Dynamic Meteorology

(WAPE: General Circulation of the Atmosphere and Synoptic meteorology)

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2) General circulation of the neutral atmosphere

a) Zonal mean climatology of $\overline{u}$ and $\overline{T}$ (cont. lecture 1)

b) Origin of the midlatitude and high latitude jets

Conservation of angular momentum

Toy-model 1

c) Trade winds and monsoonal flows in the tropics

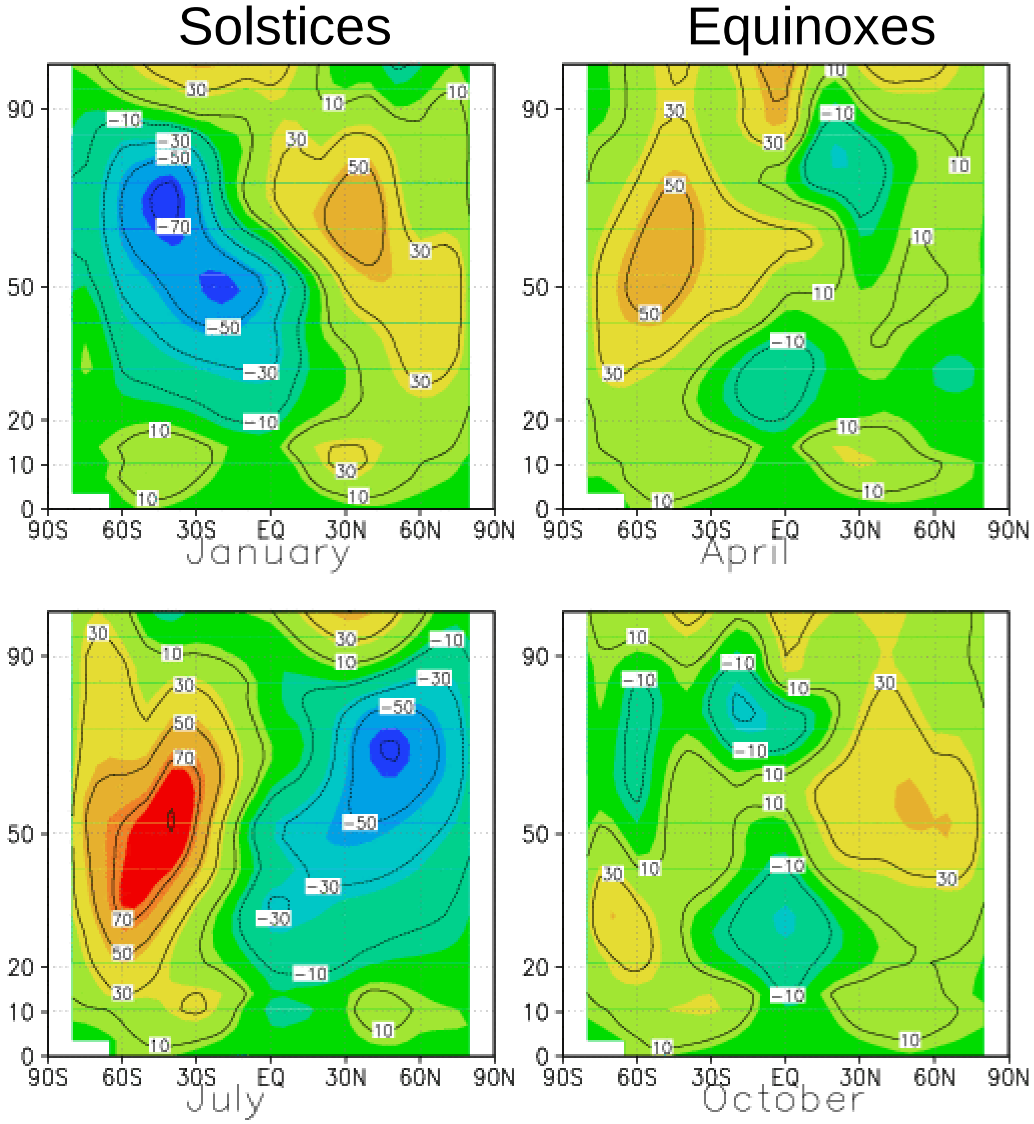
Toy-model 1b

a) Zonal mean climatologies of $\bar{u}$ and $\bar{T}$ (cont. lecture 1)

$$\bar{u} = \frac{1}{2\pi}\int_0^{2\pi} u\, d\lambda$$

Zonal mean zonal wind climatologies (CIRA dataset)

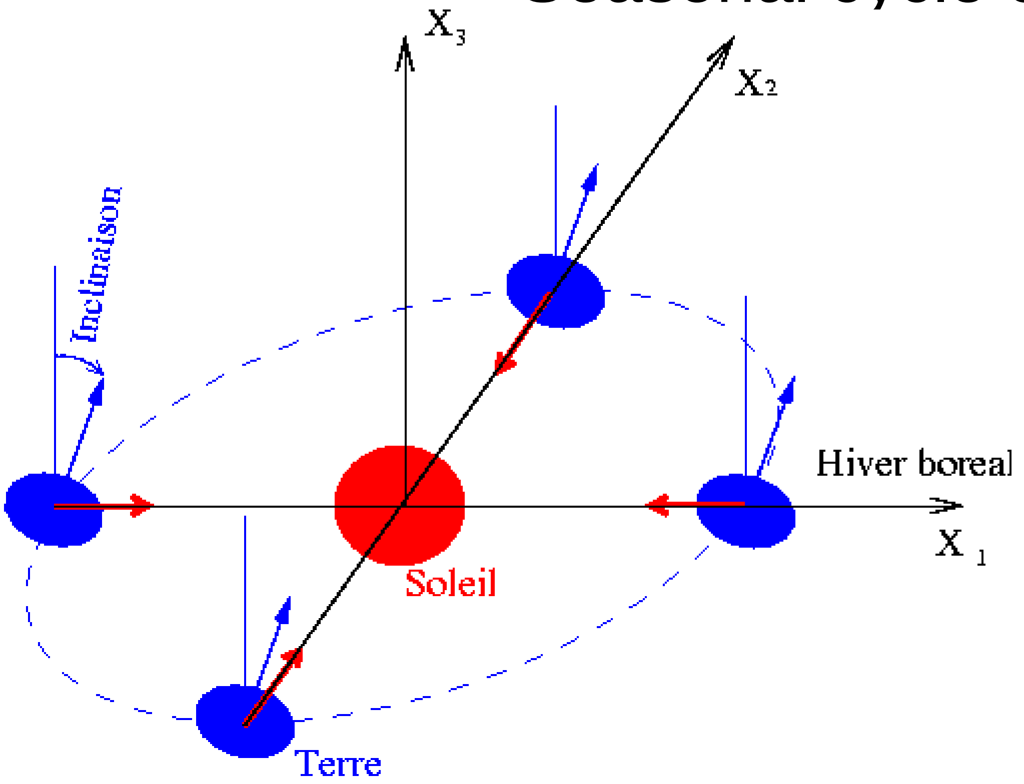
$\bar{u}$(m/s)

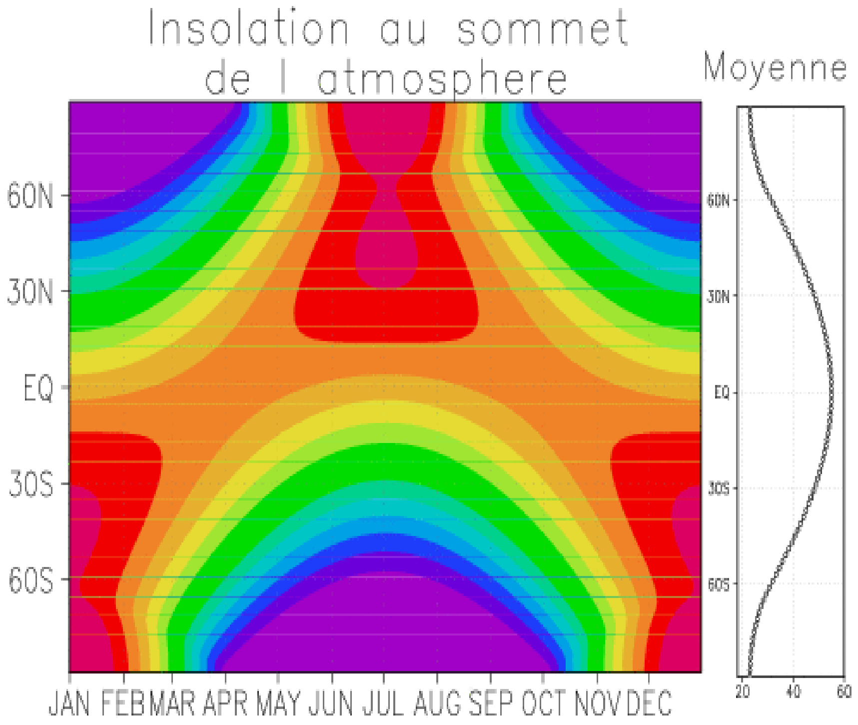


- In all seasons there are two westerly jets near below the subtropical tropopause. These westerlies extent almost down to the surface (0-16km) and characterize the midlatitude circulations.
- Still in troposphere, the winds tend to be slightly westward (easterly) in the tropics.
- In the middle atmosphere (20-90km), the winds are eastward (westerlies) in the winter hemisphere and westward in the summer hemisphere.
- In spring and fall the middle atmosphere jets are eastward in both hemisphere (equinox).
- Note, that during the winters, the jets in the southern hemisphere (July) are stronger than in the northern hemisphere (January).

a) Zonal mean climatologies of $\overline{u}$ and $\overline{T}$ (cont. lecture 1)

Seasonal cycle of the solar flux





- O3 re-emit almost instantaneously producing a chemical heating, the UV it absorbs
- The solar flux is maximum at the poles in summer, in part because the length of the day is 24h there
- Averaged over the year, the solar flux is maximum at the equator

a) Zonal mean climatologies of $\overline{u}$ and $\overline{T}$ (cont. lecture 1)

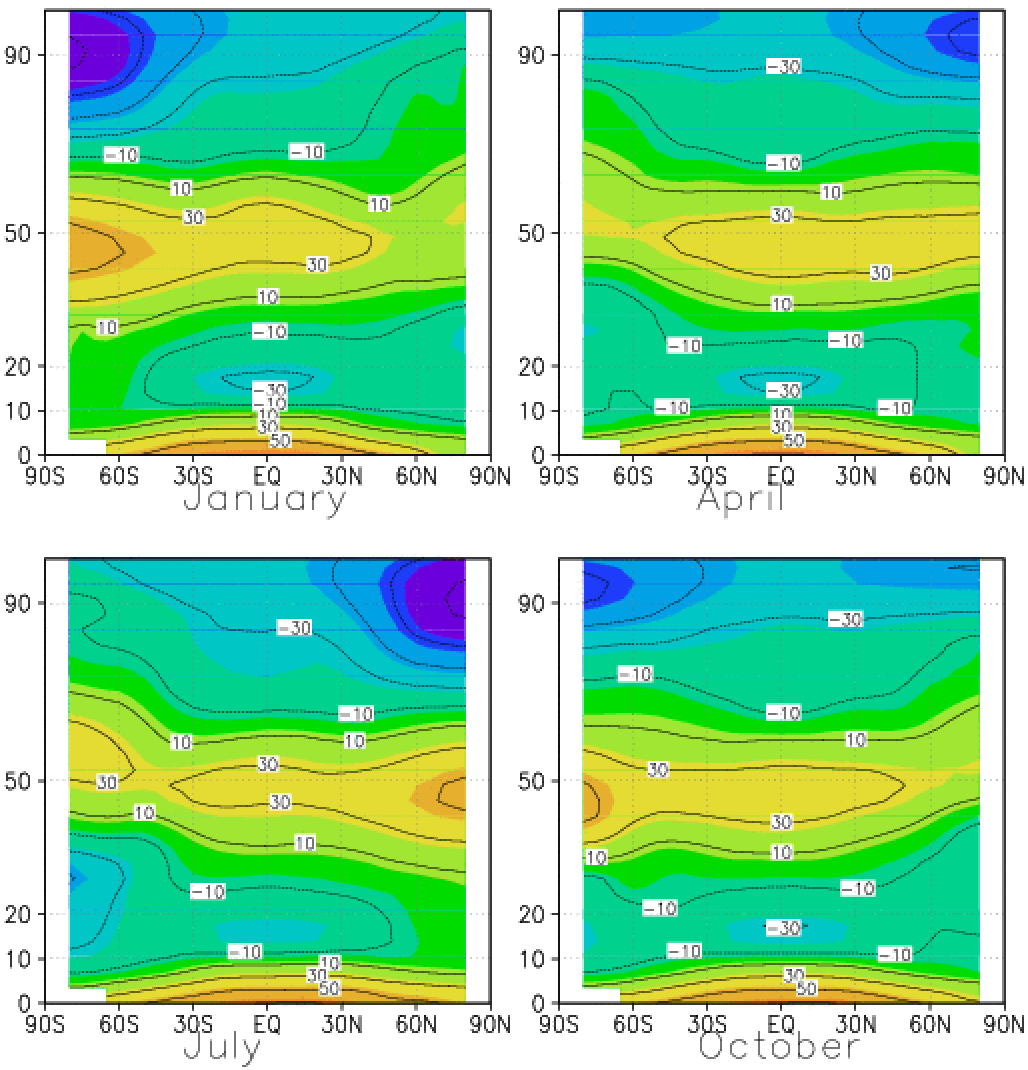
$$\overline{T} = \frac{1}{2\pi}\int_0^{2\pi} T\, d\lambda$$

Temperature (CIRA dataset)

$\overline{T}$(K)-230

Solstices

Equinoxes

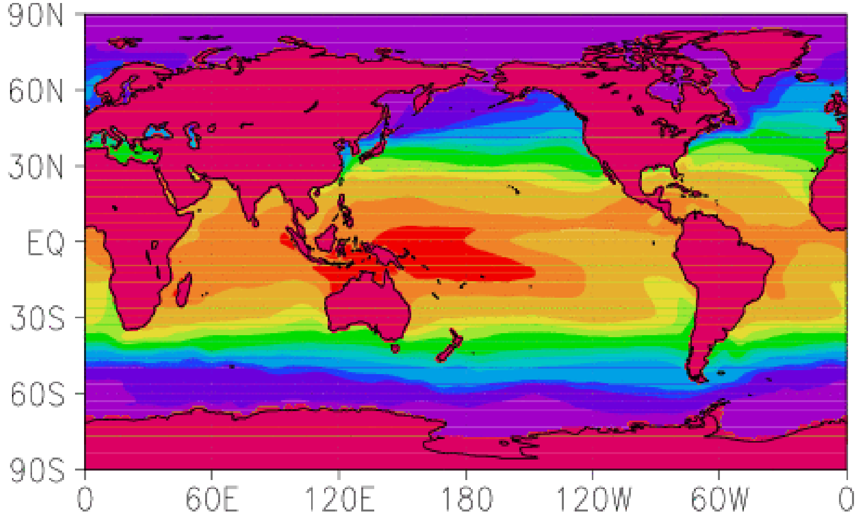


- Temperature decays with altitude in the troposphere.
- There is a minimum at the tropical tropopause (a greenhouse effect due to the presence of water vapour).
- In the stratosphere (20km<z<50km), T decreases from the summer pole to the winter pole.
- At the stratopause (50km) in the summer hemisphere, there is a max in T.
- During solstices and in the upper mesosphere (70-90km) T increases from the summer pole to the winter pole!
- Still in the solstices and at the mesopause, (90km) there are pronounced minima in T (~180K) over the summer pole!!

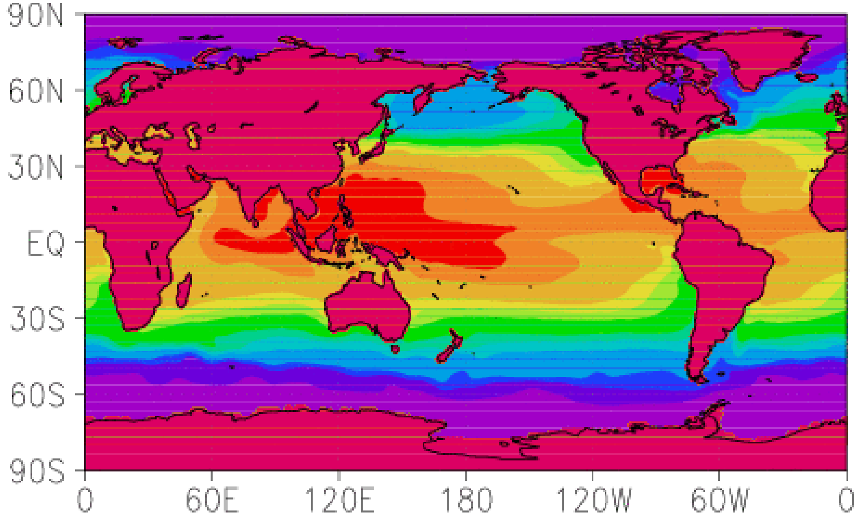
a) Zonal mean climatologies of $\overline{u}$ and $\overline{T}$ (cont. lecture 1)

The heat capacity of the ocean is very large, it allows the oceans to integrate the solar cycle. The thermal forcing is then in the IR, and absorbed by H2O and CO2 before reaching the middle atmosphere (greenhouse effect)

Temperature Surface de la mer
Janvier (ECMWF 1993–1997)



Juillet (ECMWF 1993–1997)



- The SST is always warmer in the equatorial regions
- It maintains a large humidity in the tropical regions, yielding a large greenhouse effect there.
- The troposphere is essentially forced from below, and will experience a less dramatic annual cycle than the middle atmosphere

b) Origin of the midlatitude and high latitude jets

Conservation of angular momentum

The « spherical » and Coriolis terms in the zonal momentum equation, can be re-written (Starting from the Primitive Eqs. In log pressure altitude) :

$$\frac{Du}{Dt}-\frac{uv}{a}\tan\phi-2\Omega\sin\phi\, v=\frac{-1}{a\cos\phi}\frac{\partial\Phi}{\partial\lambda}+X$$

$$\frac{D}{Dt}\underbrace{\left(a\cos\phi\, u+a^2\Omega\cos^2(\phi)\right)}_{m}=-\frac{\partial\tilde{\Phi}}{\partial\lambda}+\underbrace{a\cos\phi\, X}_{Torque}$$

Friction

Pressure

But we are interested in zonal means :
The pressure term disappears and the frictional torque stays :

Zonal mean $$\frac{\overline{Dm}}{Dt}=a\cos\varphi\,\overline{X}$$

But this is not a equation for $\overline{m}$ advected by a « mean meridional» circulation !

$$\frac{\overline{Dm}}{Dt}=\underbrace{\frac{\partial\overline{m}}{\partial t}+\frac{\overline{v}}{a}\frac{\partial\overline{m}}{\partial\phi}+\overline{w}\frac{\partial\overline{m}}{\partial z}}_{\frac{\overline{D}\overline{m}}{dt}}+\underbrace{\overline{\frac{u'}{a\cos\phi}\frac{\partial m'}{\partial\lambda}}+\overline{\frac{v'}{a}\frac{\partial m'}{\partial\phi}}+\overline{w',\frac{\partial m'}{\partial z}}}_{\text{Eddy forcing}}=\underbrace{a\cos\phi\,\overline{X}}_{\text{Mean torque}}$$

b) Origin of the midlatitude and high latitude jets

Conservation of angular momentum

Zonal mean and disturbance AAM: $\overline{m}=a^2\cos^2\phi\,\Omega+a\cos\phi\,\overline{u}$ Disturbances: $m'=m-\overline{m}=a\cos\phi\,u'$

Eddy forcing written in « flux » form, using mass conservation :

$$\rho_0\frac{\overline{D}\,\overline{m}}{Dt}=-\rho_0\overline{\frac{u'}{a\cos\phi}\frac{\partial m'}{\partial\lambda}}-\rho_0\overline{\frac{v'}{a}\frac{\partial m'}{\partial\phi}}-\rho_0\overline{w'\frac{\partial m'}{\partial z}}+\rho_0 a\cos\phi\,\overline{X}$$

Integration by « part » :

$$-\;m'\operatorname{div}\rho_0\vec{u}'=\frac{-m'}{a\cos\phi}\left(\frac{\partial}{\partial\lambda}\rho_0 u'+\frac{\partial}{\partial\phi}\rho_0\cos\phi\,v'\right)-m'\frac{\partial\rho_0 w'}{\partial z}=0$$

$$\rho_0\frac{\overline{D}\,\overline{m}}{Dt}=\frac{-1}{a\cos\phi}\frac{\partial}{\partial\phi}\rho_0\cos\phi\,\overline{v'm'}-\frac{\partial}{\partial z}\rho_0\overline{w'm'}\;+\;a\,\rho_0\cos\phi\,\overline{X}$$

Meridional flux of AAM

Vertical flux of AAM

<u>Eulerian mean formalism for the interaction between waves and the zonal mean wind :</u>

$$\frac{1}{a\cos\varphi}\frac{\overline{D}}{Dt}\overline{m}=\overline{u}_t+\overline{v}\left[(a\cos\varphi)^{-1}(\overline{u}\cos\varphi)_\varphi-f\right]+\overline{w}\,\overline{u}_z=\overline{X}-(a\cos^2\varphi)^{-1}\left(\overline{u'v'}\cos^2\varphi\right)_\varphi-\frac{1}{\rho_0}\left(\rho_0\overline{u'w'}\right)_z$$

Advection by the mean Meridional circulation

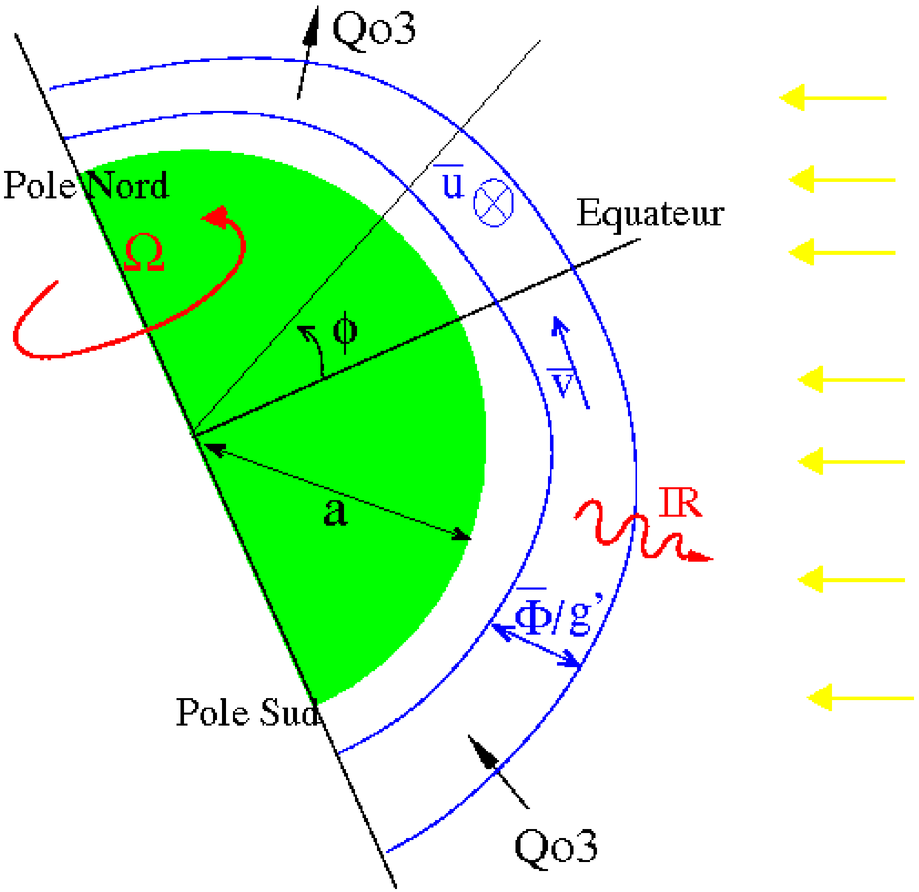
Rossby waves

Gravity waves

Equatorial waves in the tropics

b) Origin of the midlatitude and high latitude jets

Toy model 1: axisymmetric shallow water model on the sphere



Shallow water model with diabatic heatings :

$$\left(\frac{\partial}{\partial t}+\frac{\overline{v}}{a}\frac{\partial}{\partial\varphi}\right)\overline{u}-\left(2\Omega\sin\varphi+\frac{\tan\varphi}{a}\overline{u}\right)\overline{v}=0$$

$$\left(\frac{\partial}{\partial t}+\frac{\overline{v}}{a}\frac{\partial}{\partial\varphi}\right)\overline{v}+\left(2\Omega\sin\varphi+\frac{\tan\varphi}{a}\overline{u}\right)\overline{u}=-\frac{1}{a}\frac{\partial\overline{\Phi}}{\partial\varphi}$$

$$\frac{\partial\overline{\Phi}}{\partial t}+\frac{1}{a\cos\varphi}\frac{\partial\cos\varphi\,\overline{v}\,\overline{\Phi}}{\partial\varphi}=\overline{Q}-\overline{Q}_s-\alpha\left(\overline{\Phi}-\overline{\Phi}_s\right)$$

Angular momentum conservation :

$$\left(\frac{\partial}{\partial t}+\frac{\overline{v}}{a}\frac{\partial}{\partial\varphi}\right)\left(a\cos\varphi\,\overline{u}+a^2\cos^2\varphi\,\Omega\right)=0$$

Geostrophic Balance :

$$2\Omega\sin\phi\,\overline{u}=-\frac{1}{a}\frac{\partial\overline{\Phi}}{\partial\phi}$$

Thermal equilibrium :

$$\overline{Q}-\overline{Q}_s=-\alpha\left(\overline{\Phi}-\overline{\Phi}_s\right)$$

$\overline{\Phi}$ is the geopotentiel,
the depth of the layer is **Φ/g**',

g'~g is a reduced gravity.

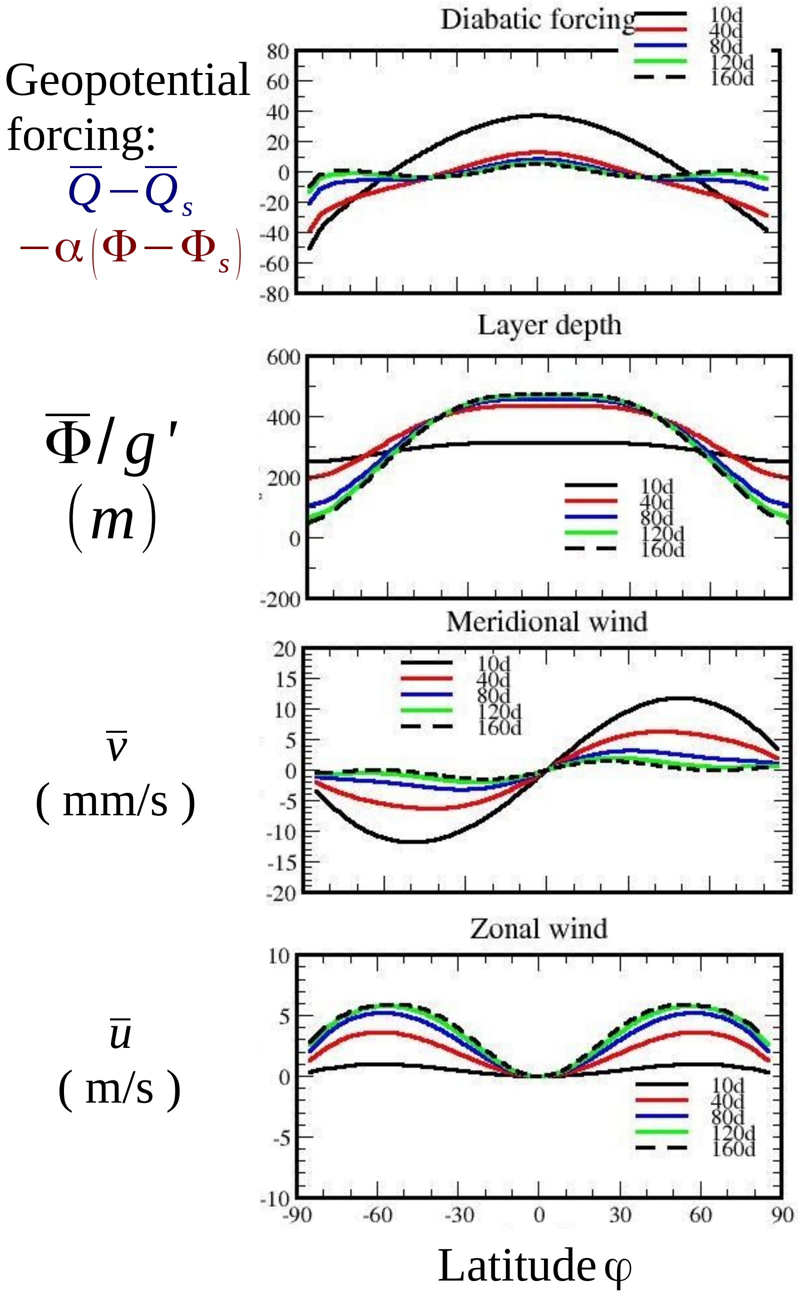
The Infrared Radiation (IR) characteristic
time scale is α~(1/5 jours).

b) Origin of the midlatitude and high latitude jets

Toy model 1 with max heating centered at the equator

Equinoxes in the middle atmosphere (03 UV Heating)

Upper troposphere all seasons (H20, CO2 IR Heating)



- At the beginning (10d) the diabatic forcing is due to O_3 only. It induces an increase of $\overline{\Phi}$ at the equator and a decrease in the mid and polar latitudes
- A radiative equilibrium between the diabatic Heating and the IR cooling is reached after 160d. The diabatic forcing is then very small.
- Initially, the heating induces a meridional motion ($\overline{v}$) toward the north in the NH, toward the south in the SH.
- $\overline{v}$ becomes very small at equilibrium (160d).

  Question the existence of the Hadley cells except in the transient cases?
- By angular momentum conservation, this meridional displacement produced positive zonal winds in both hemisphere.
- Note that meridional circulations have much smaller amplitudes than the zonal winds they produce ($\overline{v}$<<$\overline{u}$)
- Note that $\overline{u}$ is in geostrophic equilibrium with $\overline{\Phi}$ in the midlatitudes.

$$2\Omega \sin\varphi \bar{u} = -\frac{1}{a}\frac{\partial \overline{\Phi}}{\partial \varphi}$$

b) Origin of the midlatitude and high latitude jets

Toy model 1 with max heating/cooling at the South/north pole

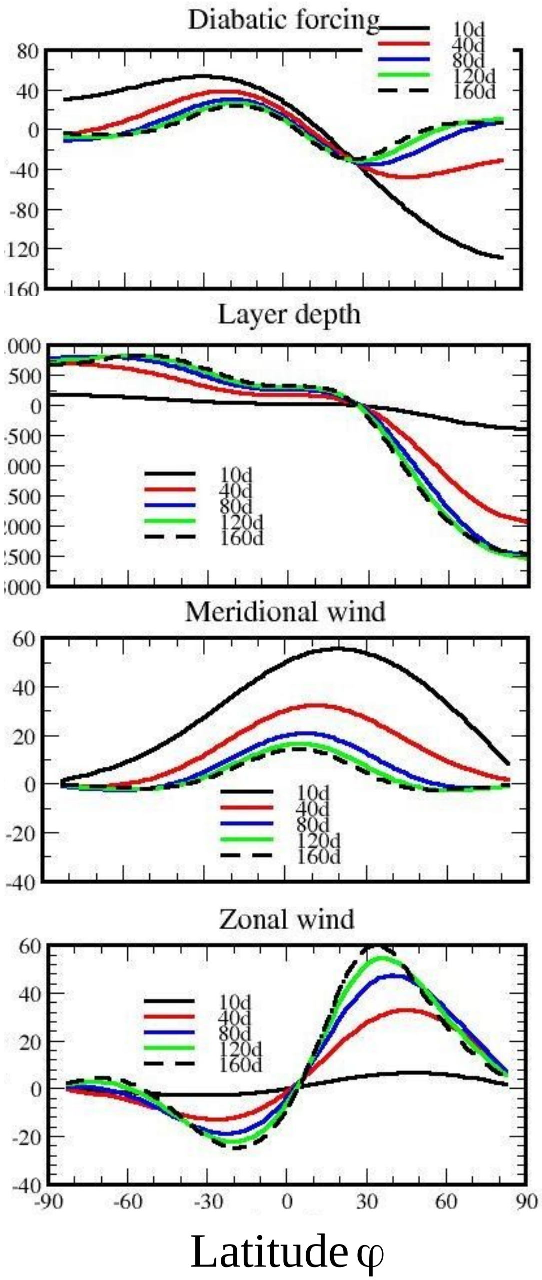
(NH winter in the middle atmosphere)

Geopotential forcing: $\overline{Q}-\overline{Q}_s$ $-\alpha\left(\Phi-\Phi_s\right)$

$\overline{\Phi}/g'$ $\left(m\right)$

$\overline{v}$ (mm/s)

$\overline{u}$ (m/s)



- At beginning (10d), the diabatic forcing due to O3 only. It induces an increase of Φ in the southern Hemisphere and a decrease in the Northern hemisphere.

- A radiative equilibrium between the forcing due to 03 and the IR cooling is reached after 160d.

- The initial forcing due to 03 induce a meridional displacement ($\overline{v}$). $\overline{v}$ becomes very small when we get near the steady state (160d).

- $\overline{v}$ becomes very small at equilibrium (160d).

Question the existence of the Brewer-Dobson cells except in the transient cases?

- By angular momentum conservation, these displacements induced negative zonal winds ($\overline{u}$) in the SH and positive zonal winds in the NH.

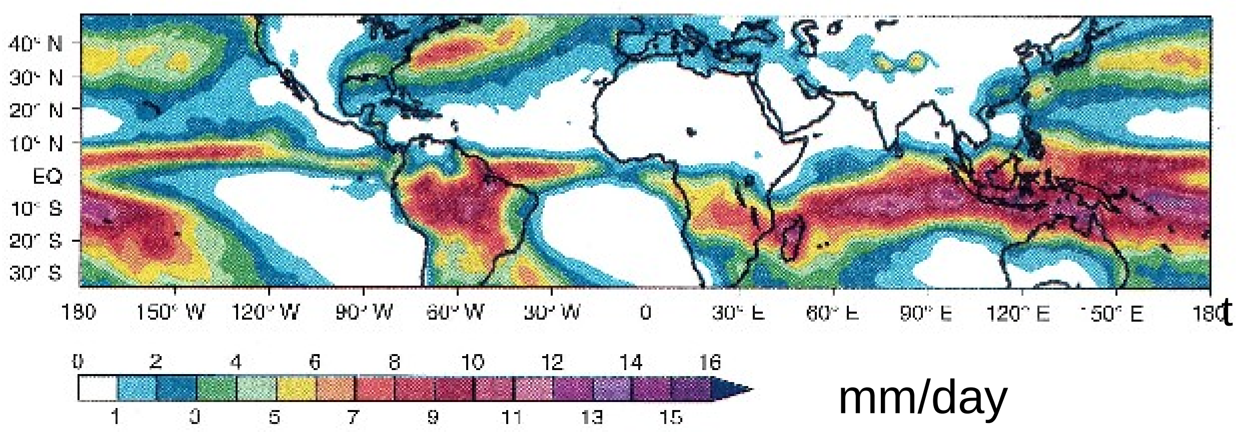
- Note that meridional circulations have much smaller amplitudes than the zonal winds they produce ($\overline{v}$<<$\overline{u}$)

- $\overline{u}$ is in geostrophic balance with Φ in the midlatitudes:

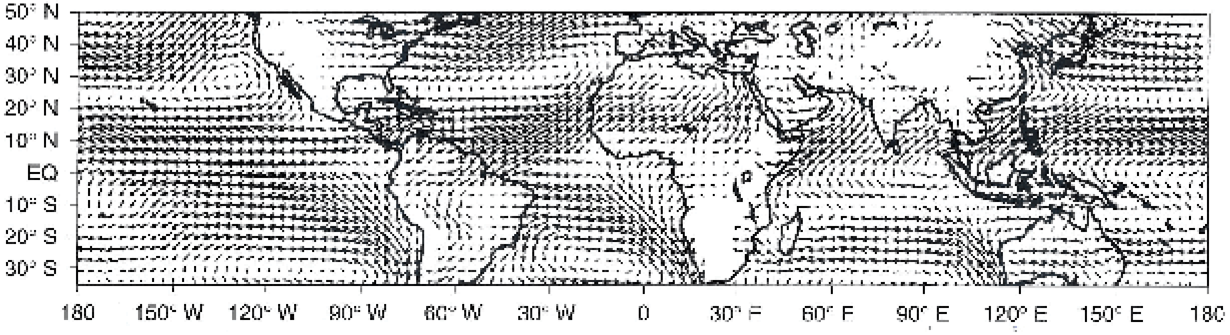
$$2\Omega\sin\varphi\overline{u}=-\frac{1}{a}\frac{\partial\overline{\Phi}}{\partial\varphi}$$

c) Trade winds and monsoonal flows

Observation of the stationary low level flow in the tropics in Jan-Feb



January February mean precipitations (GPCP data)



Winds at 925hPa, from ECMWF re-analysis

Over the Ocean the precipitations are Concentrated over the ITCZ, and SPCZ

Over land, precipitations are large over the Amazonian basin, southern subtropical Africa (monsoon regions are essentially in the SH in JF)

Mixture of lands and oceans make the maritime continent a zone of intense Convection.

Low level winds are blowing eastward at low levels (trade winds).

And also toward the monsoonal regions

Figures from J. Slingo (EAS, 2003)

c) Trade winds and monsoonal flows

Toy model 1b: linear model for the trade winds, axisymetric troposphere bounded by two rigid lids at the ground and at the tropopause (Boussinesq+Hydrostatic model):

$$\frac{\partial \bar{u}}{\partial t}-2\Omega\sin\varphi\bar{v}=-\alpha\bar{u}$$

$$\frac{\partial \bar{v}}{\partial t}+2\Omega\sin\varphi\bar{u}=-\frac{1}{a}\frac{\partial\bar{\Phi}}{\partial\varphi}-\alpha\bar{v}$$

$$\frac{\partial\bar{\Phi}}{\partial z}=\bar{b}$$

$$\frac{\partial\bar{b}}{\partial t}+N^2\bar{w}=\bar{q}-\alpha\bar{b}$$

$$\frac{1}{a\cos\varphi}\left(\frac{\partial\cos\varphi\bar{v}}{\partial\varphi}\right)+\frac{\partial\bar{w}}{\partial z}=0$$

Imposed vertical structure :

$$(\bar{q},\bar{w},\bar{b})=(Q,W,B)\sin\frac{\pi z}{D}$$

$$(\bar{u},\bar{v},\bar{\Phi})=(U,V,gH)\cos\frac{\pi z}{D}$$

A shallow water linear system
often used in tropical meteorology

$$\frac{\partial U}{\partial t}-2\Omega\sin\varphi V=-\alpha U$$

$$\frac{\partial V}{\partial t}+2\Omega\sin\varphi U=-\frac{1}{a}\frac{\partial H}{\partial\varphi}-\alpha V$$

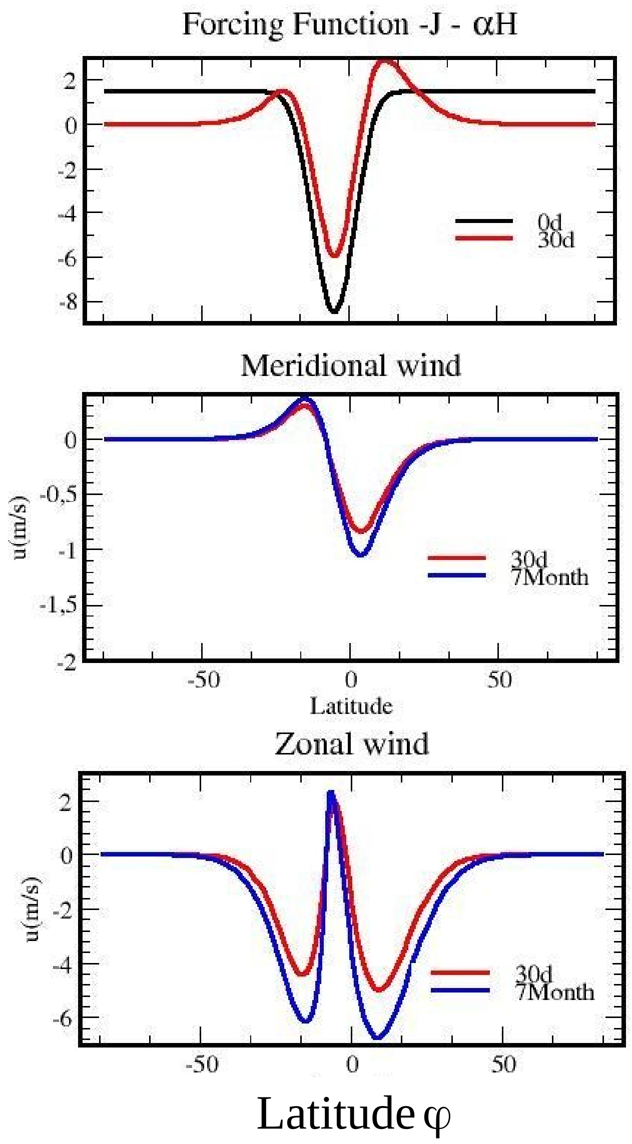
$$\frac{\partial H}{\partial t}+\frac{h}{a\cos\varphi}\frac{\partial V\cos\varphi}{\partial\varphi}=-J-\alpha H$$

$$J=\frac{DQ}{\pi g}$$

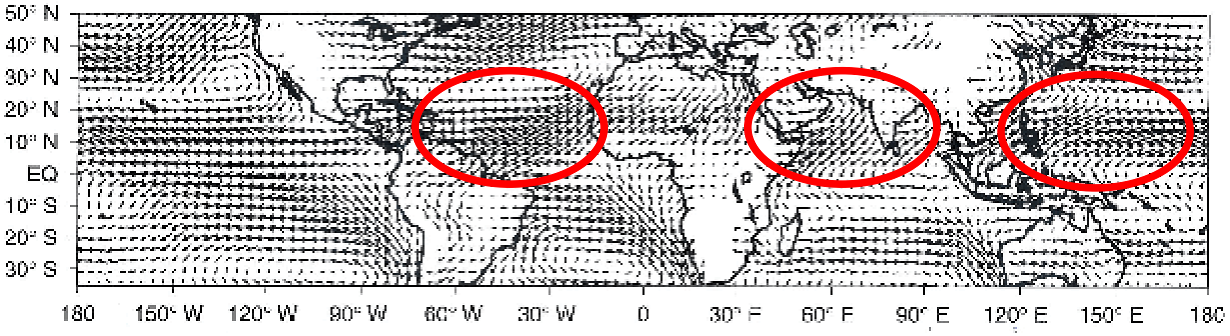
Equivalent depth: $h=\frac{N^2D^2}{\pi^2 g}\approx 100\mathrm{m}$

c) Trade winds and monsoonal flows

Toy model 1b with Forcing centered near the Equator, but in the SH (NH winter case)

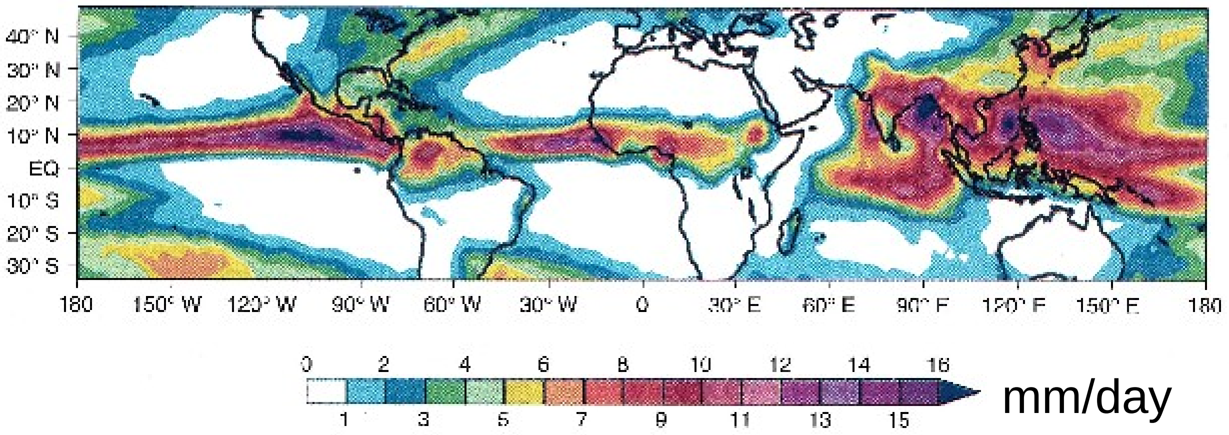


The winds converge toward the ITCZ with a substantial meridional component



c) Trade winds and monsoonal flows

Observation of the stationary low level flow in the tropics in Jul-August (favorable to monsoons in the NH subtropics)

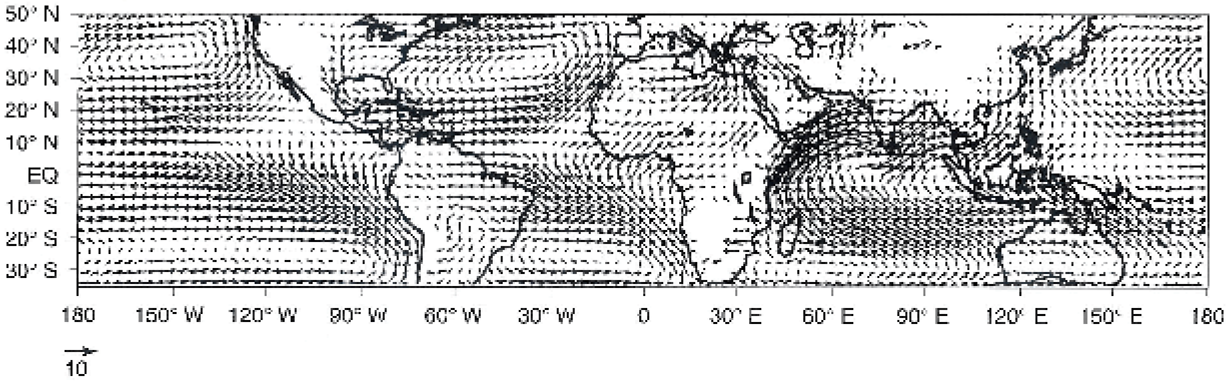


July-August mean precipitations (GPCP data)

Over the Ocean the precipitations are Concentrated over the ITCZ, and SPCZ.

But there is a shift toward the NH subtropics, in the bay of Bengal and all around India.

Note also the African Monsoon.



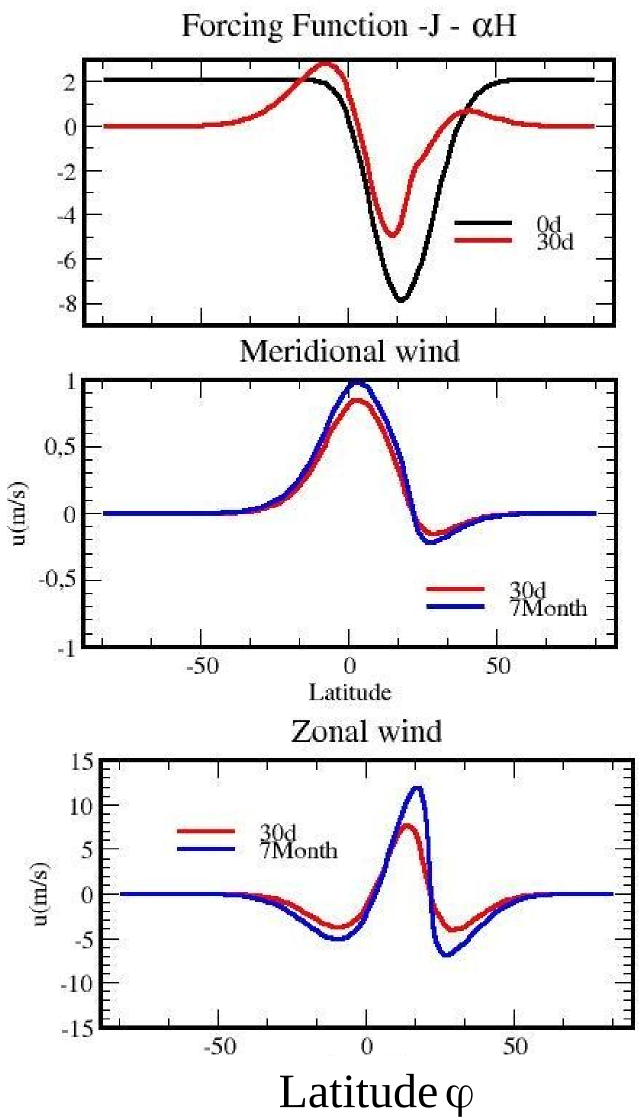
July-August Winds at 925hPa, from ECMWF re-analysis

The low level winds that feeds the Indian Monsoon in terms of moisture veers from being North-Eastward in the SH to North Westward in the SH

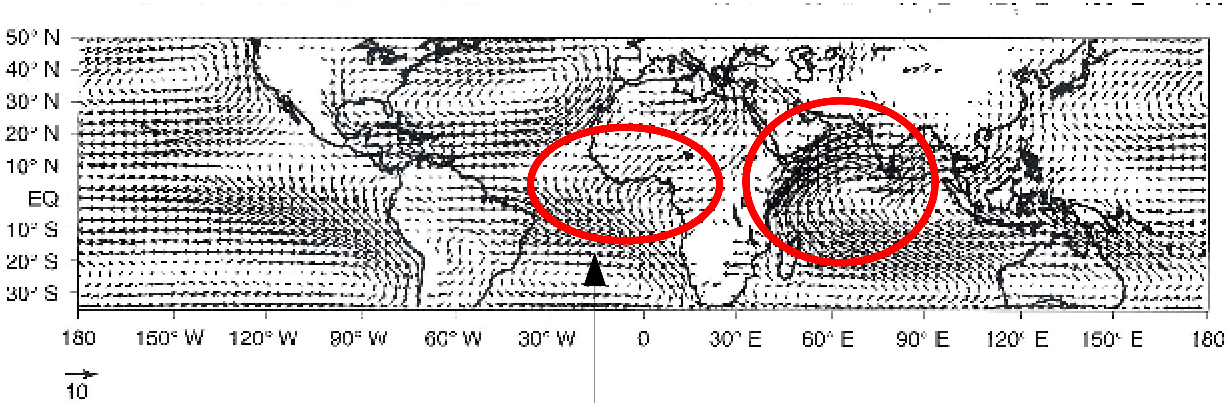
Figures from J. Slingo (EAS, 2003)

c) Trade winds and monsoonal flows

Toy model 1b with Forcing centered in the NH Subtropics, e.g. Over India and the Tibetan plateau (NH Summer case)



The monsoonal flow changes direction as it passes the equator and bring a large amount of moisture in India



African Monsoon

c) Trade winds and monsoonal flows

Why there are monsoons?

The temperature T of land warms more rapidly when summers arrive than the ocean temperature

$$\frac{dT}{dt}=-\frac{1}{\rho C_p}\frac{dF}{dz}$$

Although the Heat Capacity (C_p) and density (ρ) of soils are larger than those of water, the depth over which the heat flux *F* is distributed Δz is much thinner for lands (no turbulent flow to carry it downward, and very small thermal conductivity). For ocean, $\Delta z=60m$, just taking Into account the penetration of light

$$\frac{dT}{dt}=-\frac{1}{\rho C_p}\frac{F_{z=0}}{\Delta z}$$

H_s: sensible heat flux
H_e: evaporative heat flux
I_{net}: net radiation at the surface

$$F_{z=0}=I_{net}-H_s-H_e, \qquad I_{net}=S(1-a)+\epsilon\sigma T_g^4-\sigma T_a^4$$

σ: land emissivity
S: Solar constant
a: albedo

Before monsoon onset, lands warm, see here India and Africa in May-June: