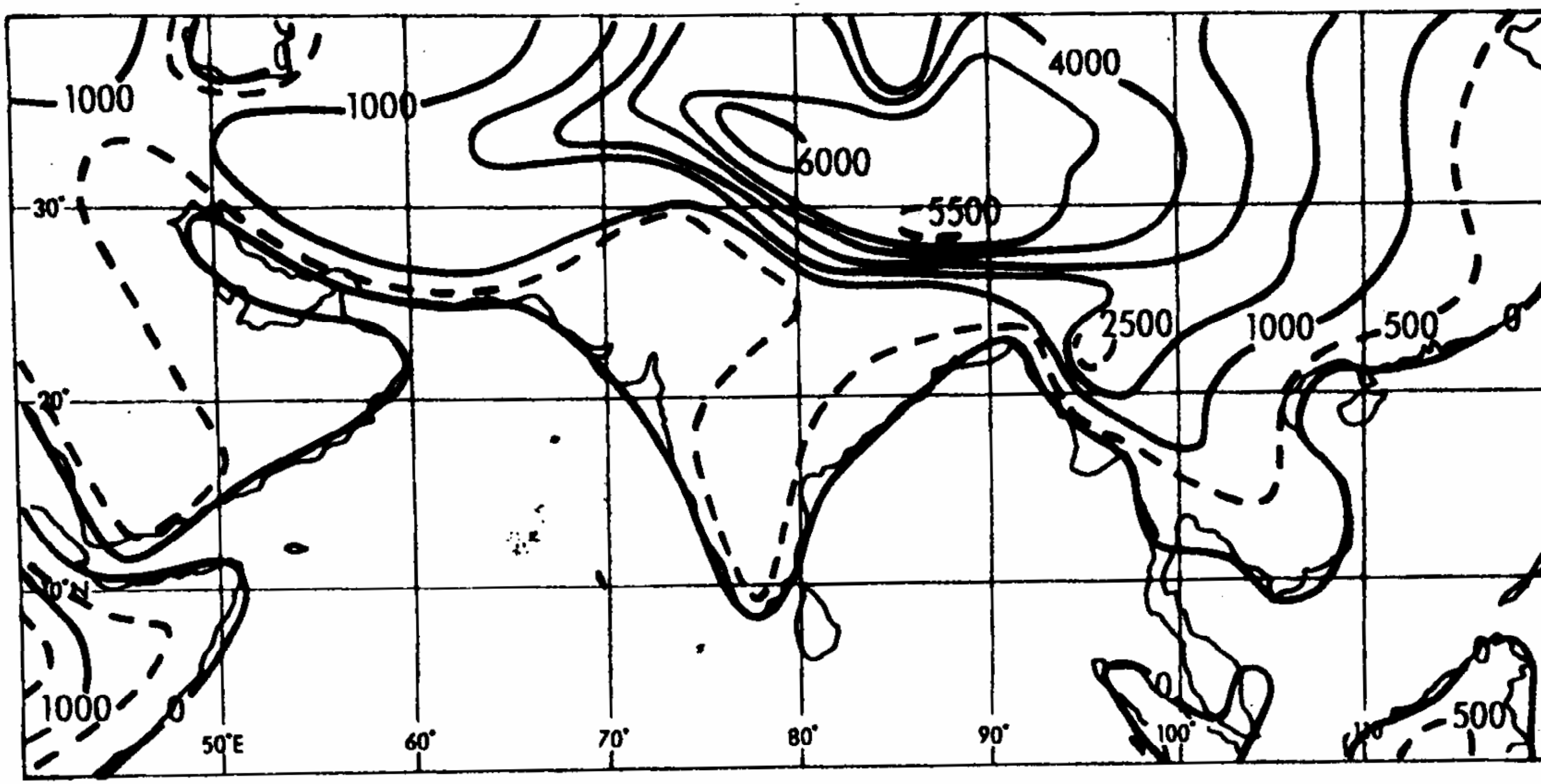
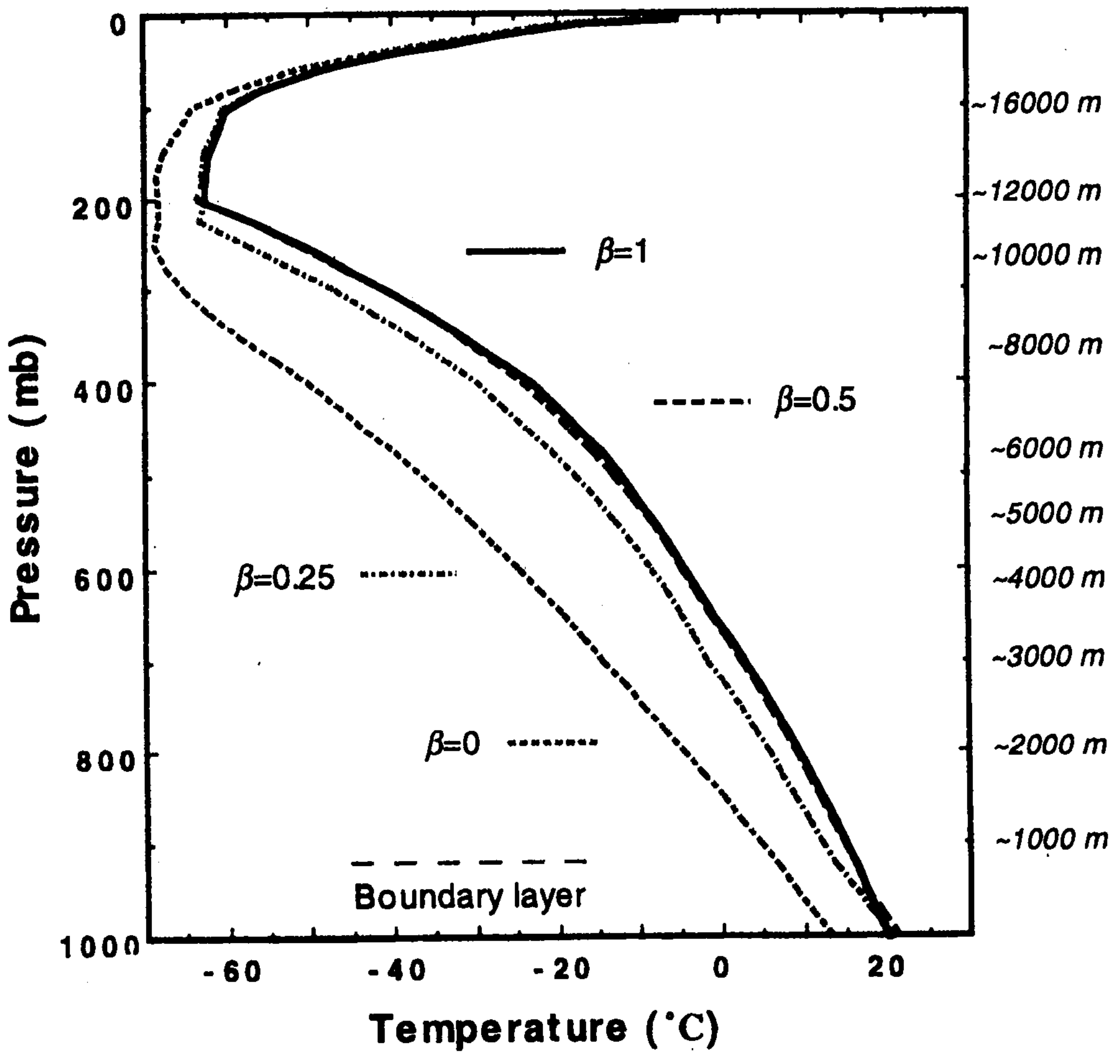
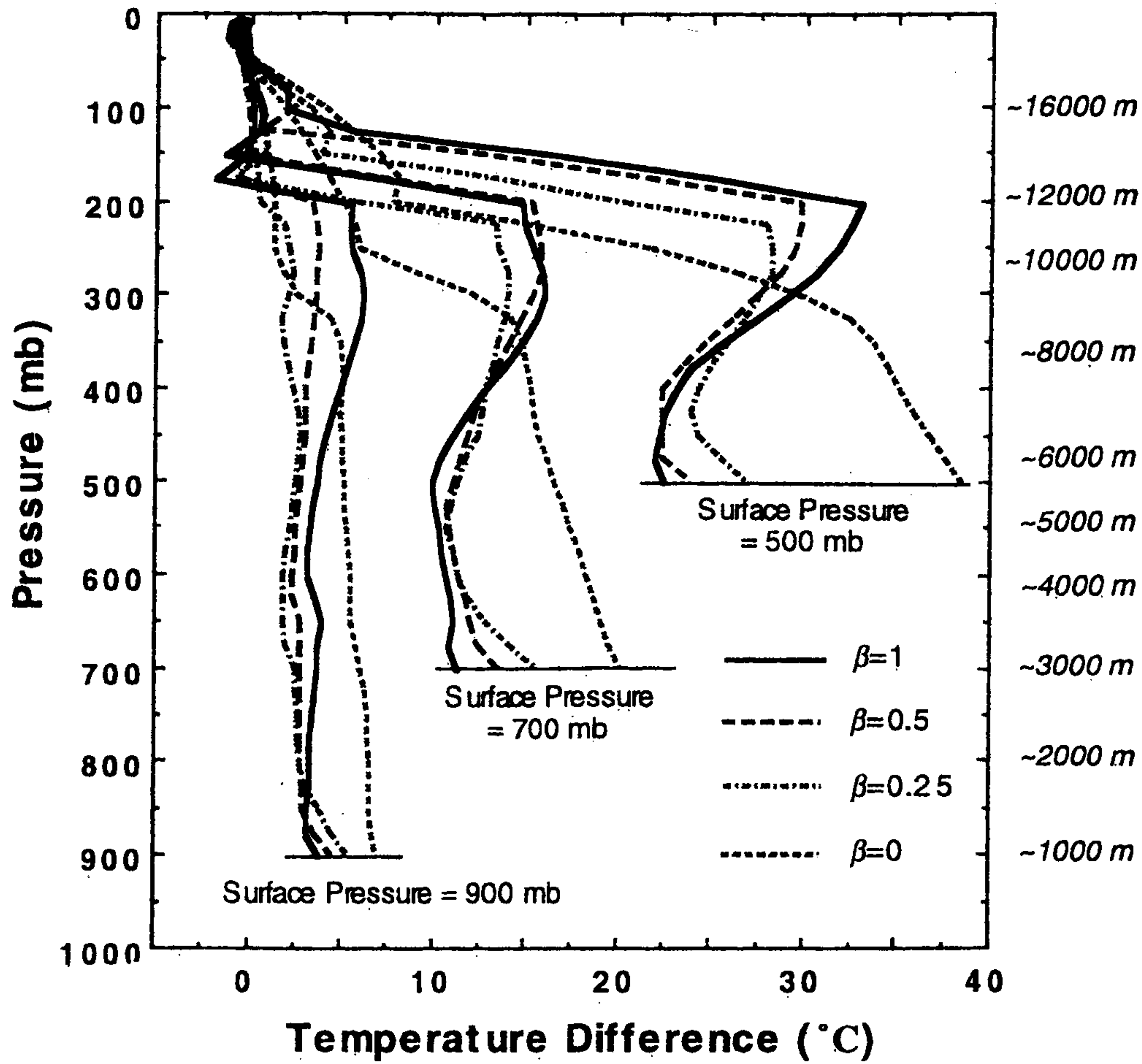
Physical Origins of the Monsoon


1000
1000
4000
6000
5500
2500
1000
500
0
30°
20°
10°N
1000
0
0
500
0
50°E
60°
70°
80°
90°
100°
110°

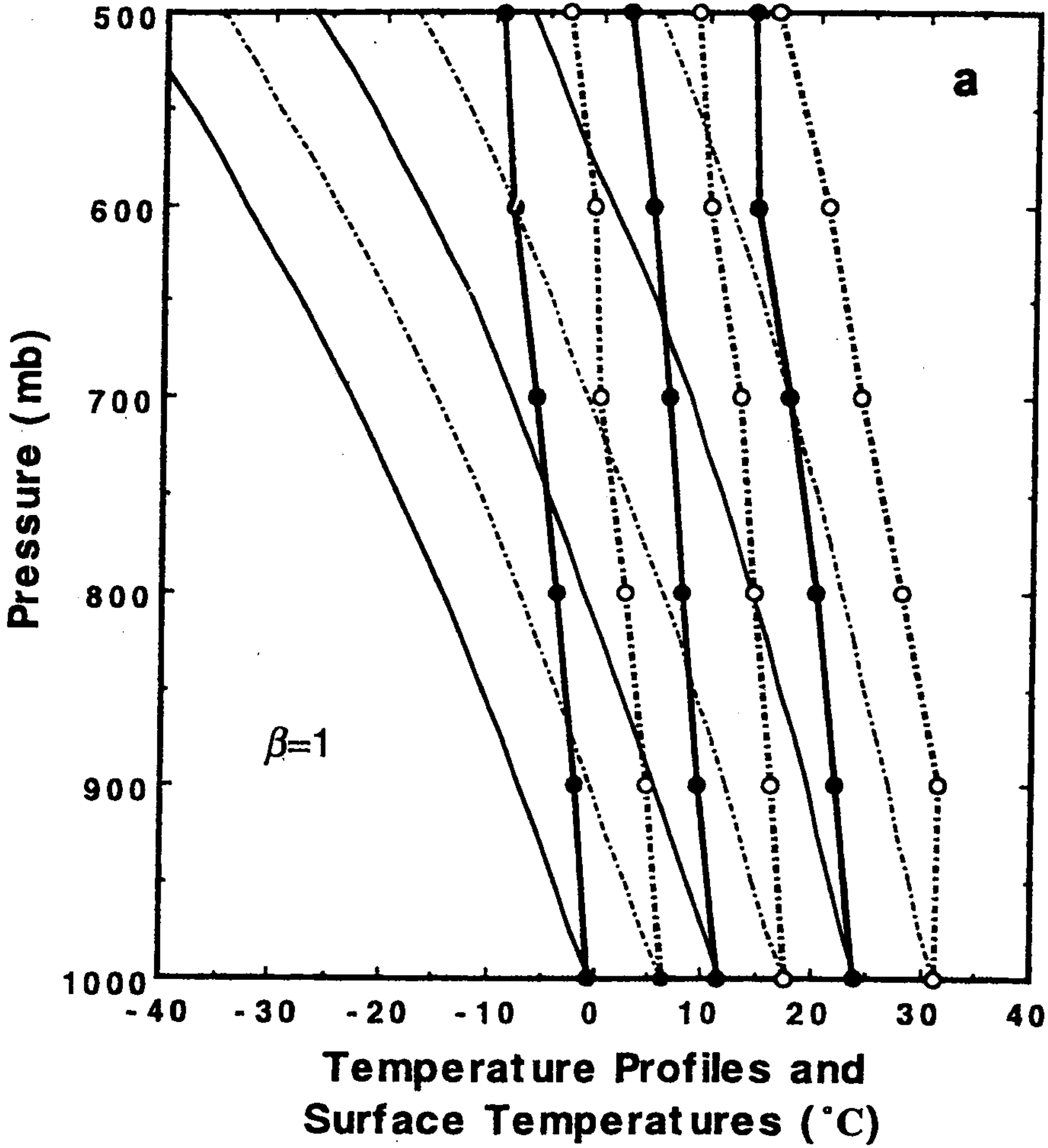
(a)

(from Molnar and Emanuel, 1999)

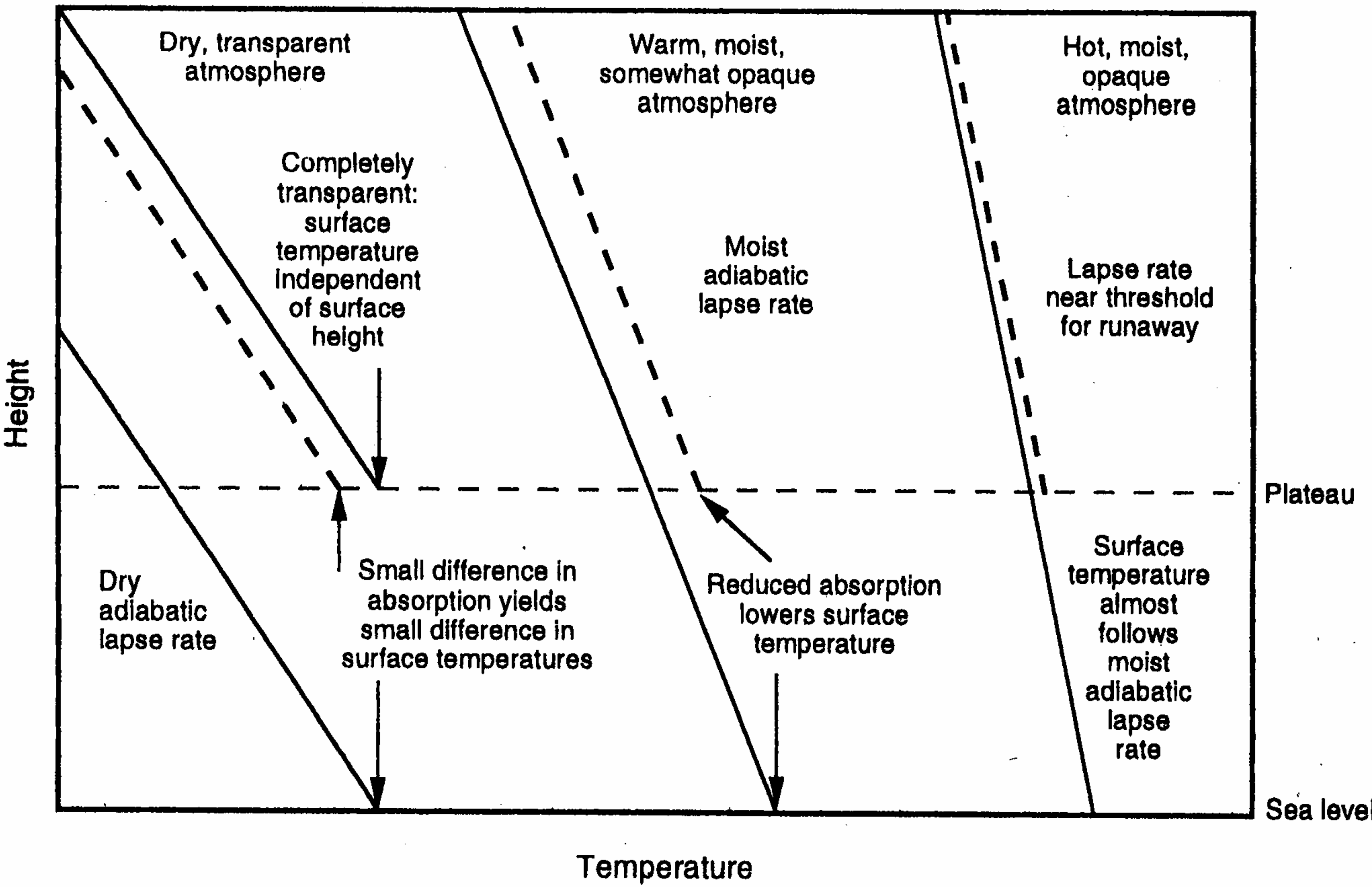




Pressure (mb)
Temperature Difference (°C)
0
100
200
300
400
500
600
700
800
900
1000
0
10
20
30
40
~16000 m
~12000 m
~10000 m
~8000 m
~6000 m
~5000 m
~4000 m
~3000 m
~2000 m
~1000 m
Surface Pressure = 500 mb
Surface Pressure = 700 mb
Surface Pressure = 900 mb
β=1
β=0.5
β=0.25
β=0



a
β=1
Pressure (mb)
500
600
700
800
900
1000
-40
-30
-20
-10
0
10
20
30
40
Temperature Profiles and
Surface Temperatures (°C)


Dry, transparent atmosphere
Warm, moist, somewhat opaque atmosphere
Hot, moist, opaque atmosphere
Completely transparent: surface temperature independent of surface height
Moist adiabatic lapse rate
Lapse rate near threshold for runaway
Height
Plateau
Dry adiabatic lapse rate
Small difference in absorption yields small difference in surface temperatures
Reduced absorption lowers surface temperature
Surface temperature almost follows moist adiabatic lapse rate
Sea level
Temperature

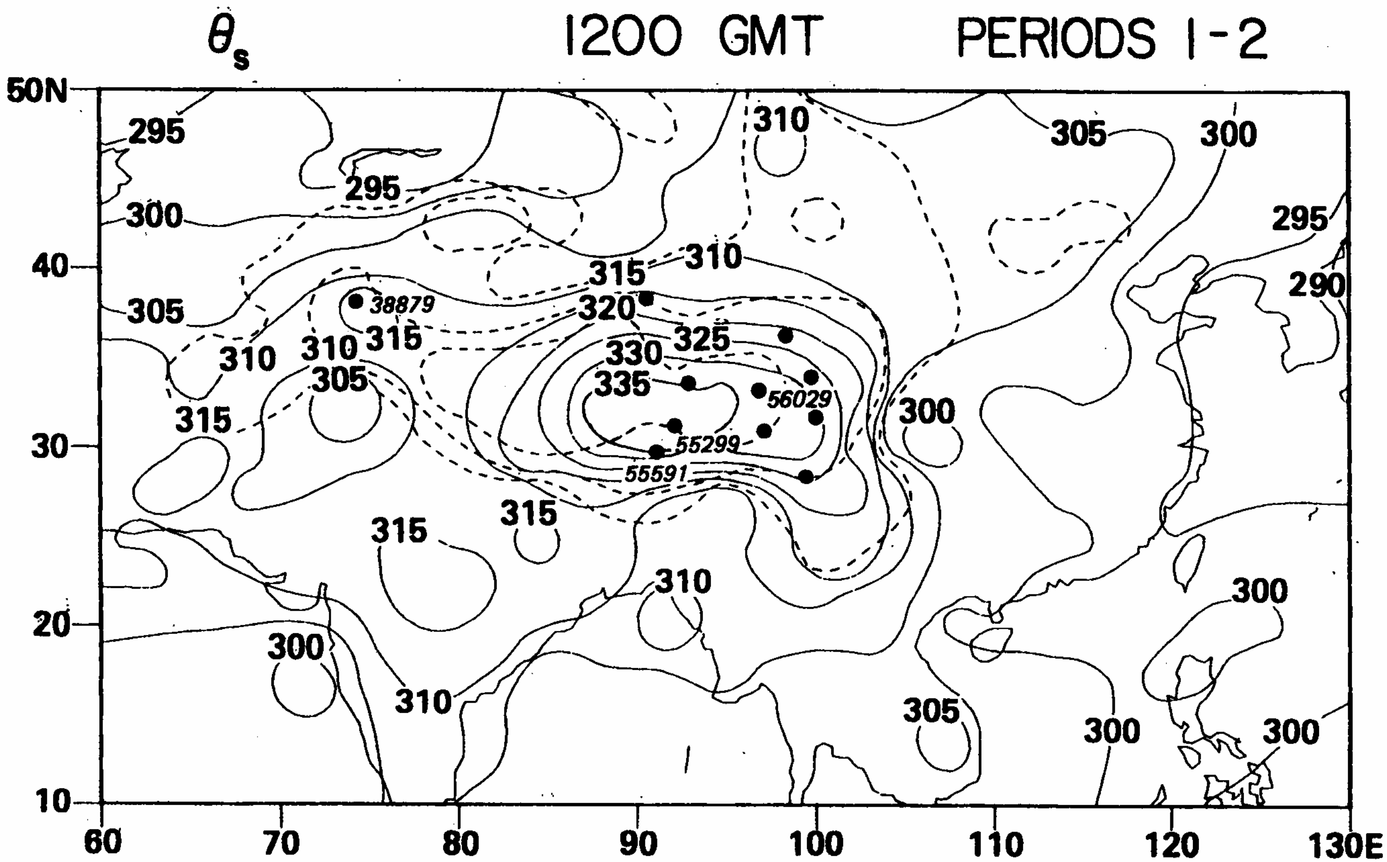
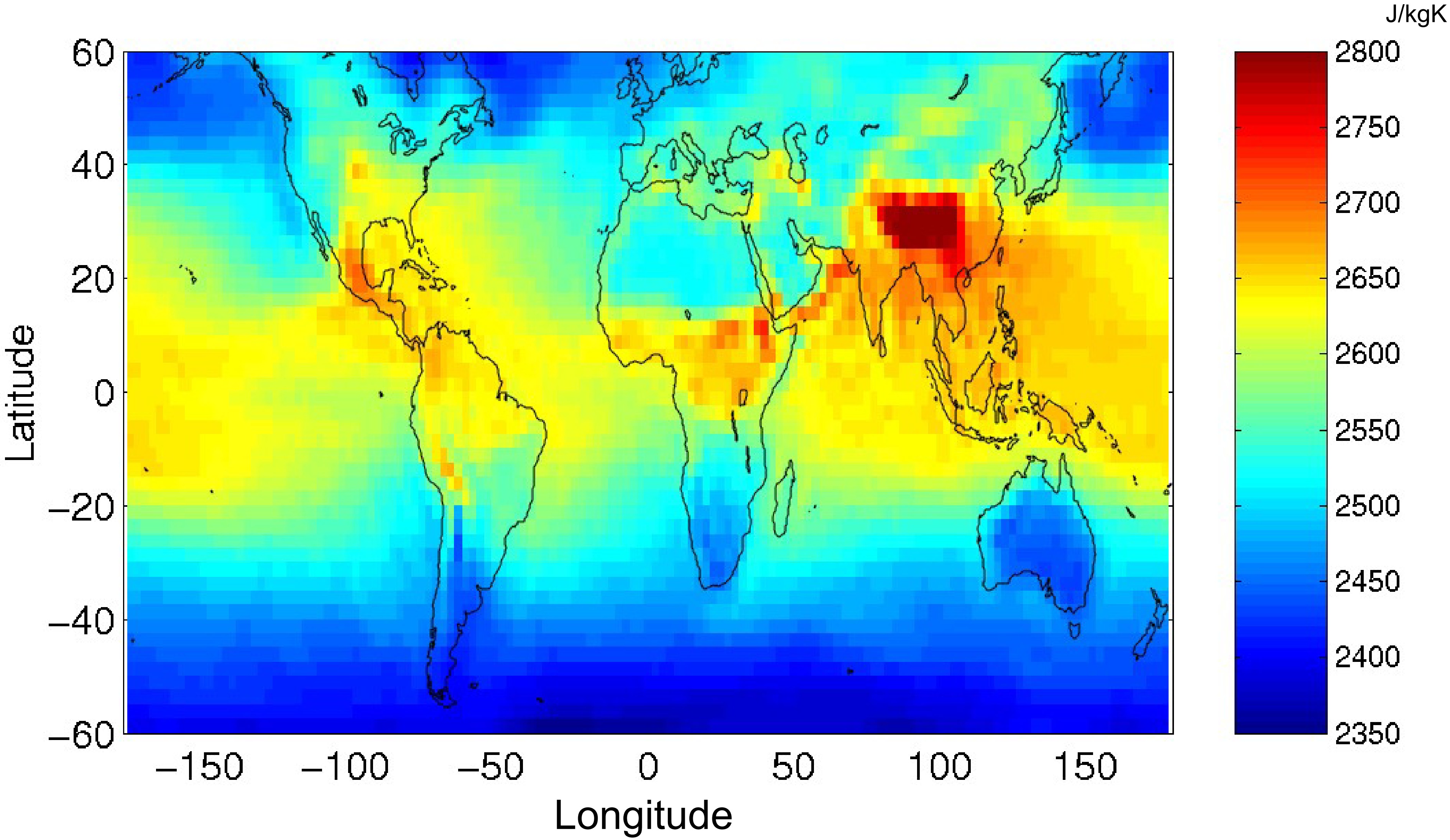


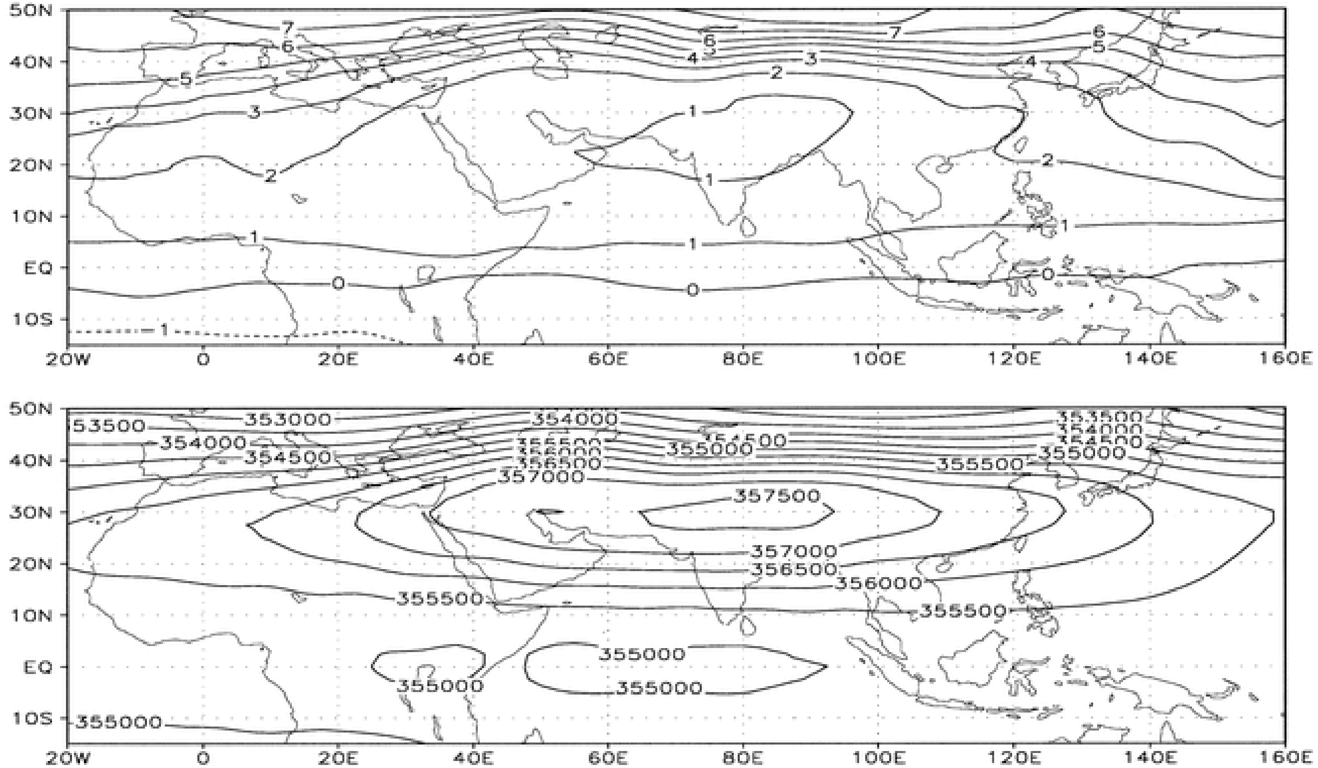
Fig. 9.14. The mean surface potential temperature θ_s at 1200 GMT during the first 10-day period. Dots represent radiosonde stations located above 3000 m. (After Luo and Yanai 1984.)

July 1 observed 1000 mb s_b



NCEP Reanalysis

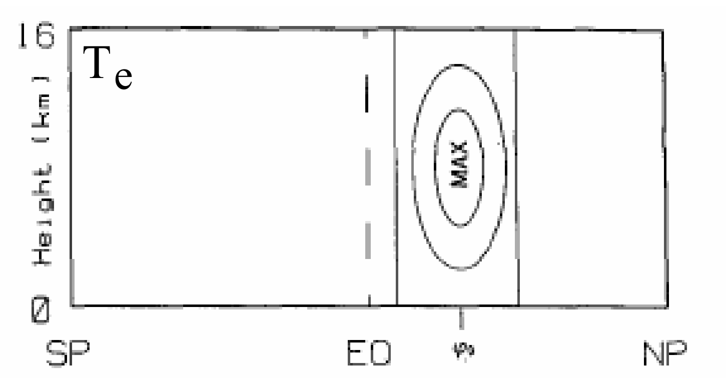
PV, M, on 370K, Jul 87-90

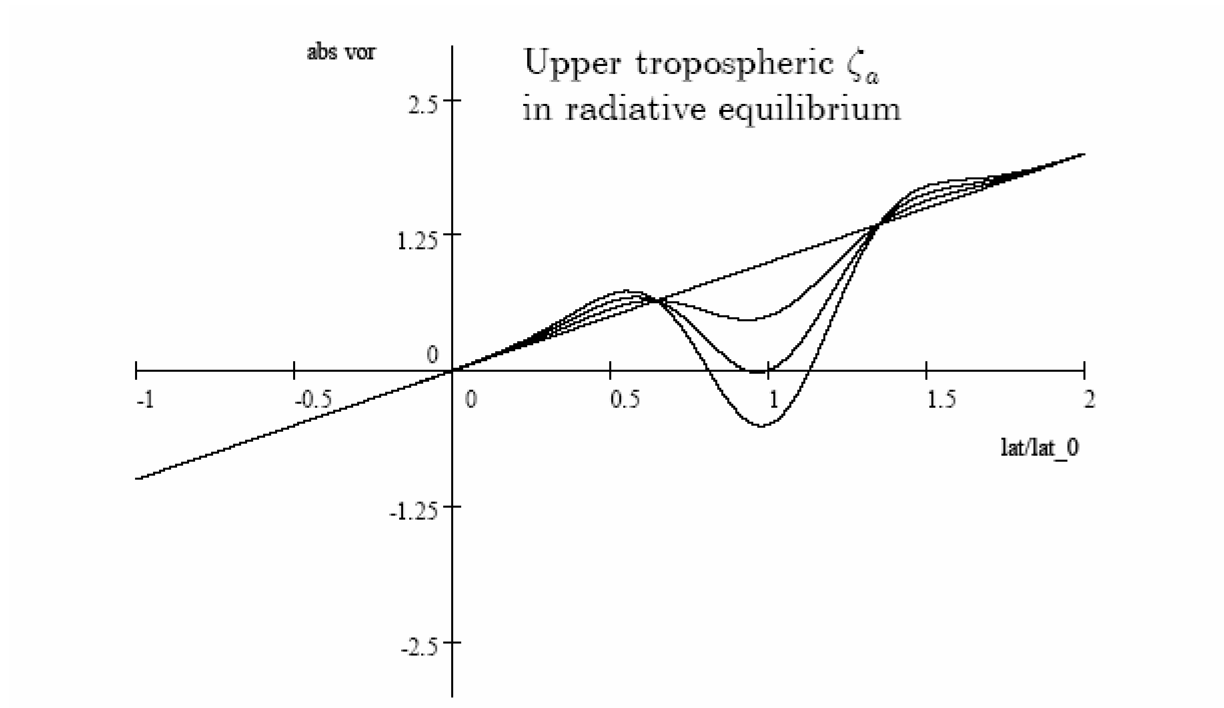


Two-D Simulations

Off-equatorial forcing in 2D

[Plumb & Hou, JAS, 1992]





Off-equatorial forcing

[Plumb & Hou, JAS, 1992]

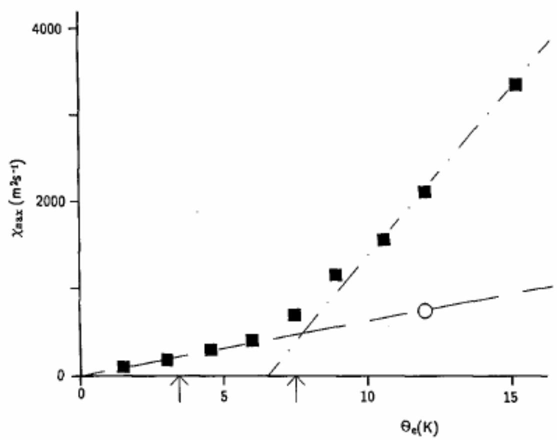
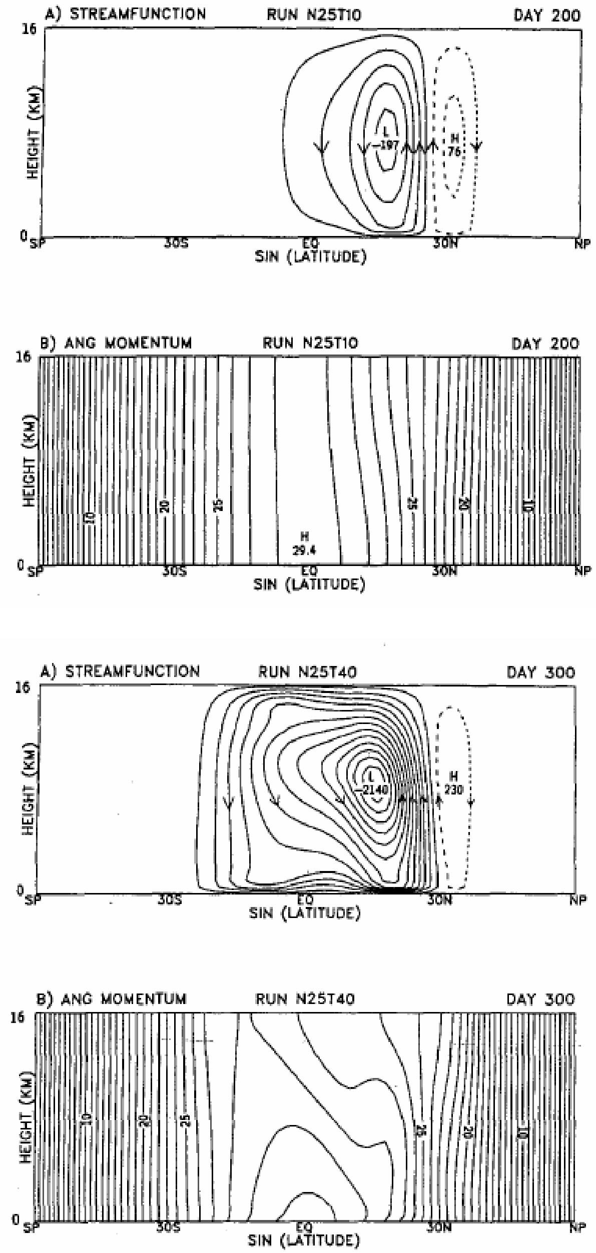
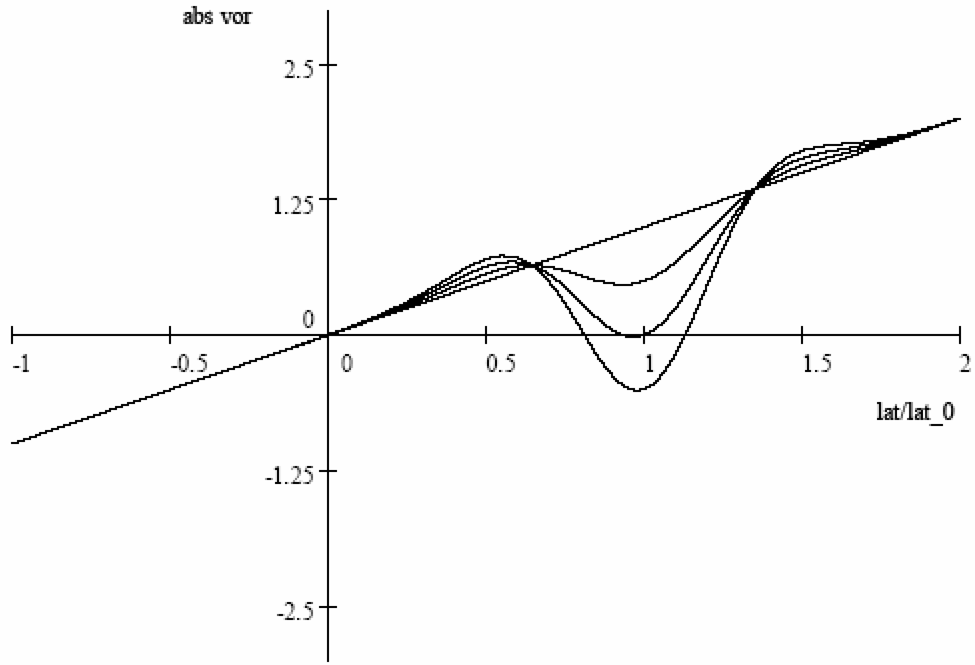


FIG. 4. Dependence of the maximum value of the steady streamfunction χ on the forcing amplitude Θ_e. The squares show points determined from results of the complete, nonlinear model. The circle shows a result from the linearized model, and the dashed line the linear dependence of χ_{max} on Θ_e. The steeper, dash–dot, line is drawn by eye and has no other significance. The two arrows show the theoretical value of Θ_e at which the TE solution becomes irregular; the left arrow is for the inviscid case, the right arrow for $\nu = 2.5\ m^2\ s^{-1}$ according to the linear model results.

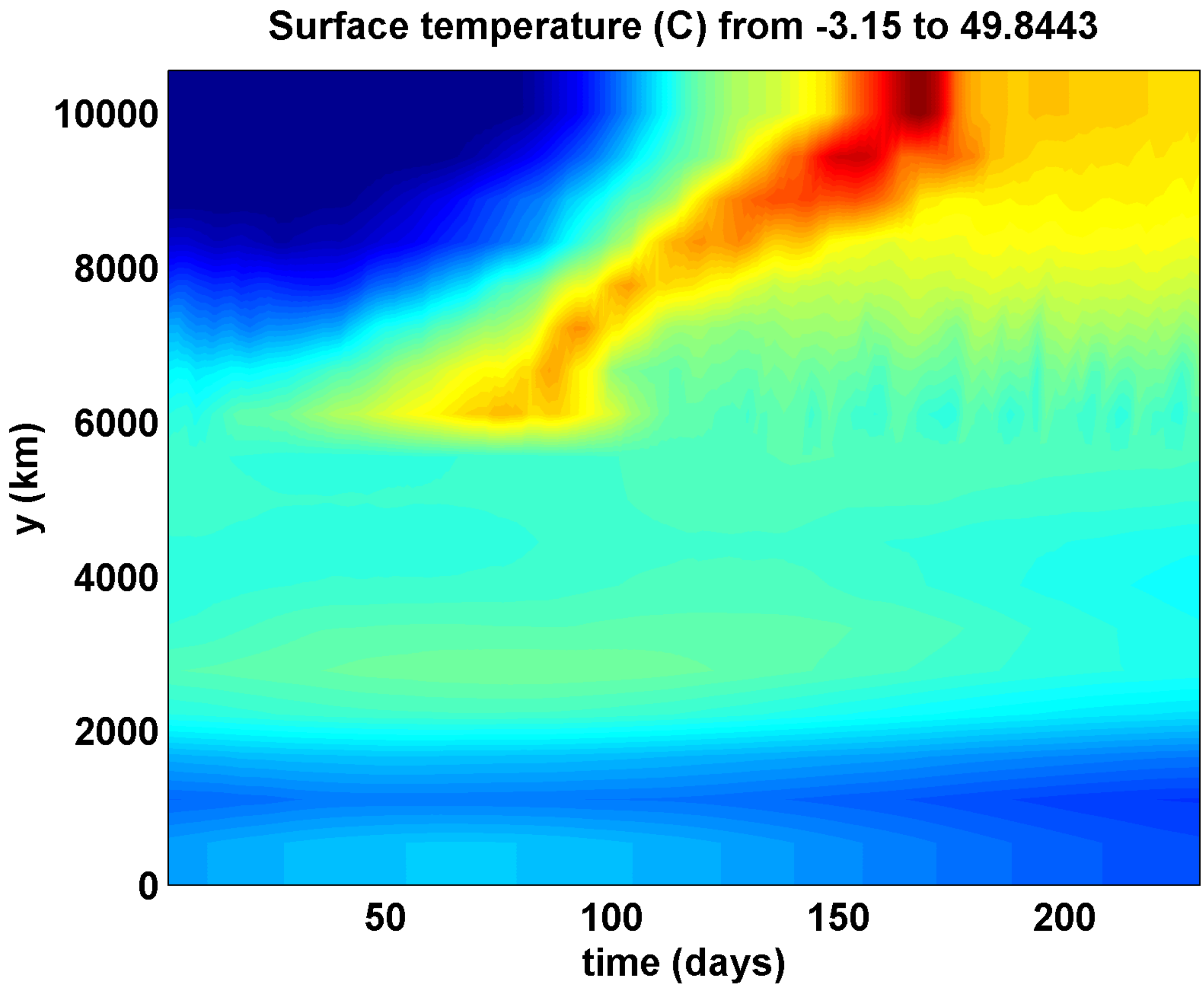


Does the $\zeta_a = 0$ criterion have any relevance under 3D dynamics?

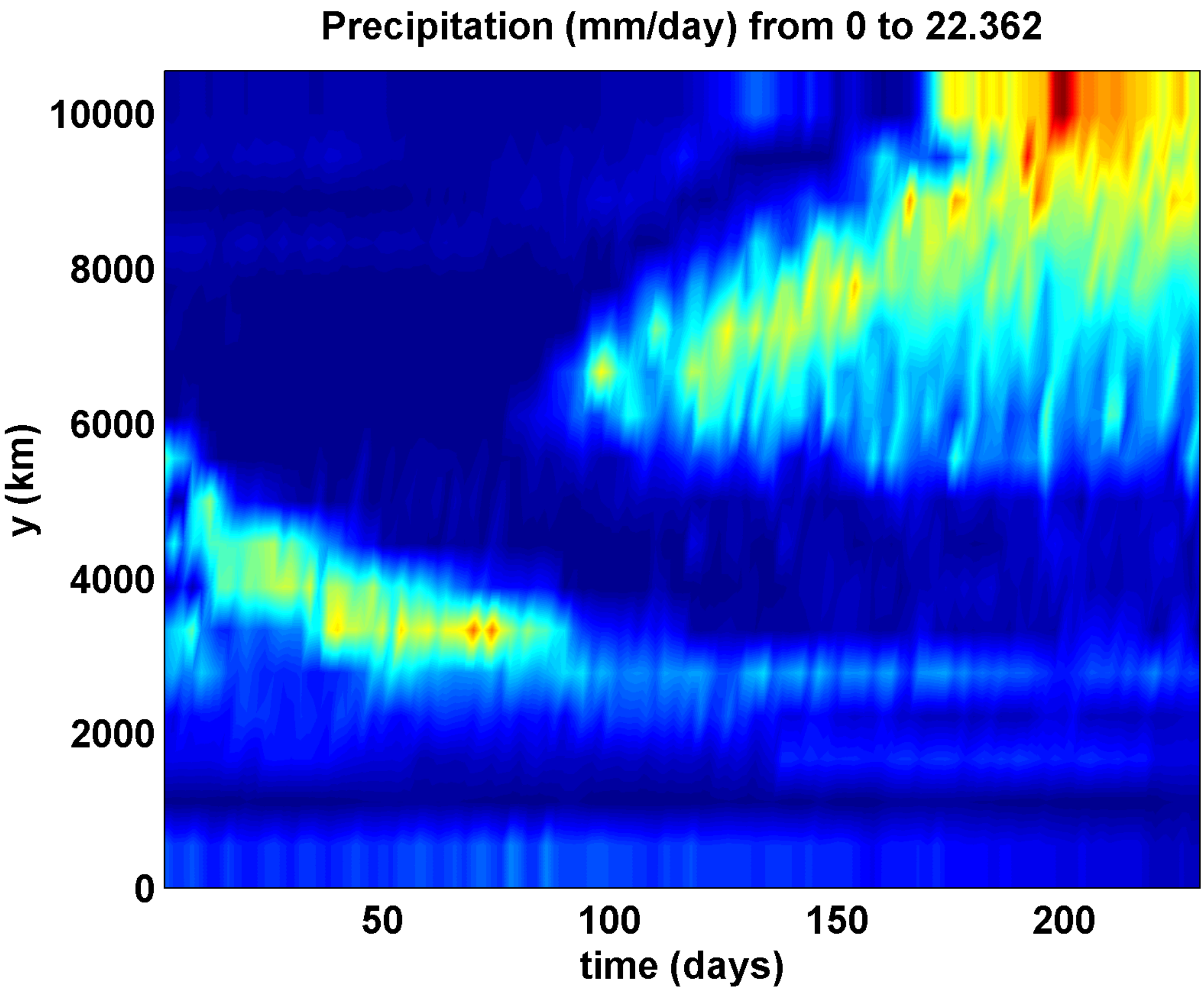


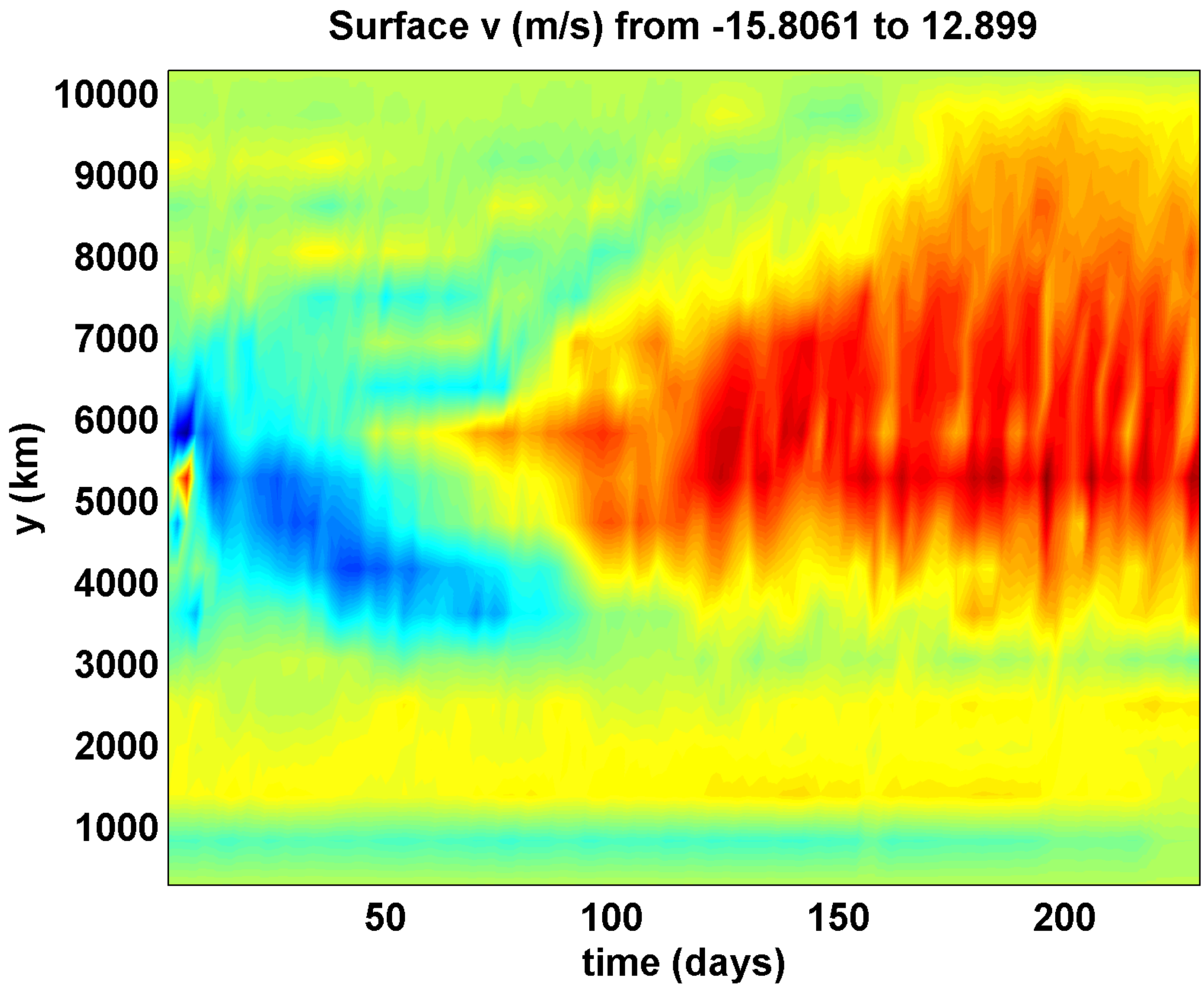
Simple PE Model

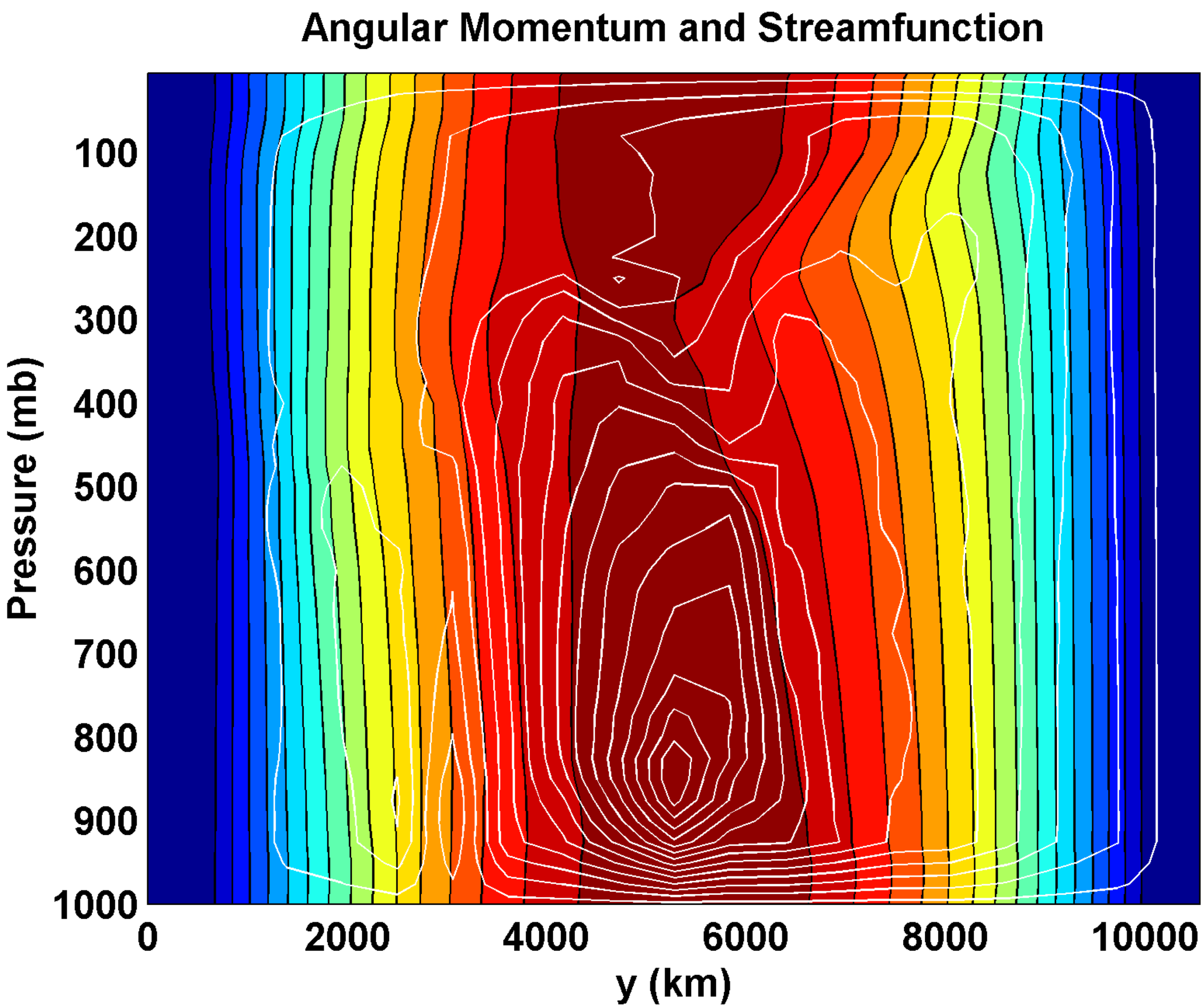
- Only 20 grid columns, N-S
- High resolution in vertical
- Convection, radiation, and cloud schemes
- Land poleward of 12 N (y=6000 km)
- Slab ocean
- Annual cycle of insolation

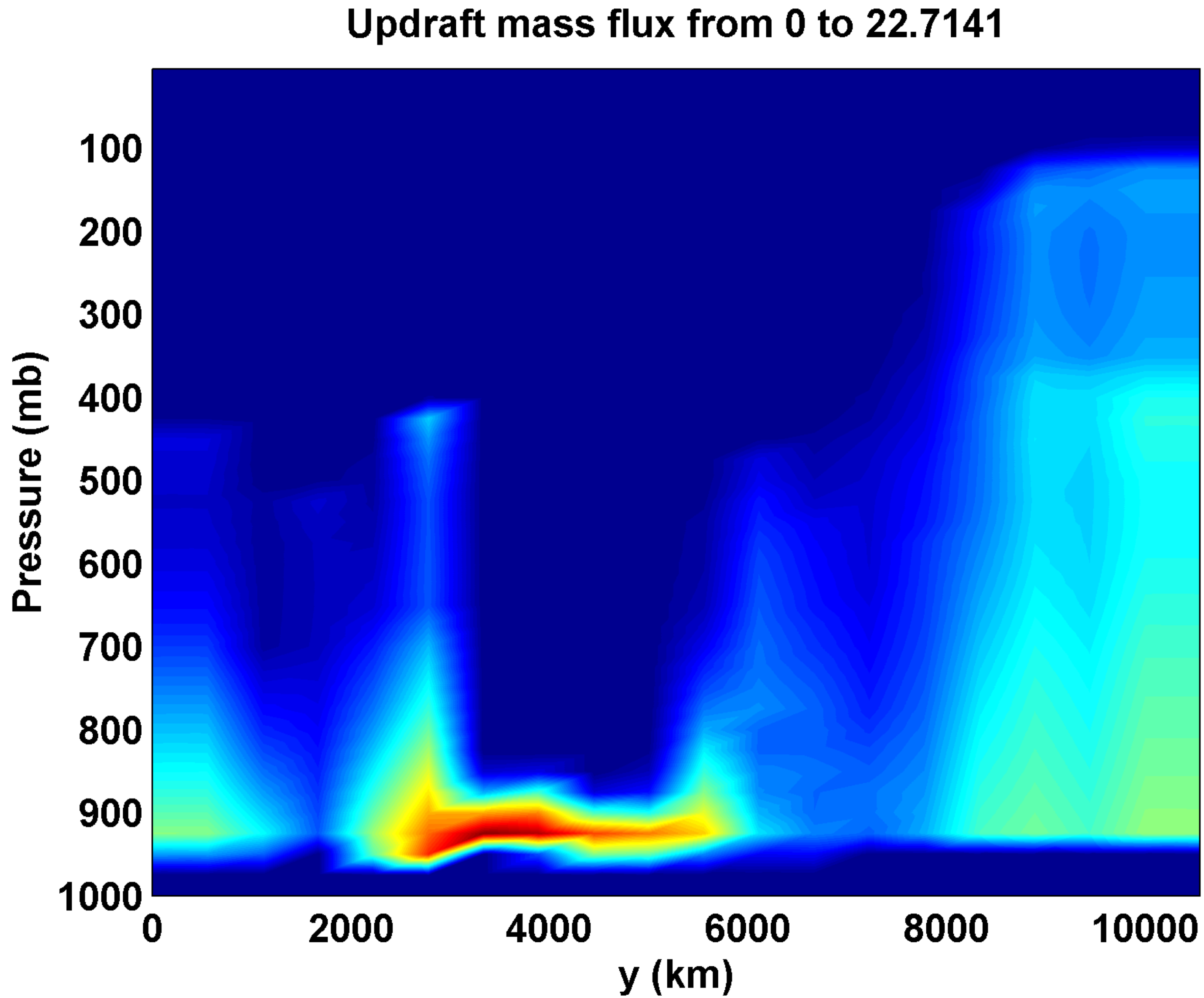


Surface temperature (C) from -3.15 to 49.8443
10000
8000
6000
4000
2000
0
y (km)
50
100
150
200
time (days)

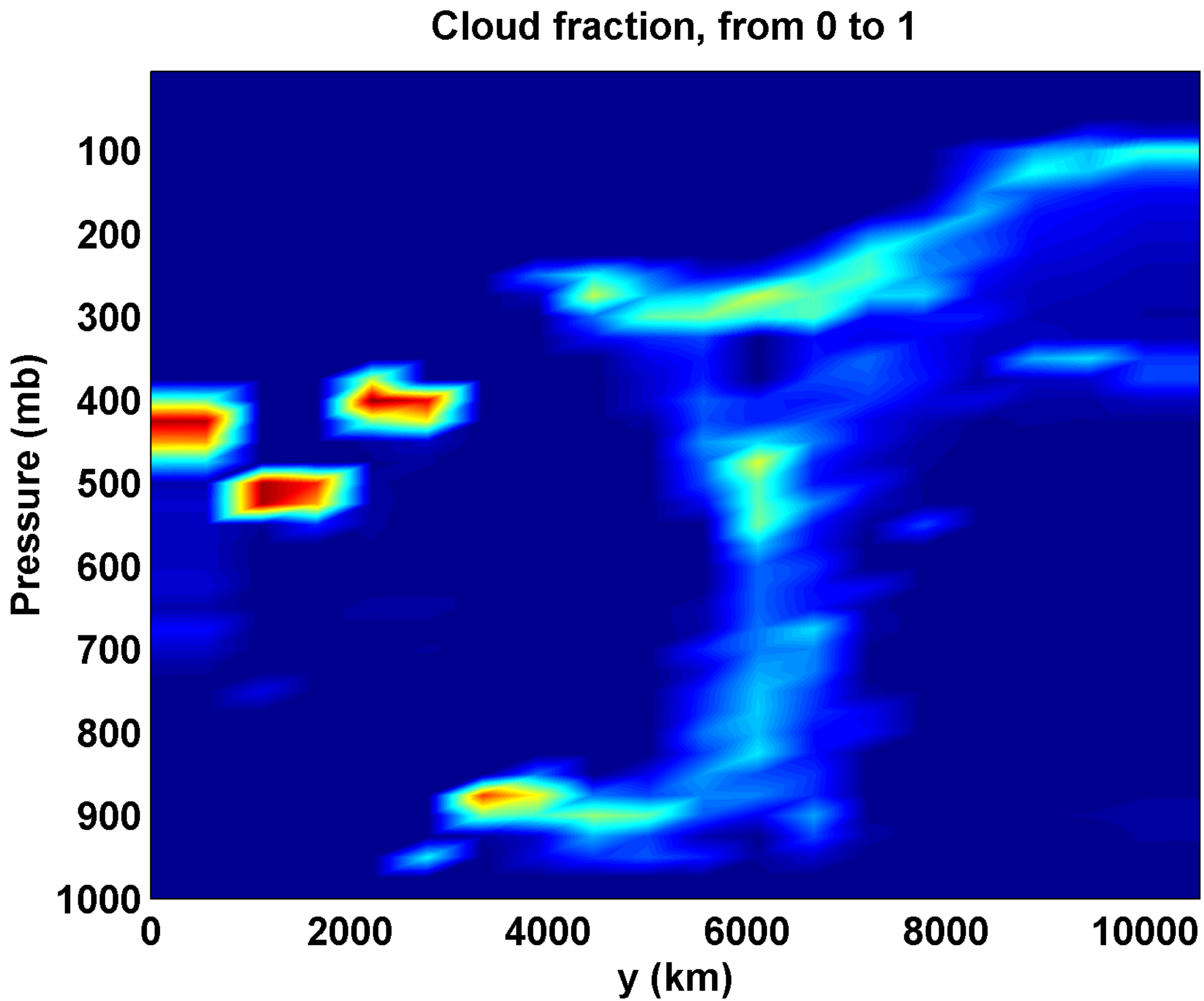

Precipitation (mm/day) from 0 to 22.362
10000
8000
6000
4000
2000
0
y (km)
50
100
150
200
time (days)


Surface v (m/s) from -15.8061 to 12.899
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
y (km)
50
100
150
200
time (days)

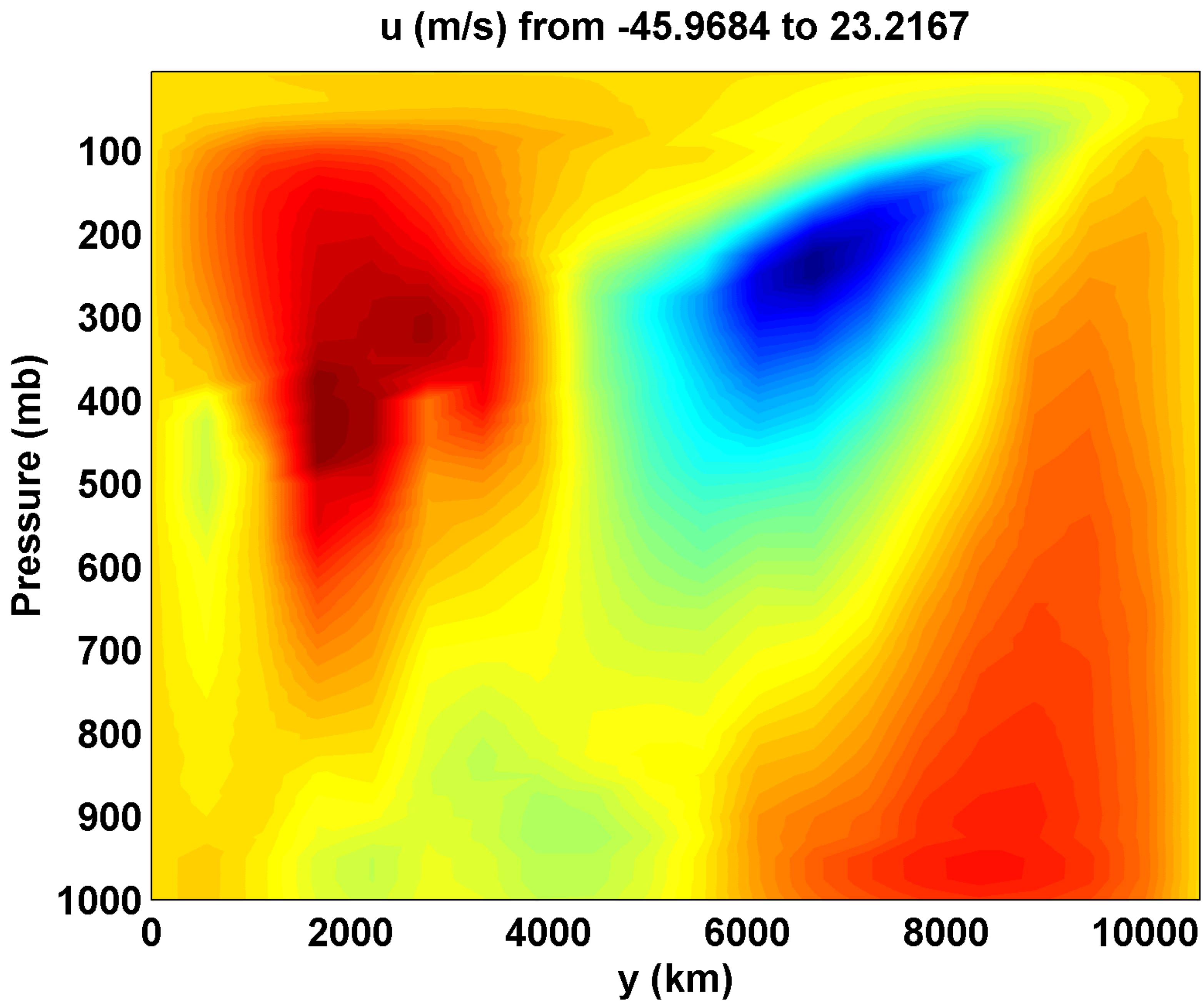

Angular Momentum and Streamfunction
Pressure (mb)
100
200
300
400
500
600
700
800
900
1000
0
2000
4000
6000
8000
10000
y (km)



Updraft mass flux from 0 to 22.7141
Pressure (mb)
100
200
300
400
500
600
700
800
900
1000
0
2000
4000
6000
8000
10000
y (km)



Cloud fraction, from 0 to 1
Pressure (mb)
100
200
300
400
500
600
700
800
900
1000
0
2000
4000
6000
8000
10000
y (km)


u (m/s) from -45.9684 to 23.2167
Pressure (mb)
100
200
300
400
500
600
700
800
900
1000
0
2000
4000
6000
8000
10000
y (km)

Model runs (*Nikki Prive*)

MIT model   64S – 64N; 4 degree resolution

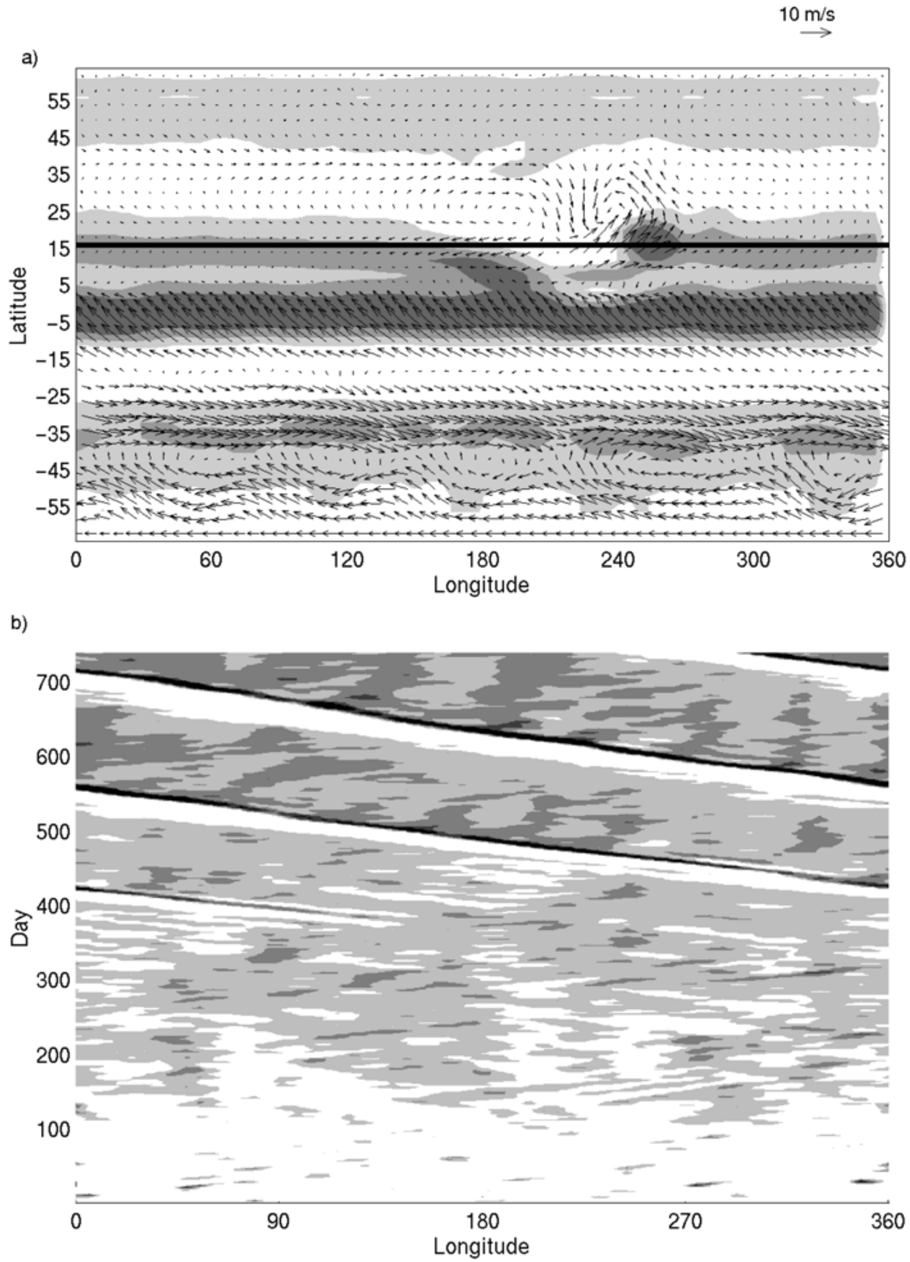
Moist model with simple lower boundary conditions

Ocean: specified SST

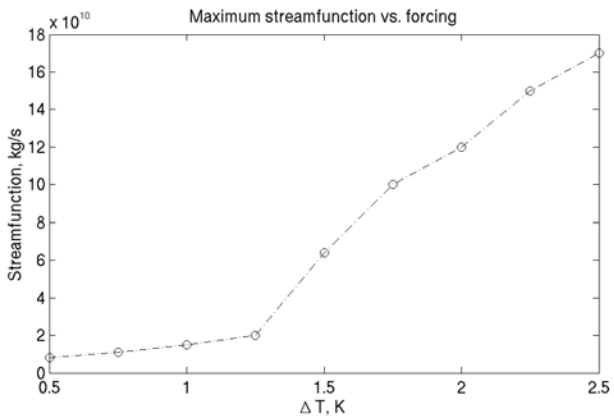
Land: specified total surface heat flux, bucket hydrology

Moist convection parameterization (Emanuel)

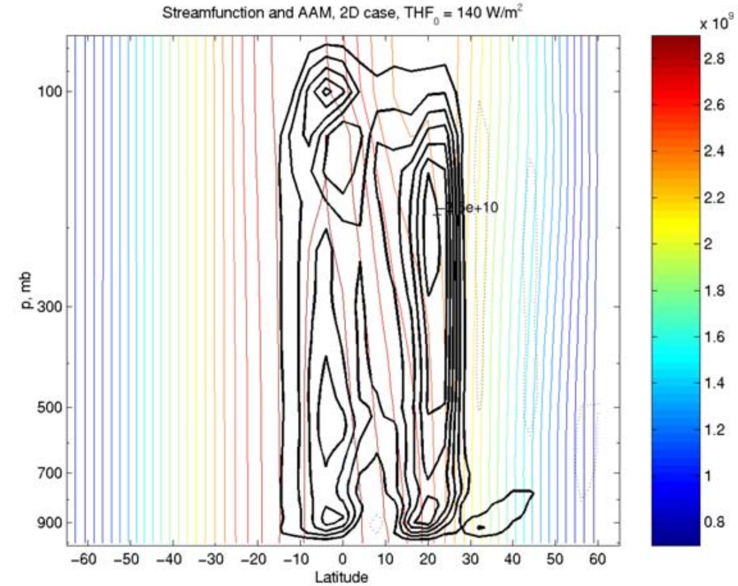
“Radiation”: Newtonian relaxation to 200K

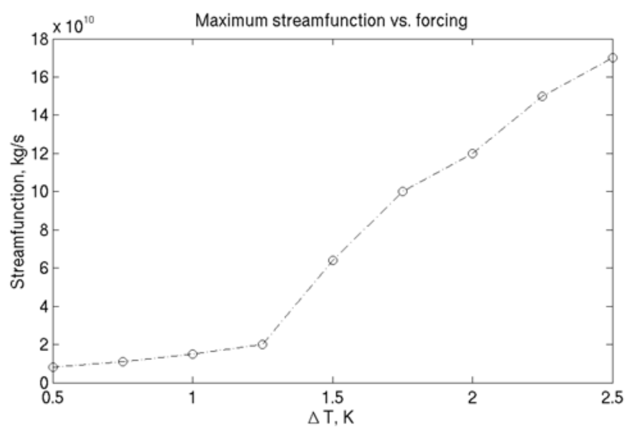


10 m/s
a)
Latitude
55
45
35
25
15
5
−5
−15
−25
−35
−45
−55
0
60
120
180
240
300
360
Longitude
b)
Day
700
600
500
400
300
200
100
0
90
180
270
360
Longitude

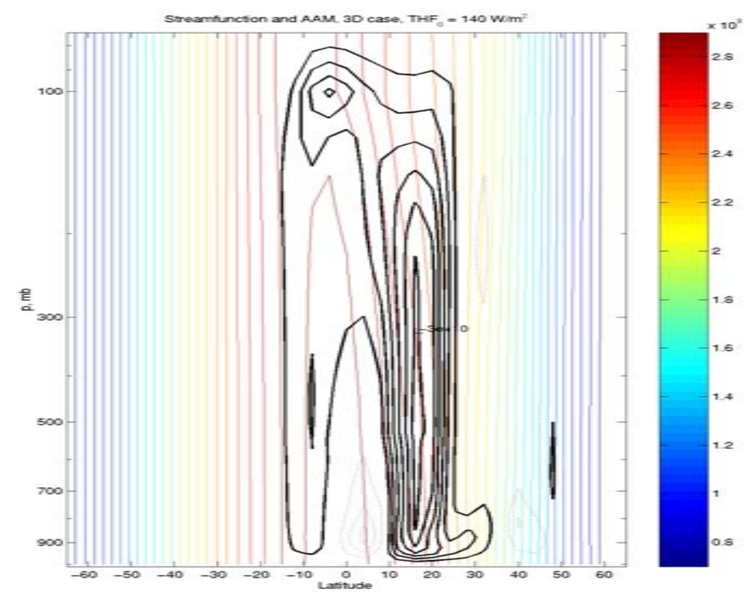

Maximum streamfunction vs. forcing
x 10^10
Streamfunction, kg/s
Δ T, K
0.5 1 1.5 2 2.5
0 2 4 6 8 10 12 14 16 18

2D


Streamfunction and AAM, 2D case, THF$_0$ = 140 W/m^2
x 10^9
p, mb
Latitude
100 300 500 700 900
-60 -50 -40 -30 -20 -10 0 10 20 30 40 50 60
0.8 1 1.2 1.4 1.6 1.8 2 2.2 2.4 2.6 2.8


Maximum streamfunction vs. forcing
x 10^10
Streamfunction, kg/s
Δ T, K
0.5 1 1.5 2 2.5
0 2 4 6 8 10 12 14 16 18

3D


Streamfunction and AAM, 3D case, THF$_0$ = 140 W/m^2
p, mb
Latitude
100 300 500 700 900
-60 -50 -40 -30 -20 -10 0 10 20 30 40 50 60

3) Cross-equatorial flow: does three-dimensionality matter?

Cross-equatorial flow *[Pauluis, JAS, 2004]*

Strong SST gradient across equator

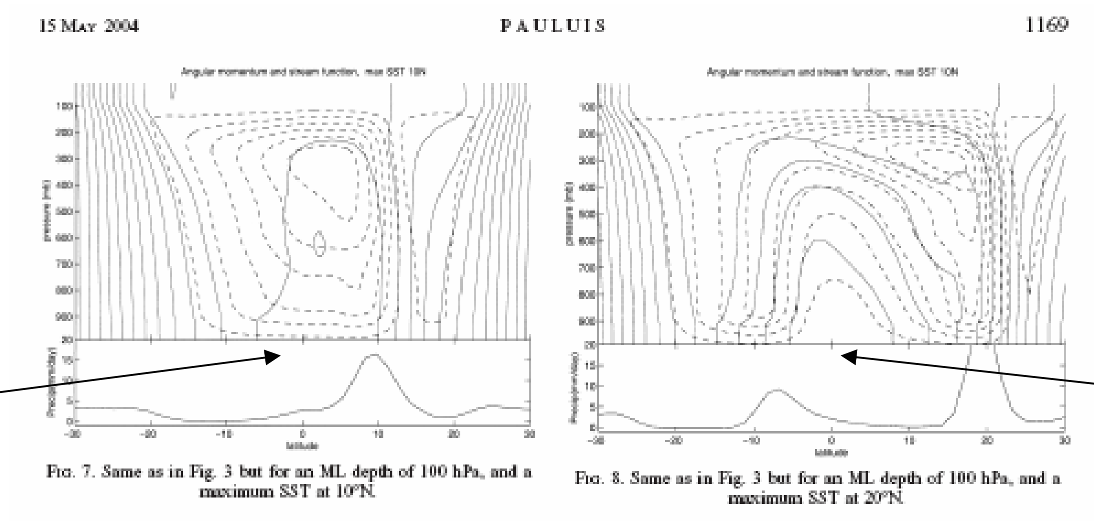


Fig. 7. Same as in Fig. 3 but for an ML depth of 100 hPa, and a maximum SST at 10°N.

Fig. 8. Same as in Fig. 3 but for an ML depth of 100 hPa, and a maximum SST at 20°N.

Weak SST gradient across equator

Shallow boundary layer

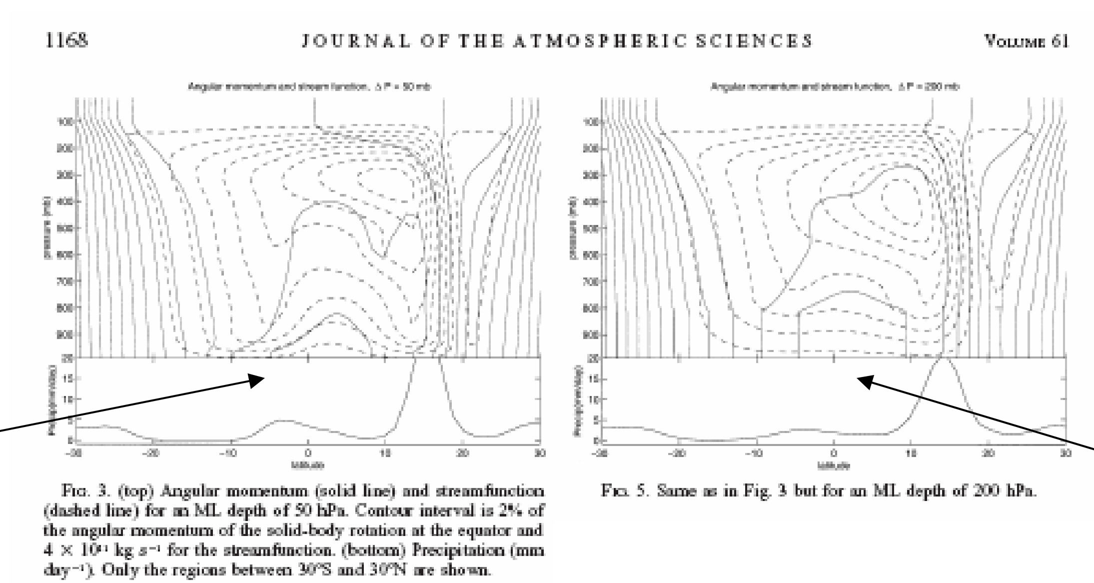
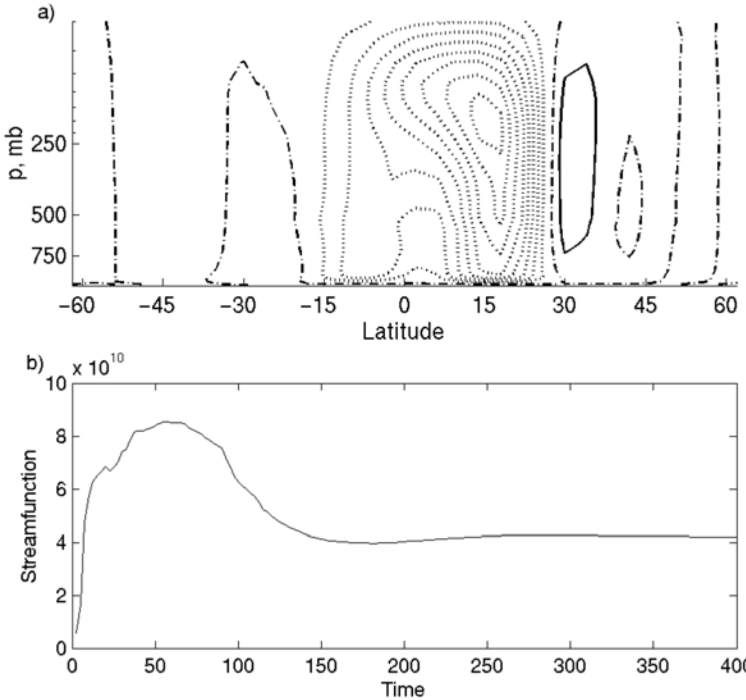


Fig. 3. (top) Angular momentum (solid line) and streamfunction (dashed line) for an ML depth of 50 hPa. Contour interval is 2% of the angular momentum of the solid-body rotation at the equator and 4×10^{11} kg s^{-1} for the streamfunction. (bottom) Precipitation (mm day^{-1}). Only the regions between 30°S and 30°N are shown.

Fig. 5. Same as in Fig. 3 but for an ML depth of 200 hPa.

Deep boundary layer

3D



2D

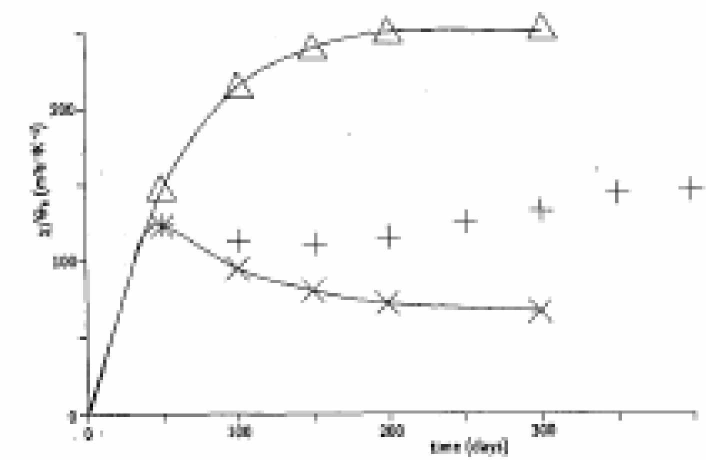
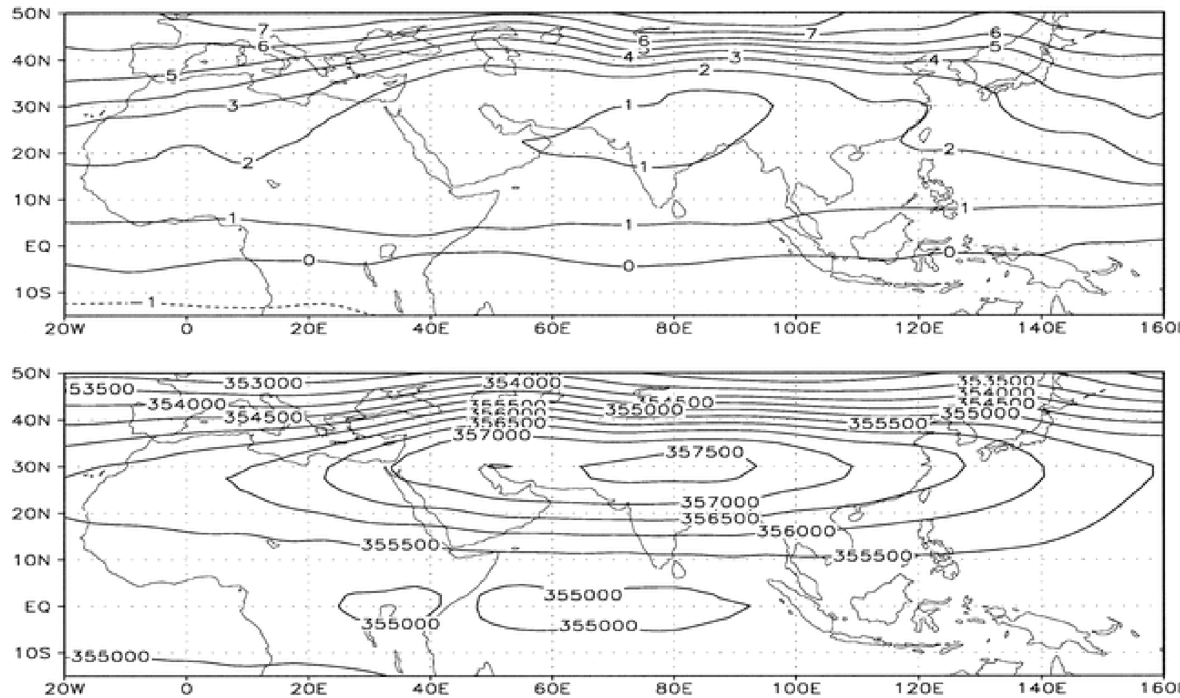
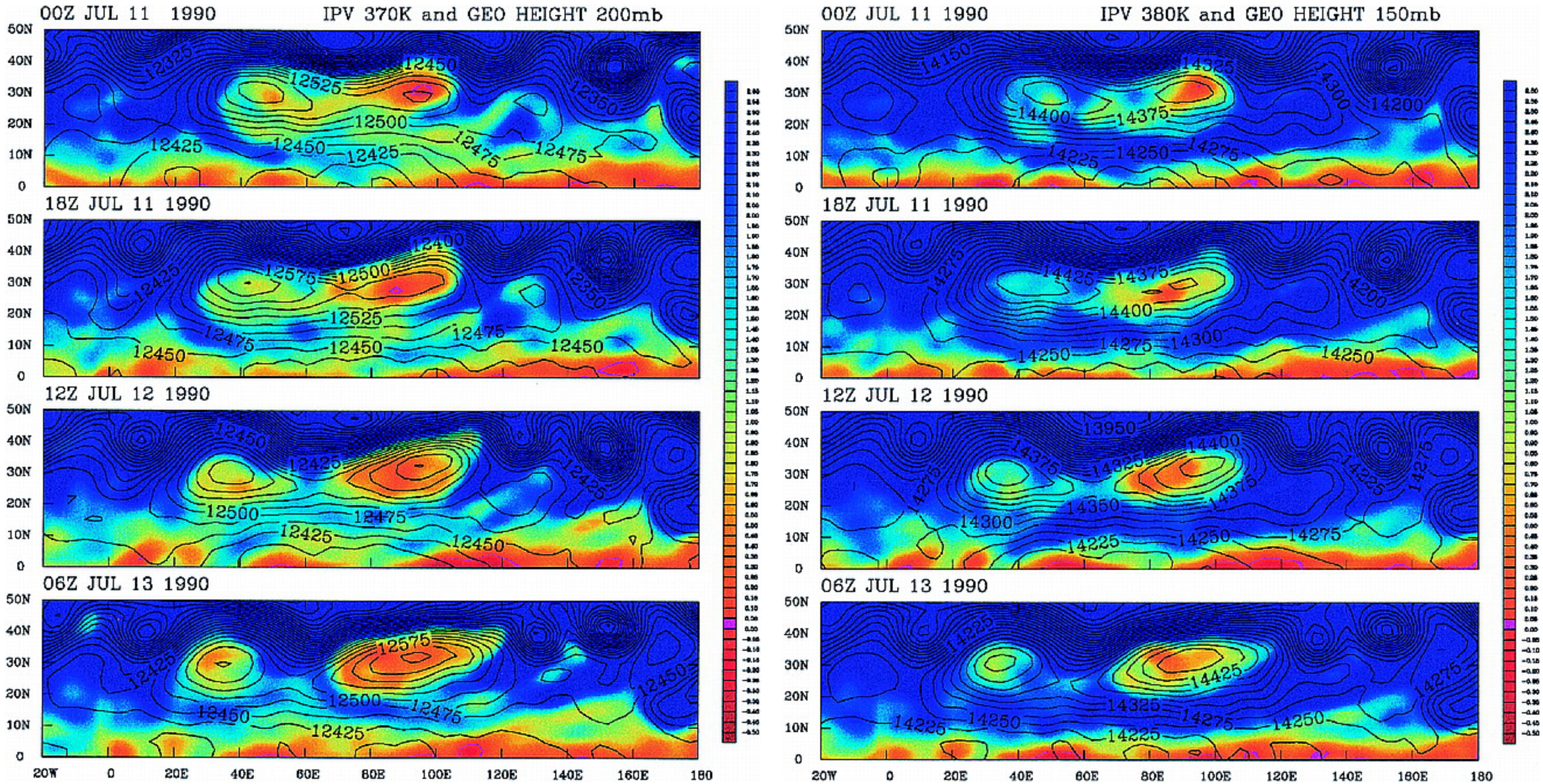


FIG. 6. Development of the circulation with time. Plot shows the streamfunction, χ_{max} scaled by θ_e, as a function of time. Cases (values of θ_e) are: (×) 3.0 K, (+) 7.5 K, (Δ) 12.1 K.


50N
40N
30N
20N
10N
EQ
10S
20W
0
20E
40E
60E
80E
100E
120E
140E
160E
7
6
5
4
3
2
1
0
353000
353500
354000
354500
355000
355500
356000
356500
357000
357500

Upper tropospheric PV on θ = 370/380K and Z on 200/150hPa

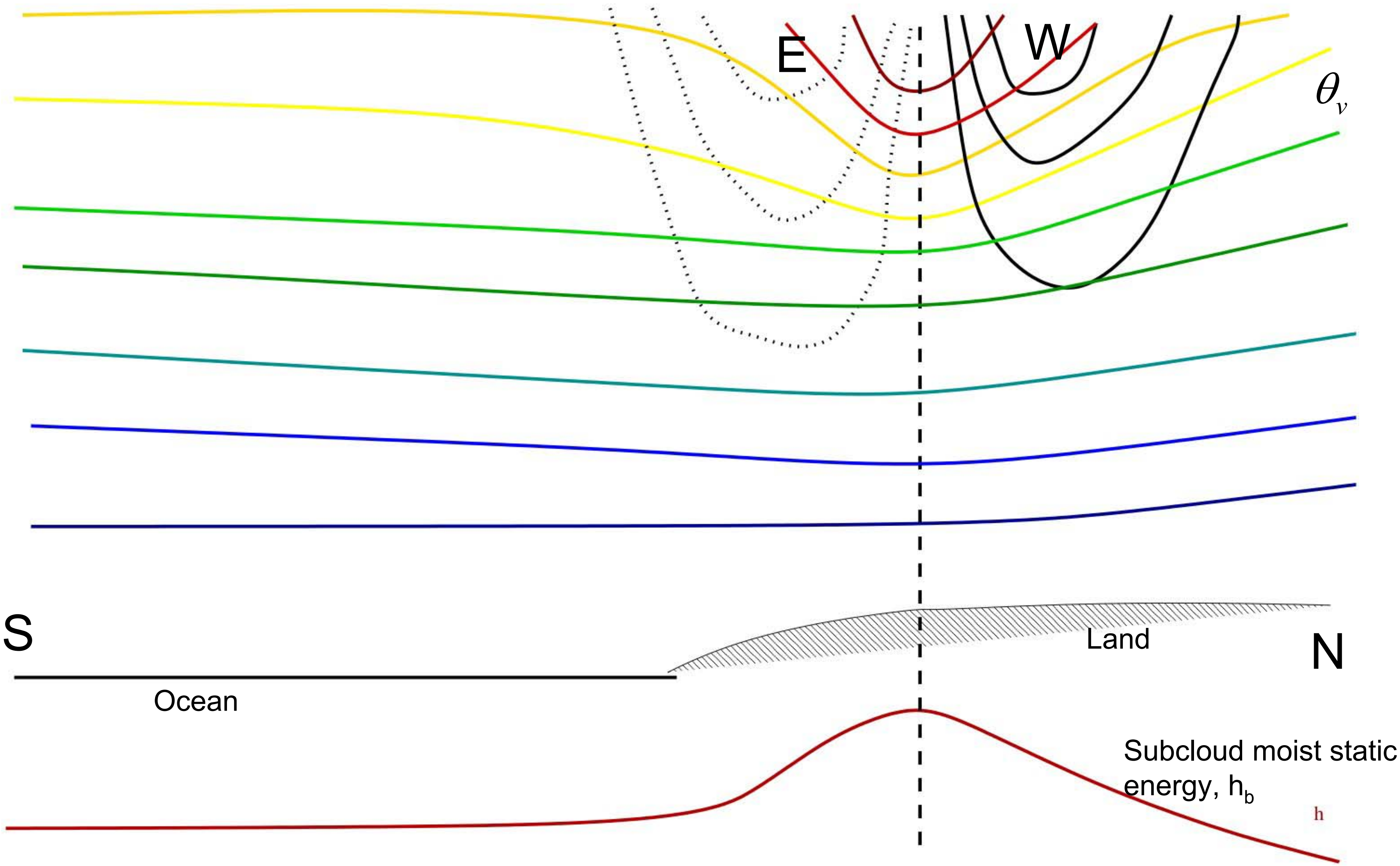
[Hsu & Plumb 1999]



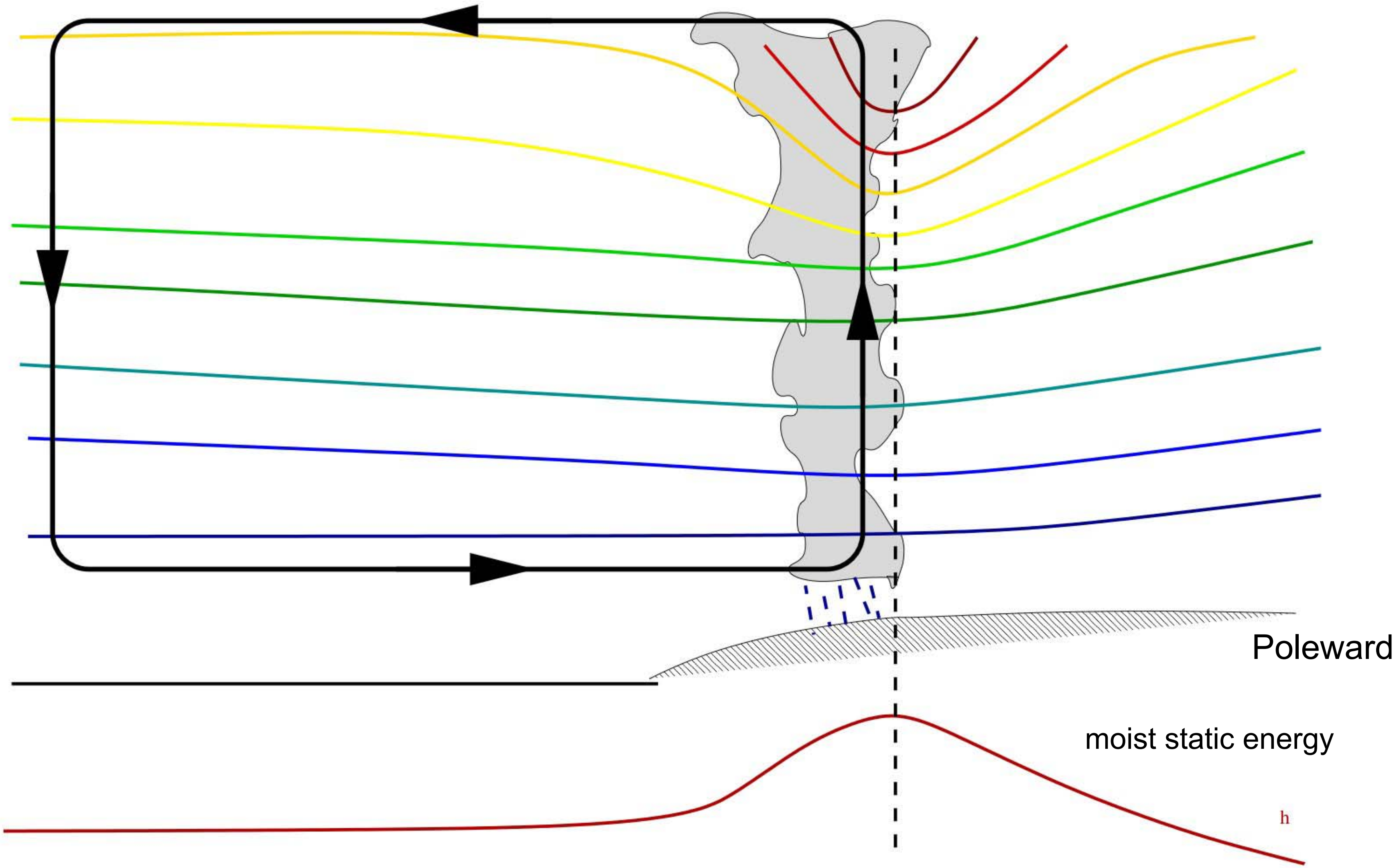
Theories of Monsoon Location

- Plumb and Hou (1992), Emanuel (1995), Zheng (1998)
  - Explained axisymmetric circulation induced by local subtropical forcing
- Rodwell and Hoskins (1995)
  - Rossby waves induce subsidence to the west of the monsoon, creating east-west asymmetry
- Xie and Saiki (1999)
  - Hydrological feedbacks limit inland progression of the monsoon
- Chou, Neelin, and Su (2001)
  - Advection of low moist static energy air, hydrological feedbacks, and Rodwell-Hoskins effect all limit poleward extent of the monsoon

Impact of local h_b maximum over land



Resulting meridional circulation and precipitation

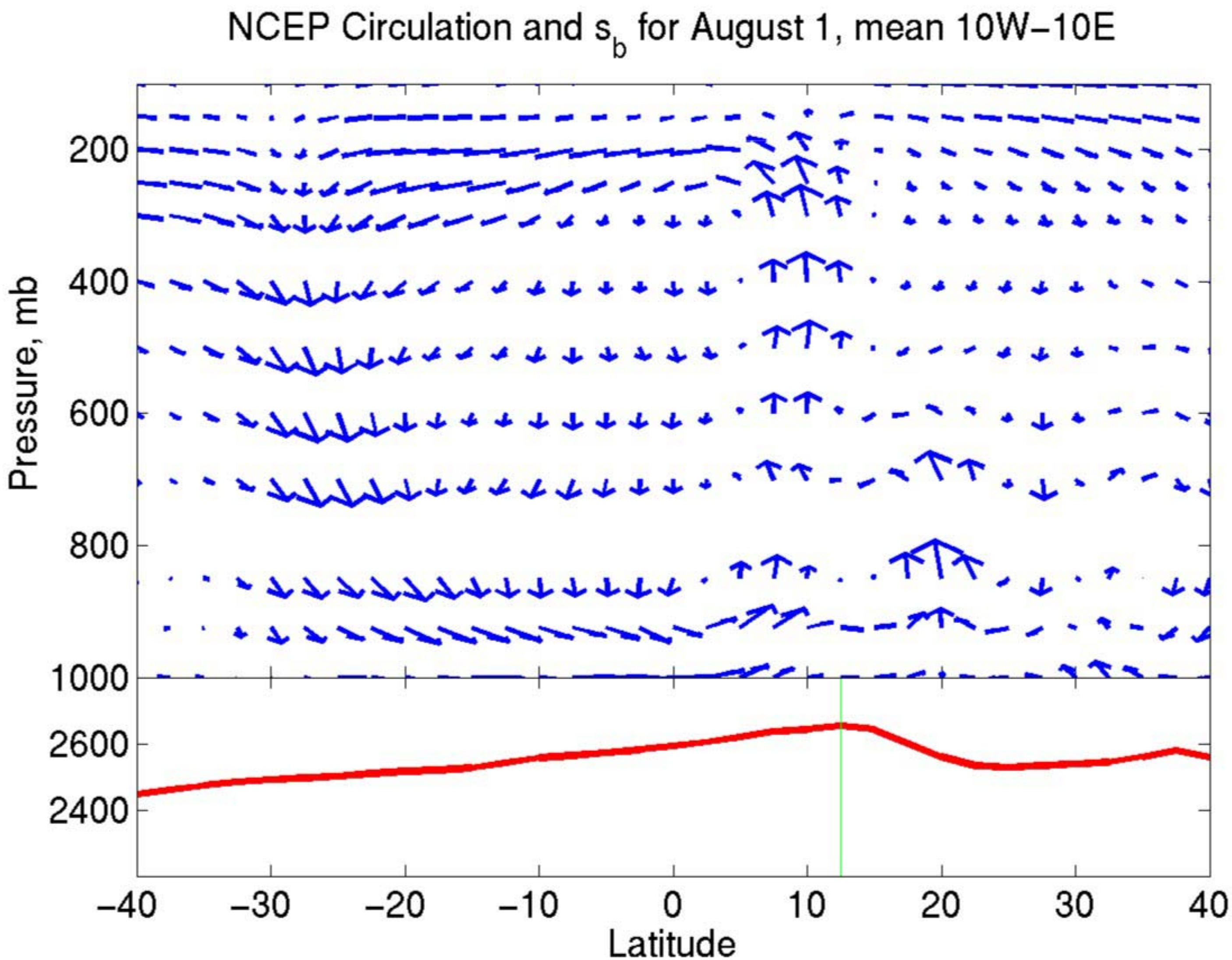


Factors that affect s_b

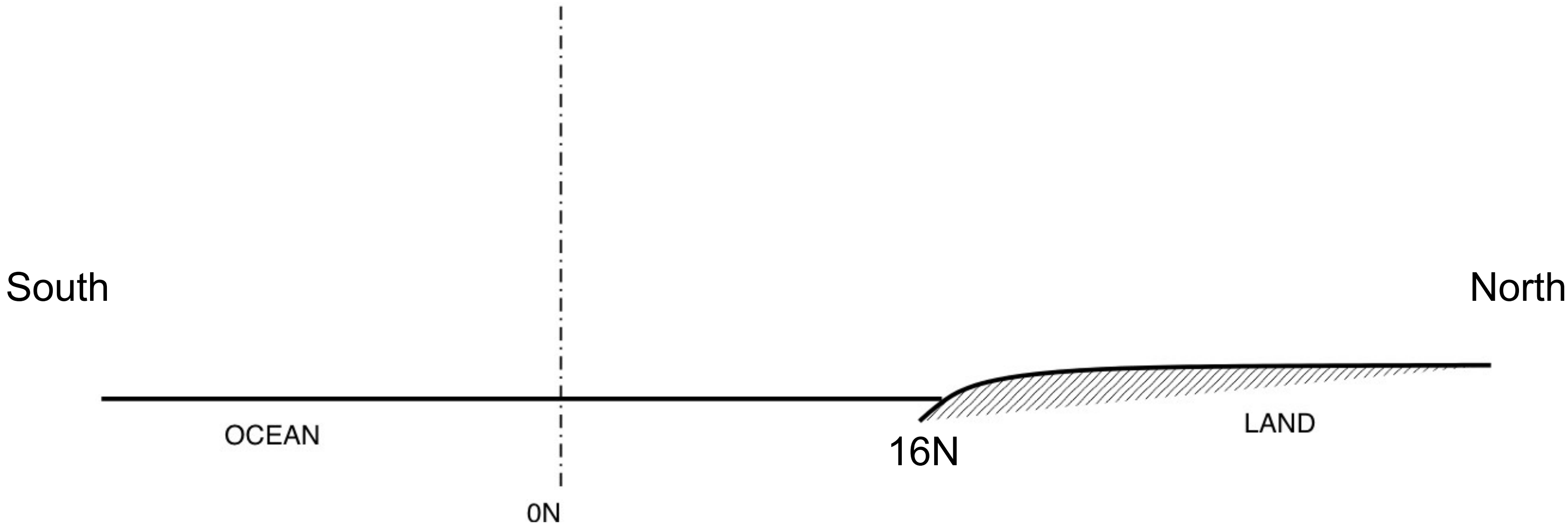
- Surface heat fluxes
- Evaporation of precipitation in convective downdrafts
- Radiative cooling
- Entrainment at the top of the subcloud layer
- Advection by large-scale flow

Circulation may have a strong impact on the subcloud s_b distribution through these feedbacks.

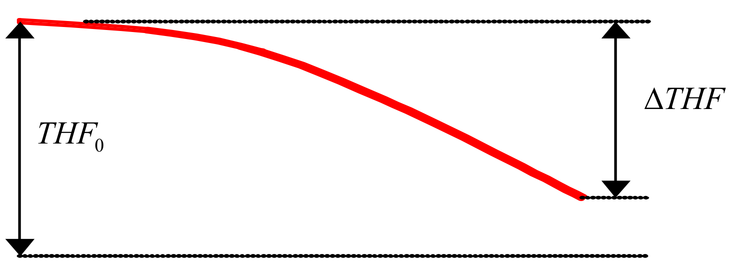
Observed circulation and subcloud s_b



Model Setup



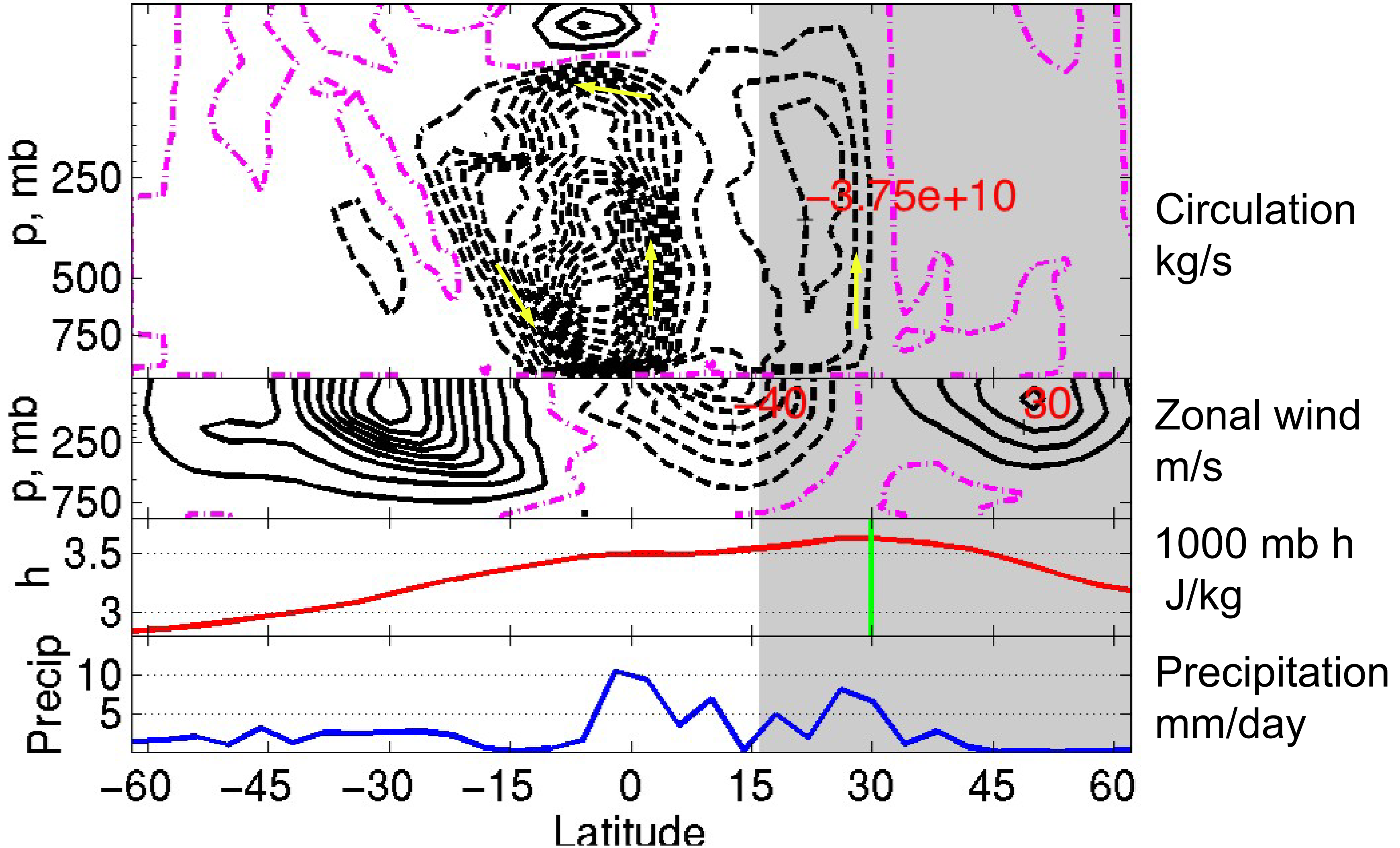
Over land, the surface forcing is determined by



$$THF(\phi) = LHF(T_S) + SHF(T_S)$$

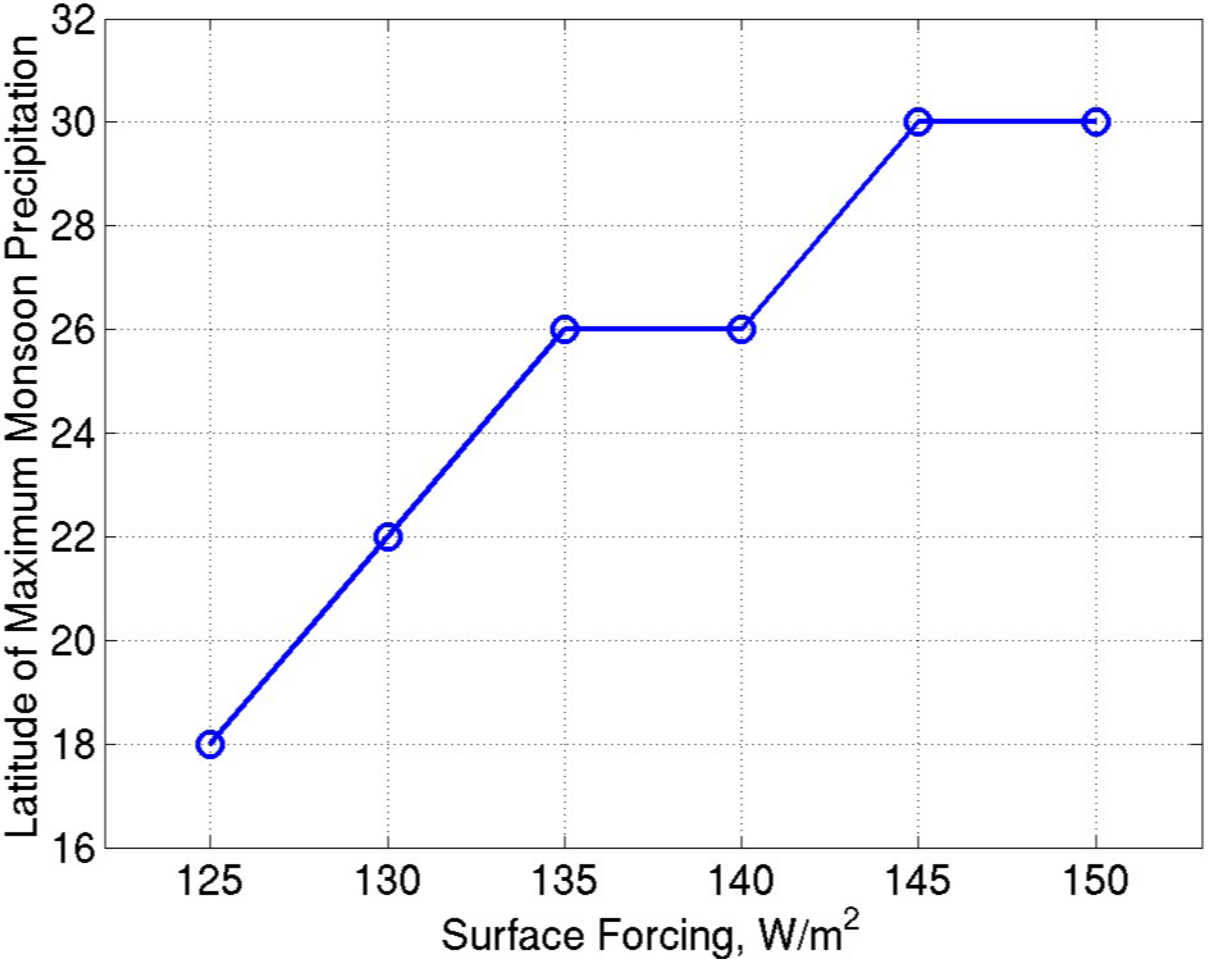
$$THF(\phi) = THF_0 - \Delta THF \sin^2(\phi - \phi_0)$$

2D Monsoon



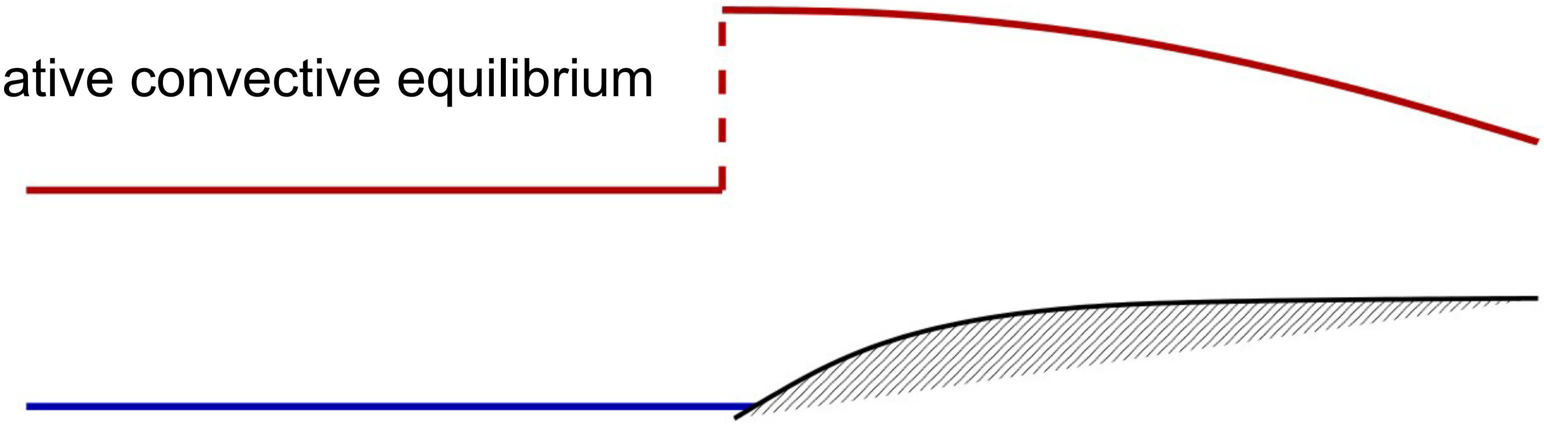
p, mb
250
500
750
−3.75e+10
Circulation kg/s
p, mb
250
750
−40
30
Zonal wind m/s
h
3.5
3
1000 mb h J/kg
Precip
10
5
Precipitation mm/day
−60 −45 −30 −15 0 15 30 45 60
Latitude

Monsoon Latitude

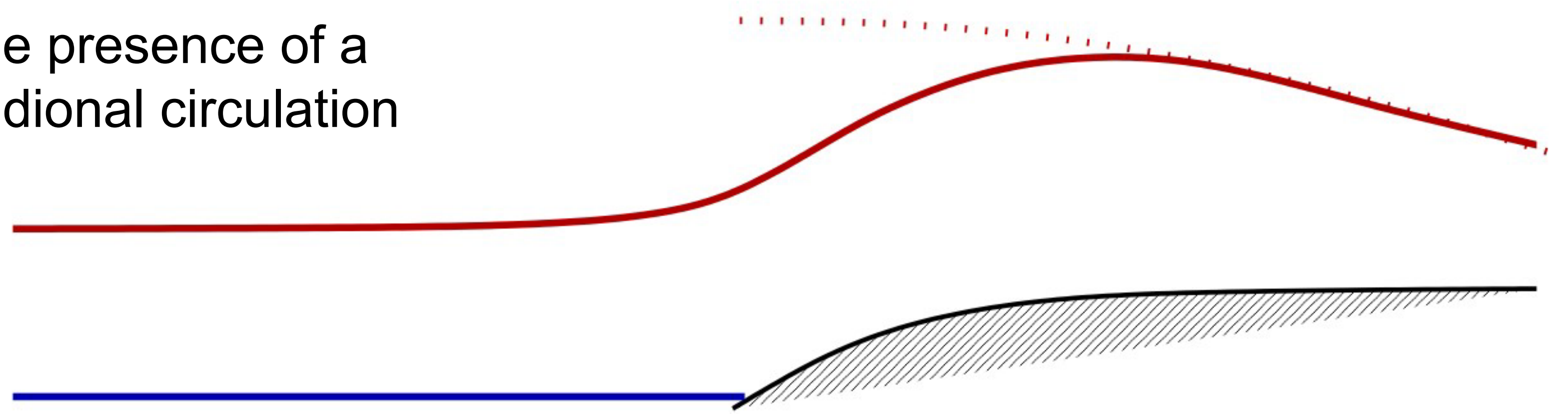


Impact of flow on h_b

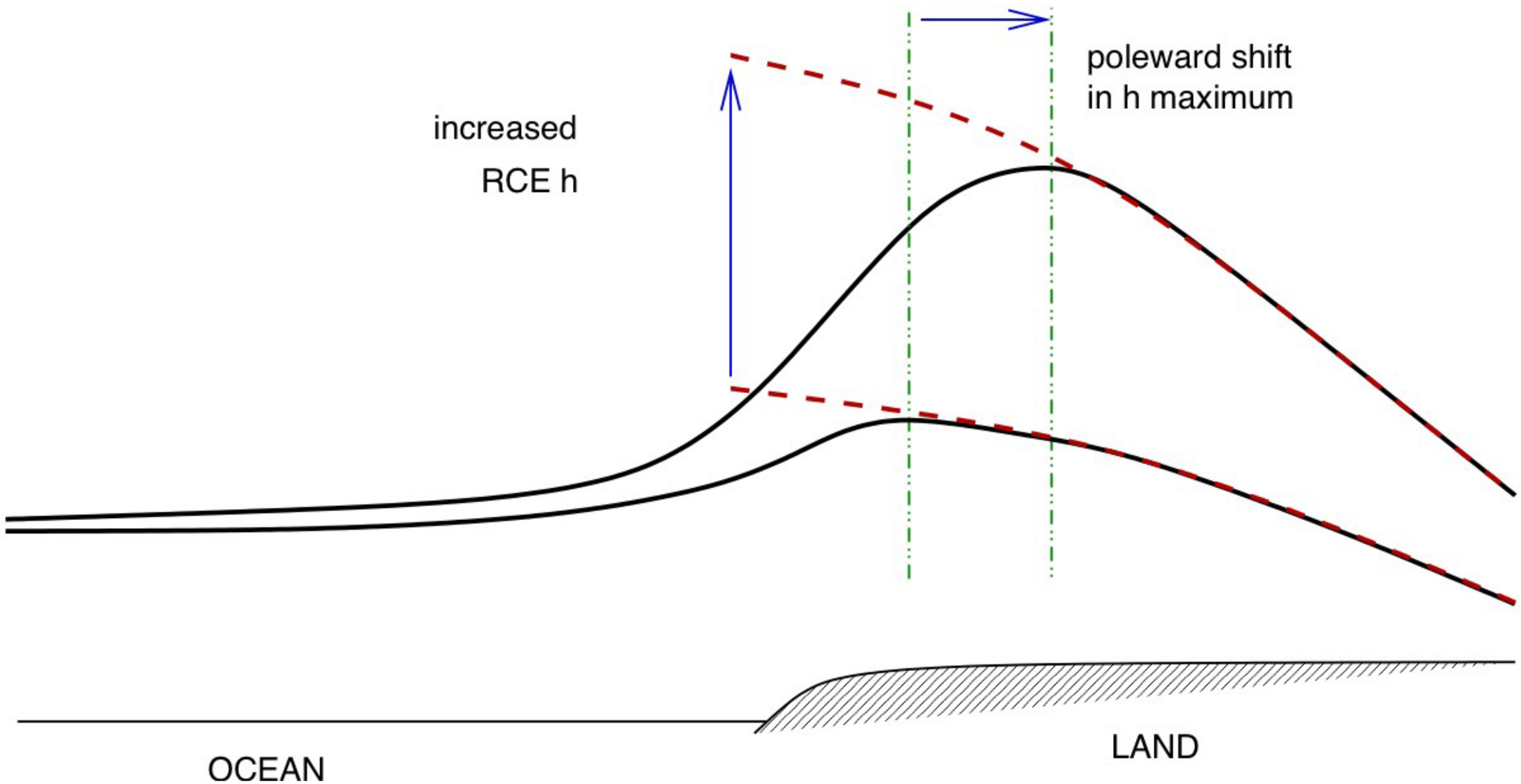
Radiative convective equilibrium



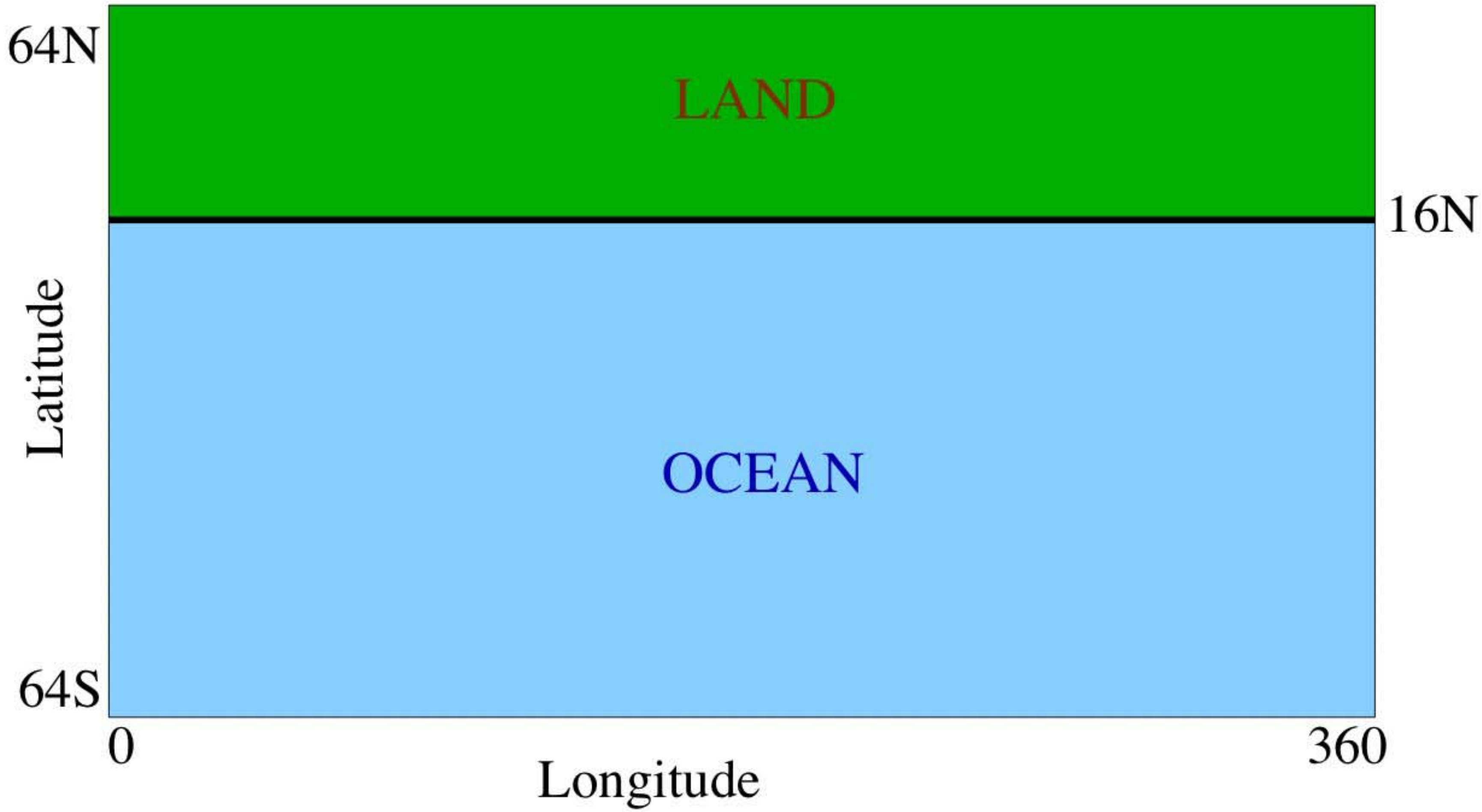
In the presence of a meridional circulation



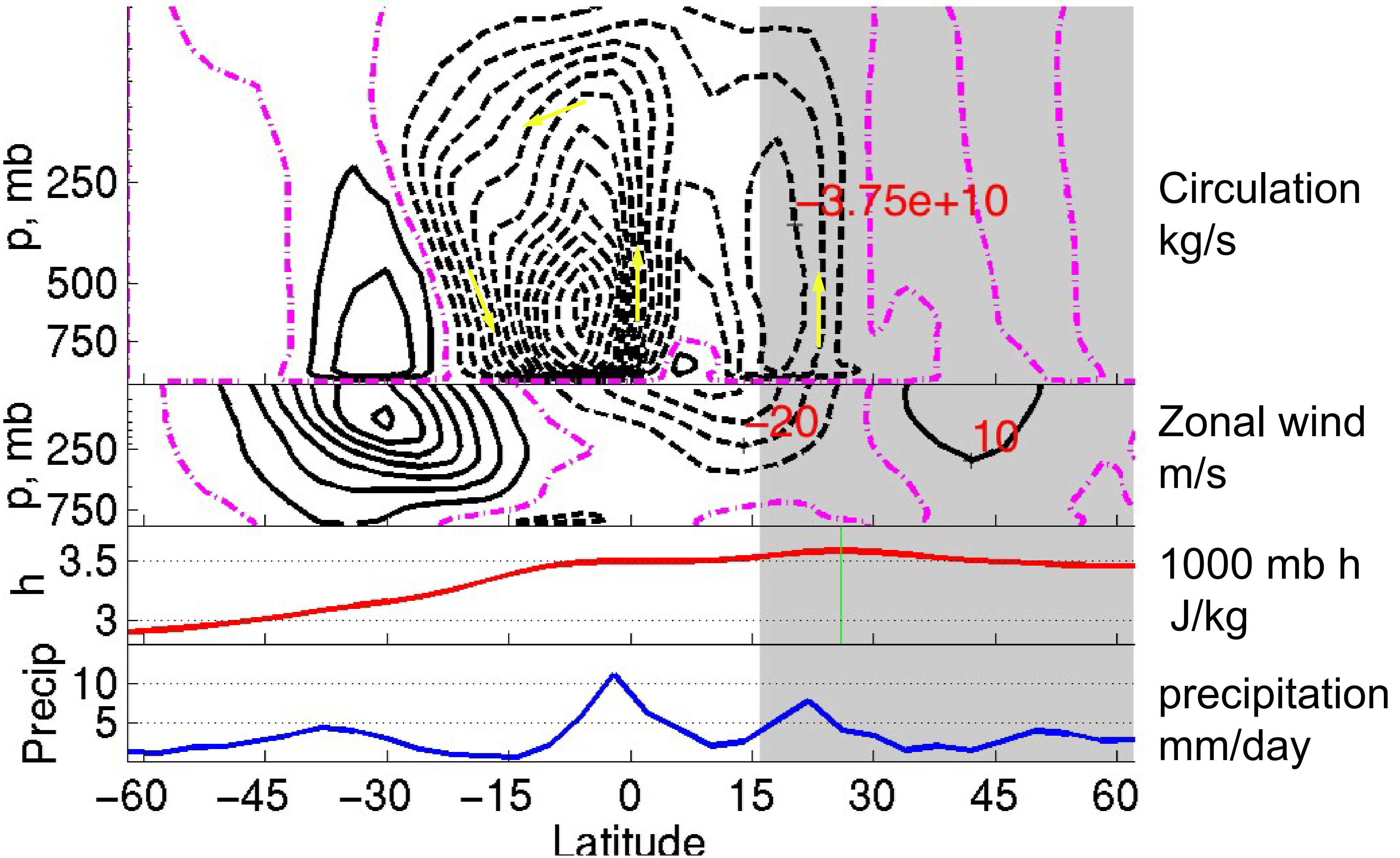
So what is going on with h_b?



Expand to 3D

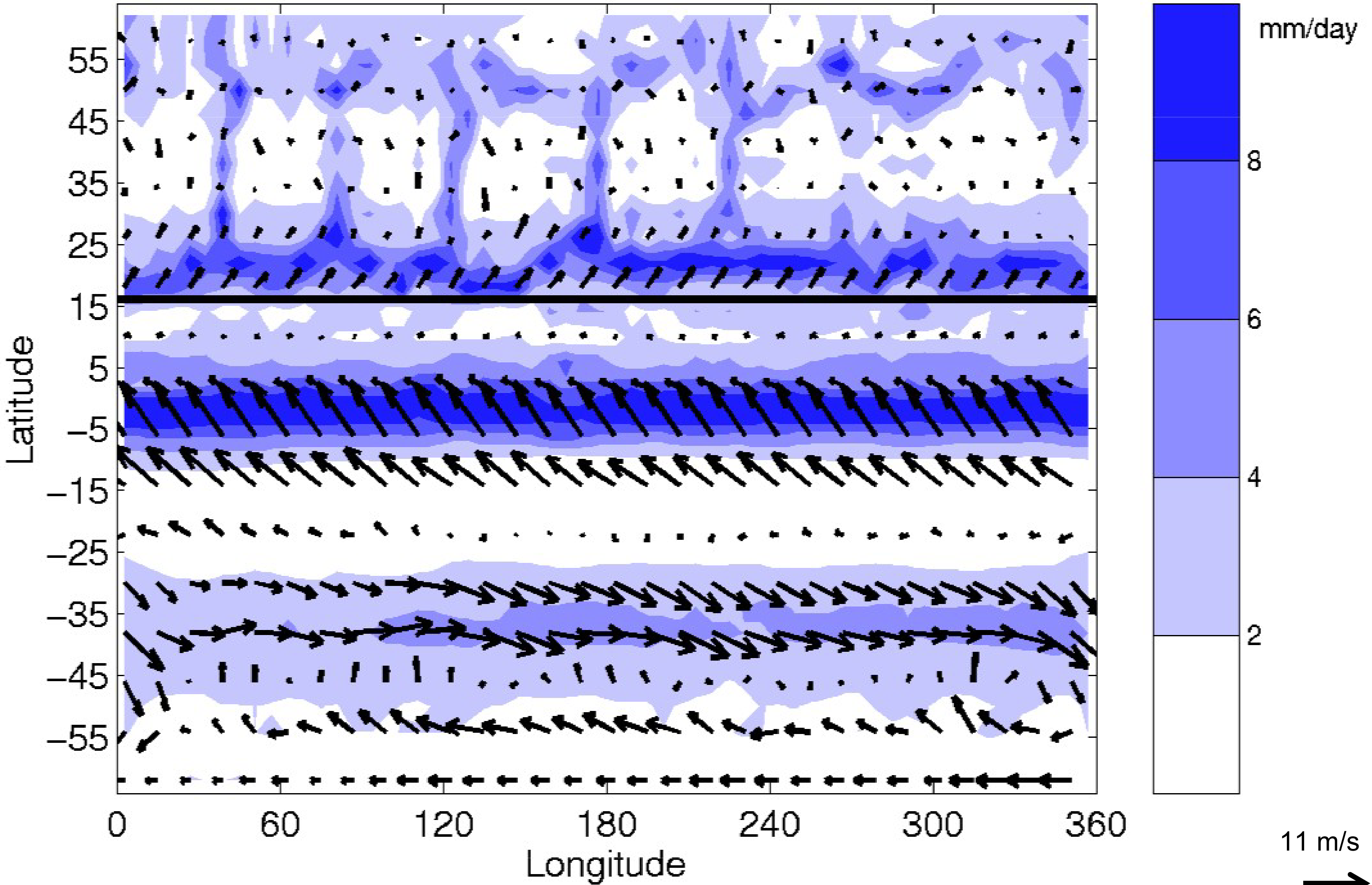


3D Monsoon

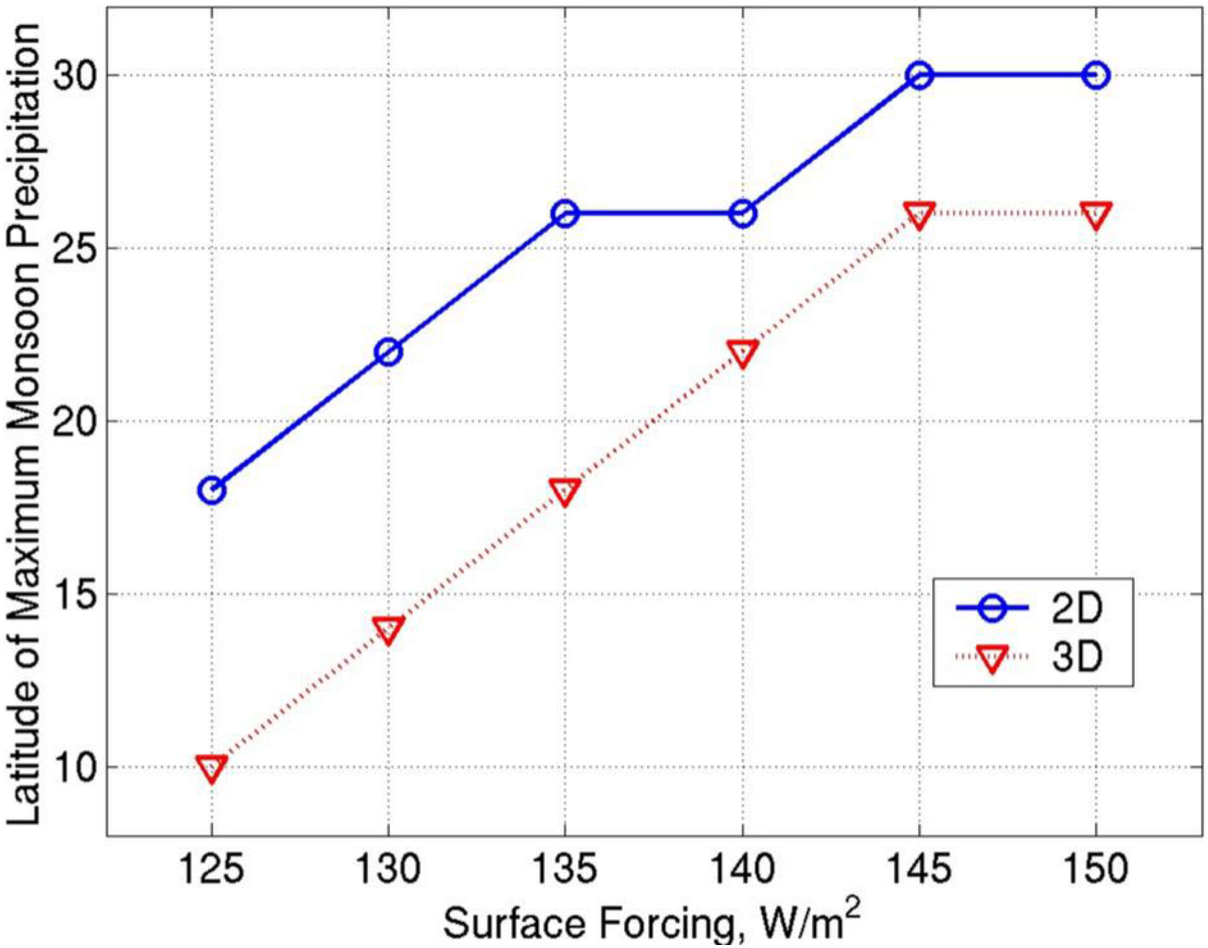


p, mb
250
500
750
−3.75e+10
p, mb
250
750
−20
10
h
3.5
3
Precip
10
5
−60
−45
−30
−15
0
15
30
45
60
Latitude
Circulation kg/s
Zonal wind m/s
1000 mb h J/kg
precipitation mm/day

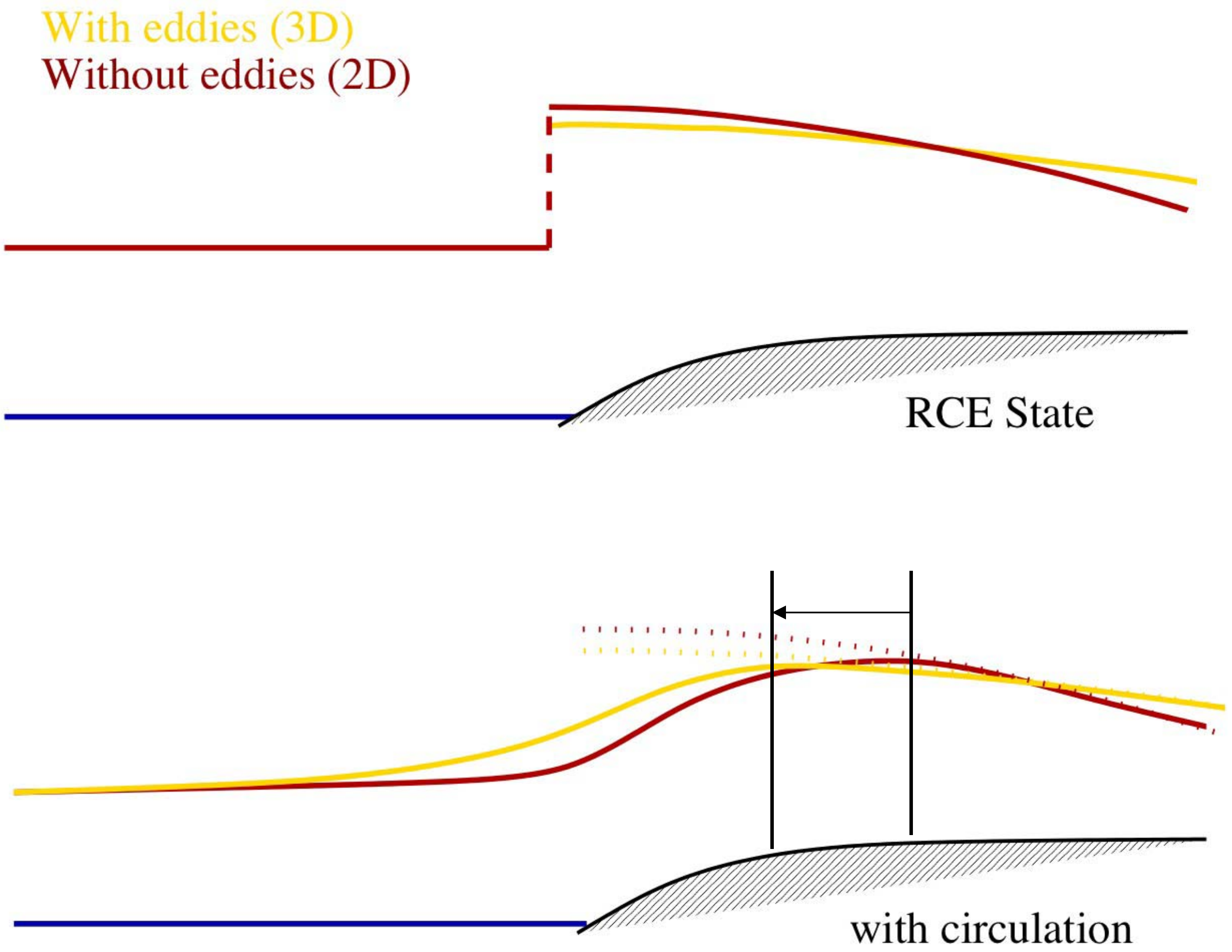
1000 mb winds and precipitation



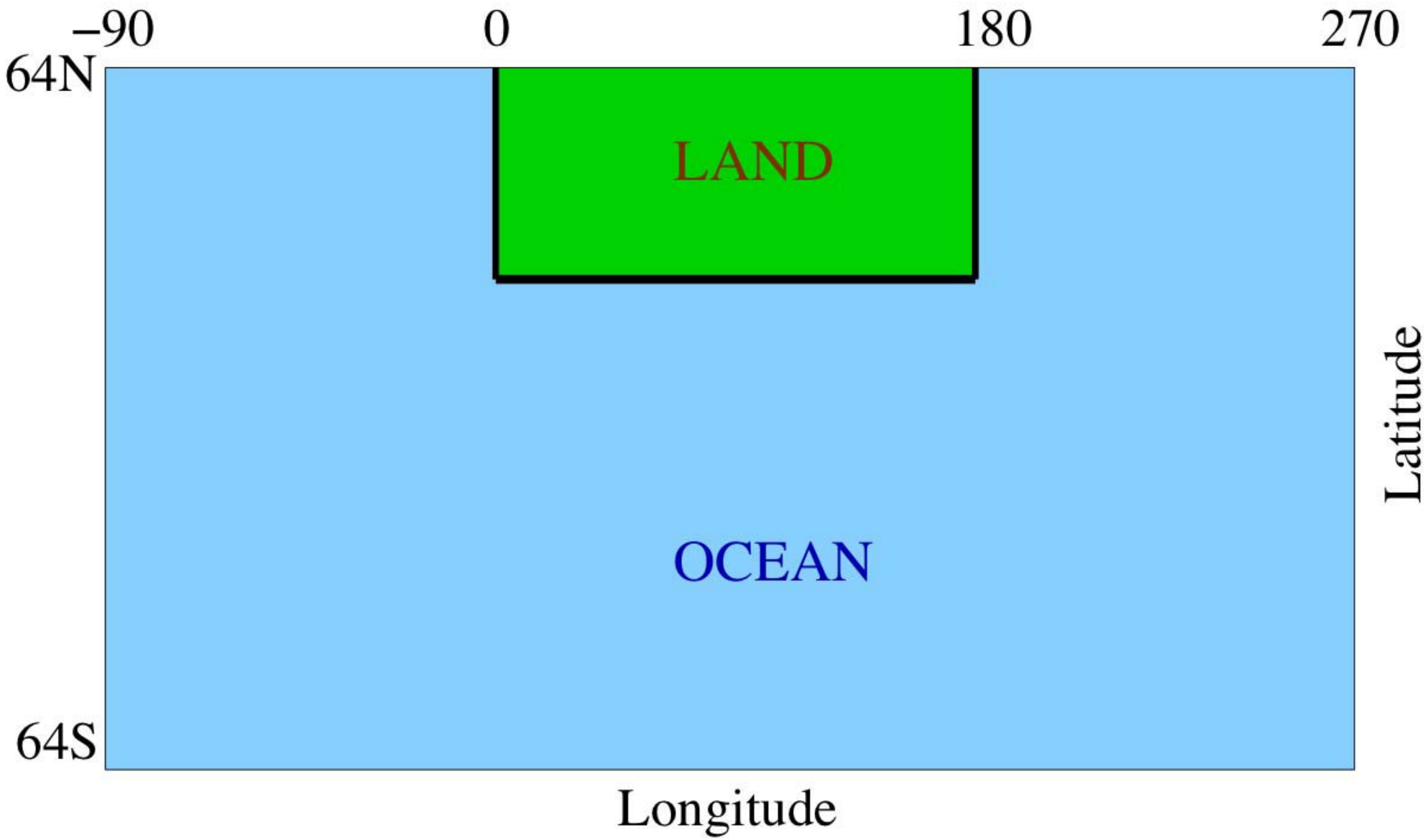
Monsoon Latitude: 2D vs 3D



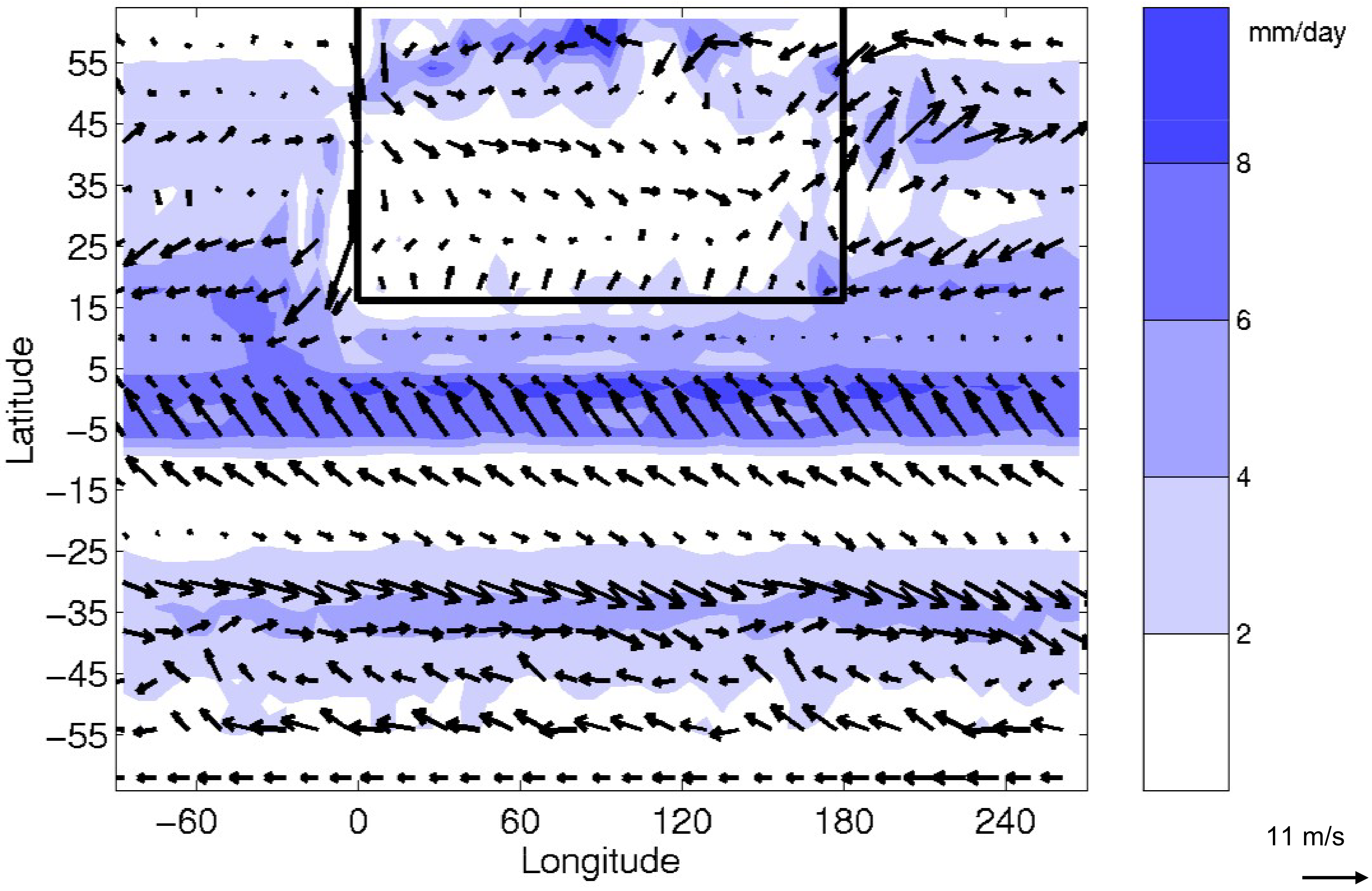
Impact of eddies on subcloud h



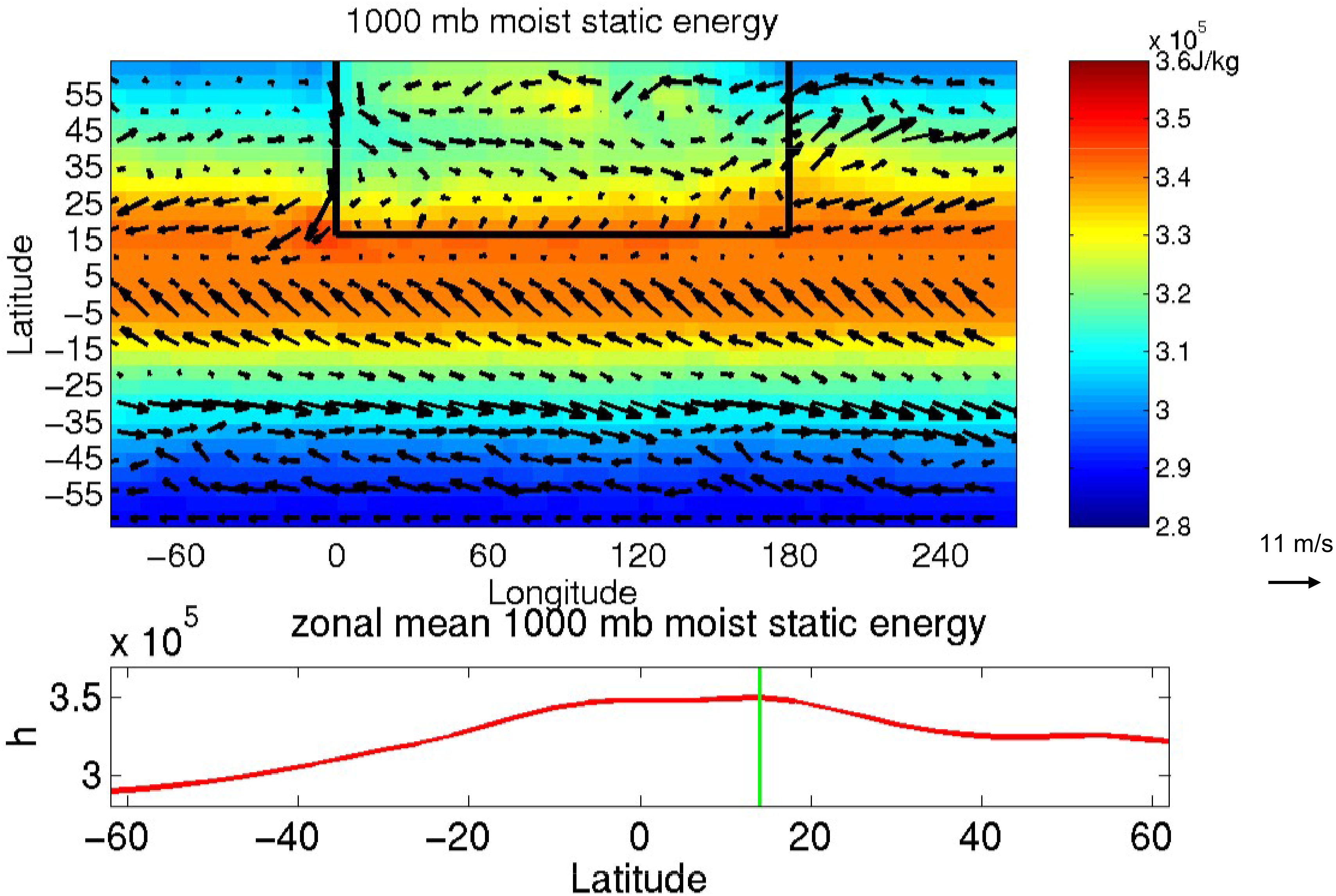
Introduce continental asymmetry



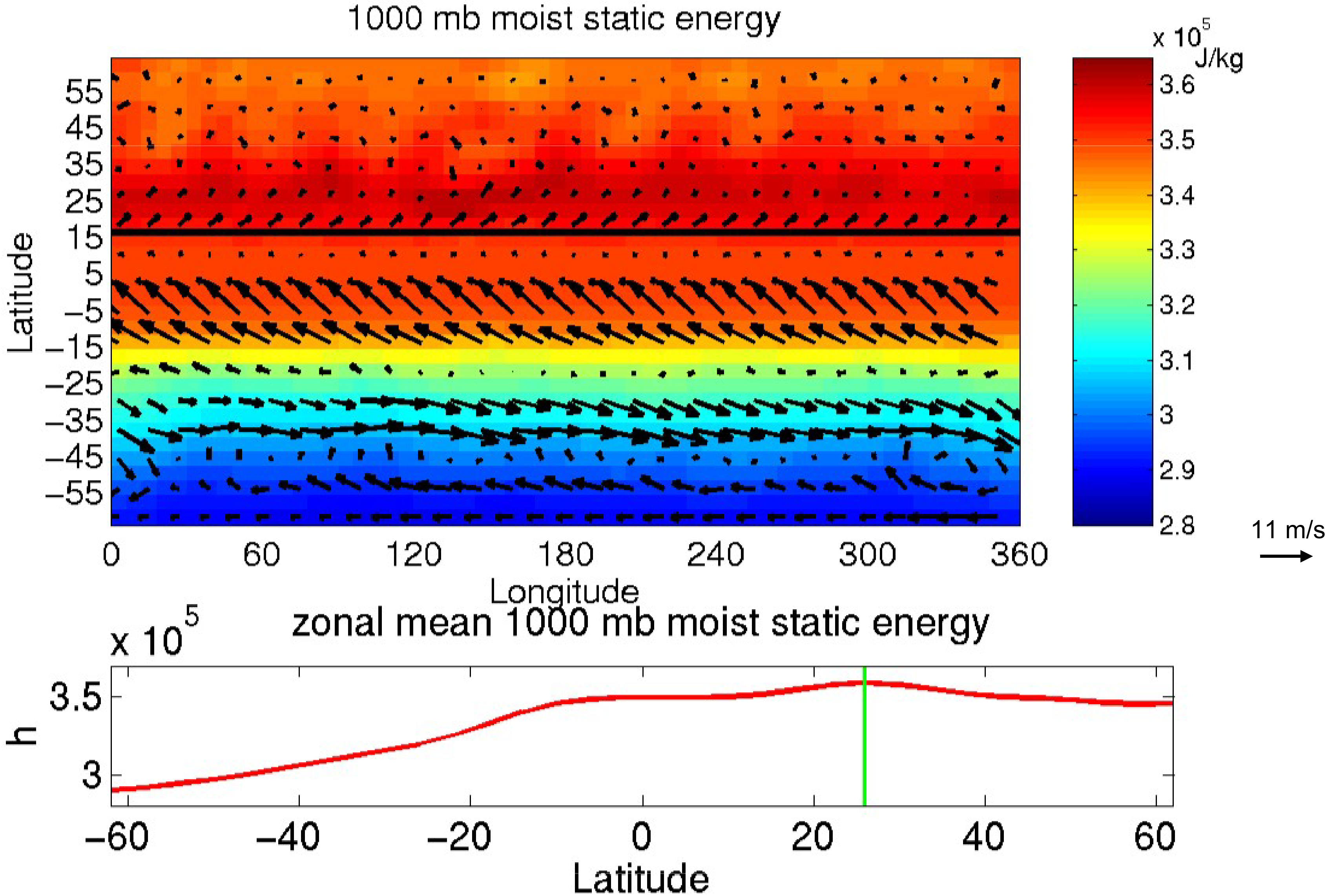
What happened to the monsoon?



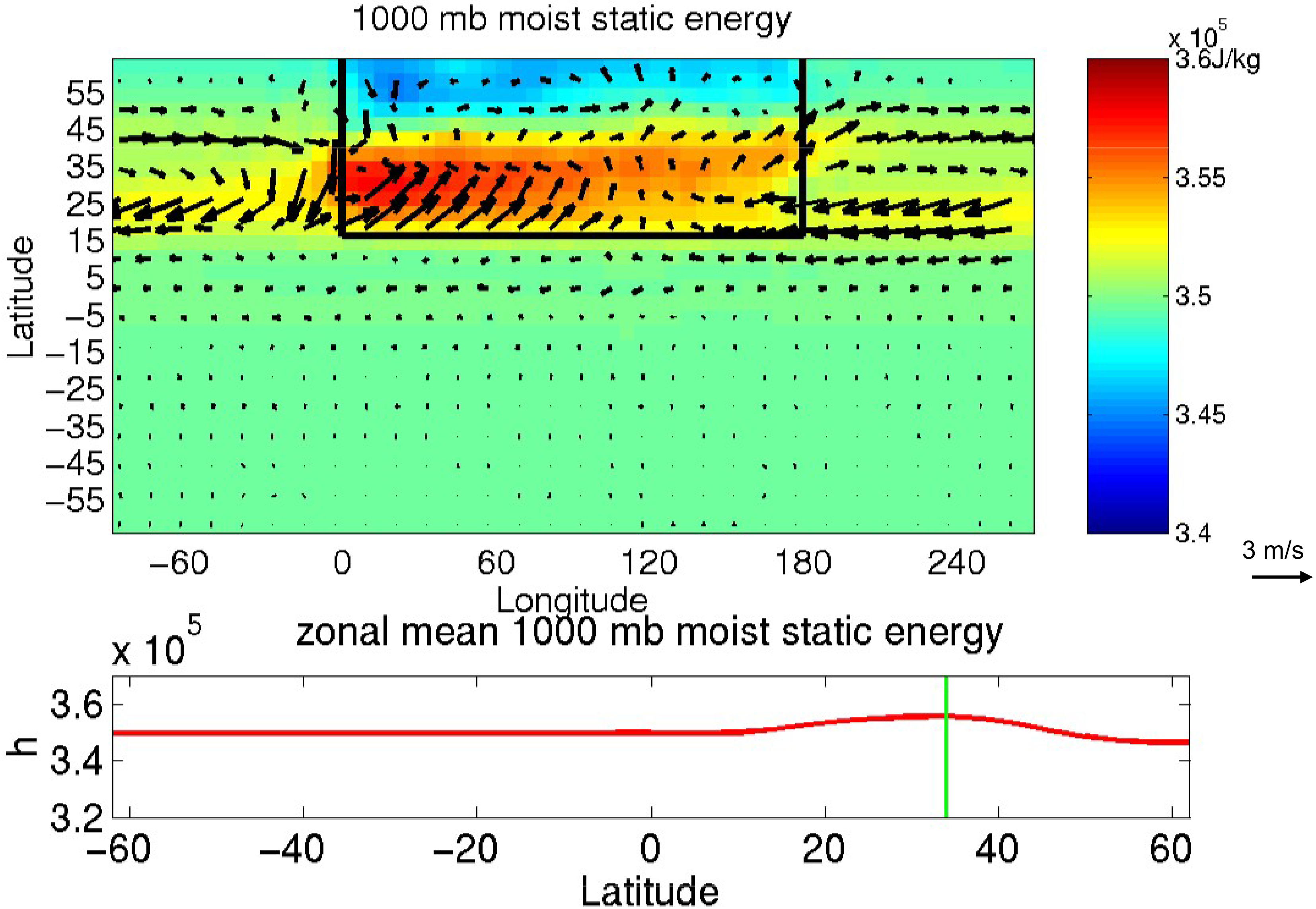
Impact of advection of low h_b air



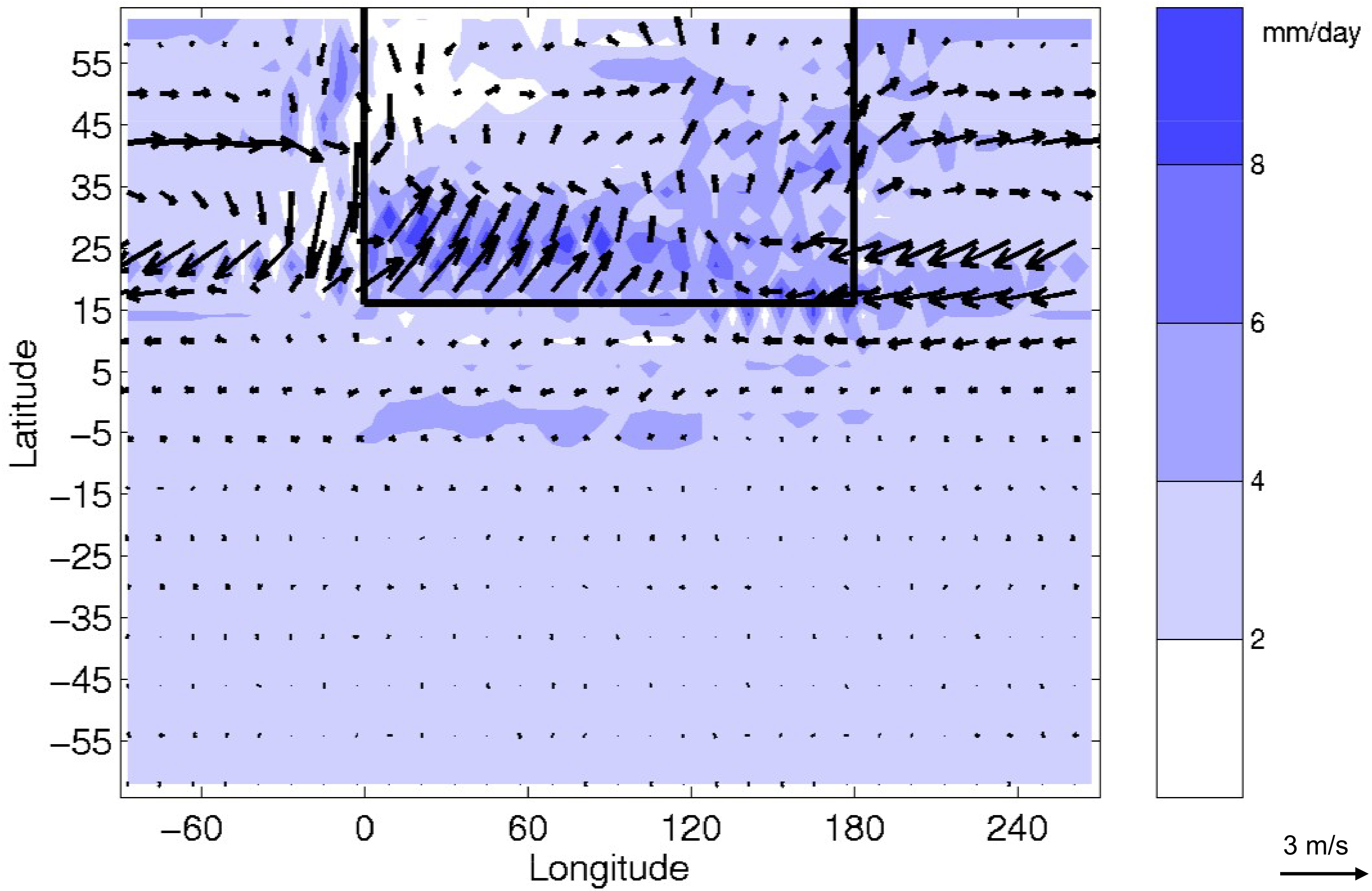
Comparison h_b distribution



Subcloud h_b with warm ocean



Remove the source of low h_b…



Extension of Held-Hou to 3 Dimensions

Assume *geostrophic* balance:

$$\frac{\partial \mathbf{V}}{\partial p} = -\frac{1}{2\Omega \sin\theta} \hat{k} \times \nabla_p \alpha$$

$$= -\frac{1}{2\Omega \sin\theta} \left(\frac{\partial T}{\partial p} \right)_{s*} \hat{k} \times \nabla_p s*$$

Integrate upward, taking **V**=0 at surface:

$$\mathbf{V}_T = \frac{1}{2\Omega \sin\theta} \left(T_s - T_T \right) \hat{k} \times \nabla s_b$$

Demand that absolute vorticity at tropopause have the same sign as *f*:

$$\sin\theta\left[4\Omega^2\sin\theta+\nabla\bullet\left(\frac{1}{\sin\theta}\left(T_s-T_T\right)\nabla s_b\right)\right]\geq 0$$