Dynamic Meteorology

(WAPE: General Circulation of the Atmosphere and Variability)

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3) Meridional circulations and the role of the Eddies

a) Tropospheric Hadley and Ferrel cells, Eddy-driven jets

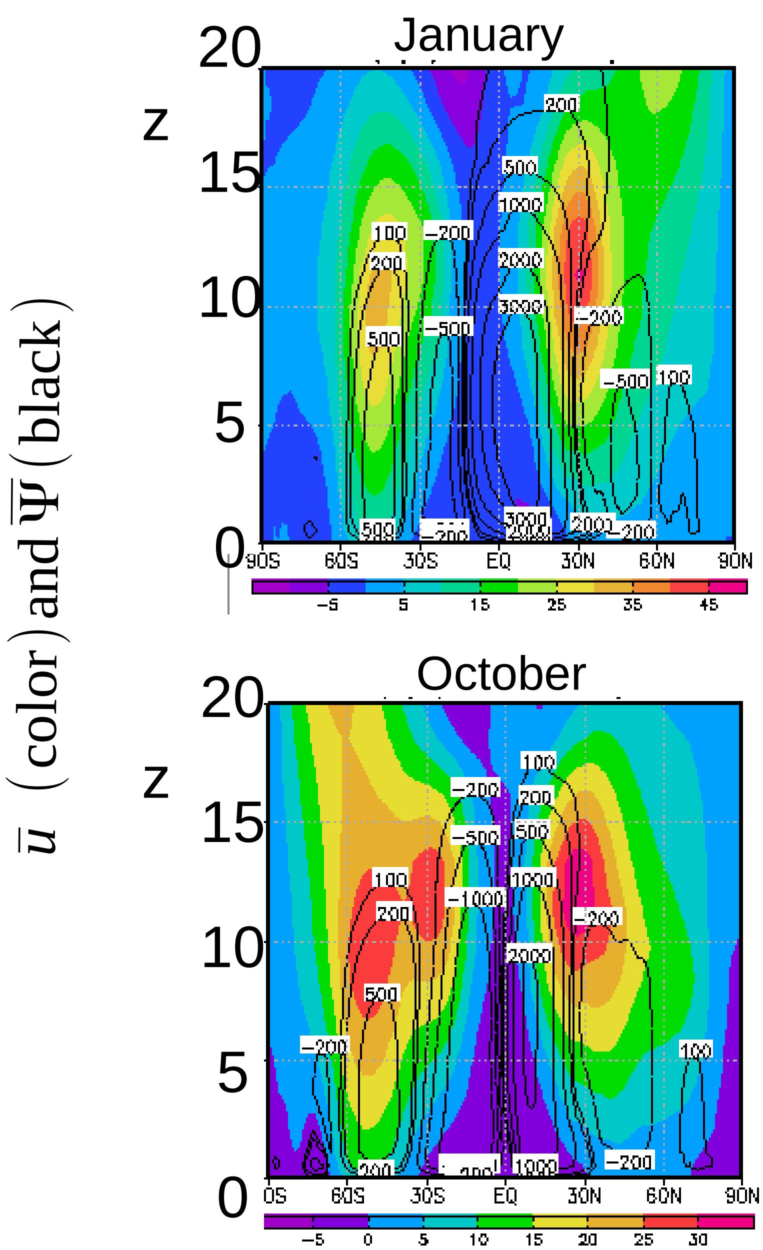
b) “Eulerian” and “transformed Eulerian” mean formalisms
Residual circulation, Eliasen Palm fluxes

c) Middle atmosphere Brewer-Dobson circulation
Transformed Eulerian mean streamfunction

a) Tropospheric Hadley and Ferrel Cells, Eddy-driven jets

Zonal mean meridional circulations in the troposphere visualized with a streamfunction

ERAI 1981-1985



$$\overline{v} = -\frac{1}{\rho_0 \cos\phi}\frac{\partial \overline{\Psi}}{\partial z} \qquad \overline{w} = +\frac{1}{\rho_0 a \cos\phi}\frac{\partial \overline{\Psi}}{\partial \phi}$$

- As expected from the shallow water model, the winds are positive at the subtropical end of the Hadley cells
- In the midlatitude regions, there seems to be a Cells which upper branch points toward the tropics, these are the Ferrel Cell
- Even poleward, there seem to be polar cells with upper branch toward the poles
- During Equinox, the Hadley and Ferrel cells are almost symmetric
- Note in the SH the presence of a secondary jet in the midlatitudes.
- Since this jet is not located at the exit of the Hadley cell, this secondary jet is “Eddy driven”.

a) <u>Tropospheric Hadley and Ferrel Cells, Eddy-driven jets</u>

Wave-mean flow interaction equations in the **Eulerian mean formalism**

Zonal mean: $\bar{u}(\varphi,z,t)=\frac{1}{2\pi}\int_0^{2\pi} u\, d\lambda$ Disturbance: $u'(\lambda,\varphi,z,t)=u-\bar{u}$

Angular momentum budget :

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

Rossby waves

Gravity waves in Midlatitudes

Equatorial waves in the tropics

Geostrophic approximation: $f\bar{u}\approx-\frac{1}{a}\bar{\Phi}_\phi$ (midlatitudes only!)

Hydrostatic approximation : $\bar{\Phi}_z=\frac{R\bar{T}}{H}$

Zonal mean meridional circulation satisfies mass conservation : $\frac{1}{a\cos\phi}(\bar{v}\cos\phi)_\phi+\frac{1}{\rho_0}(\rho_0\bar{w})_z=0$

Thermodynamics :

$$\bar{\theta}_t+a^{-1}\bar{v}\bar{\theta}_\phi+\bar{w}\bar{\theta}_z=\bar{Q}-(a\cos\phi)^{-1}(\overline{v'\theta'}\cos\phi)_\phi-\frac{1}{\rho_0}(\rho_0\overline{w'\theta'})_z$$

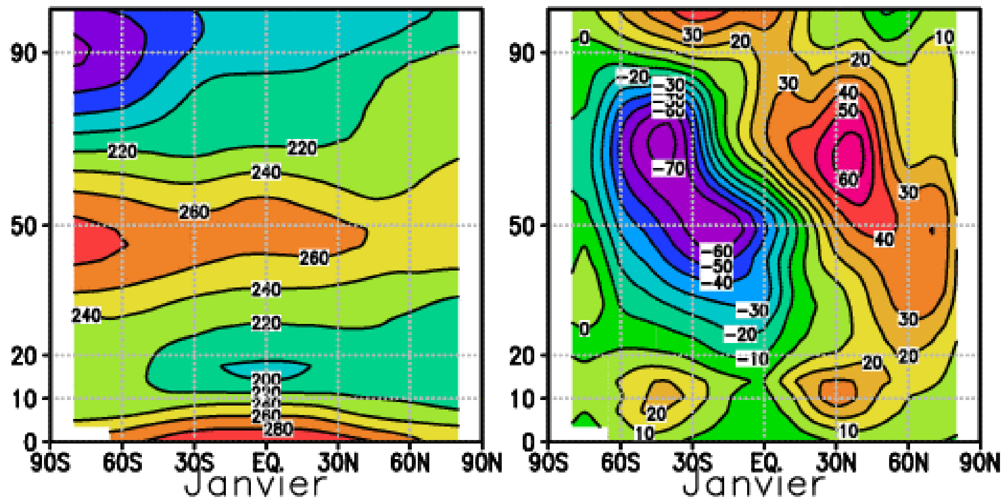
Baroclinic instabilities (troposphere)

Rossby waves (upper troposphere middle atmosphere)

a) Tropospheric Hadley and Ferrel Cells, Eddy-driven jets

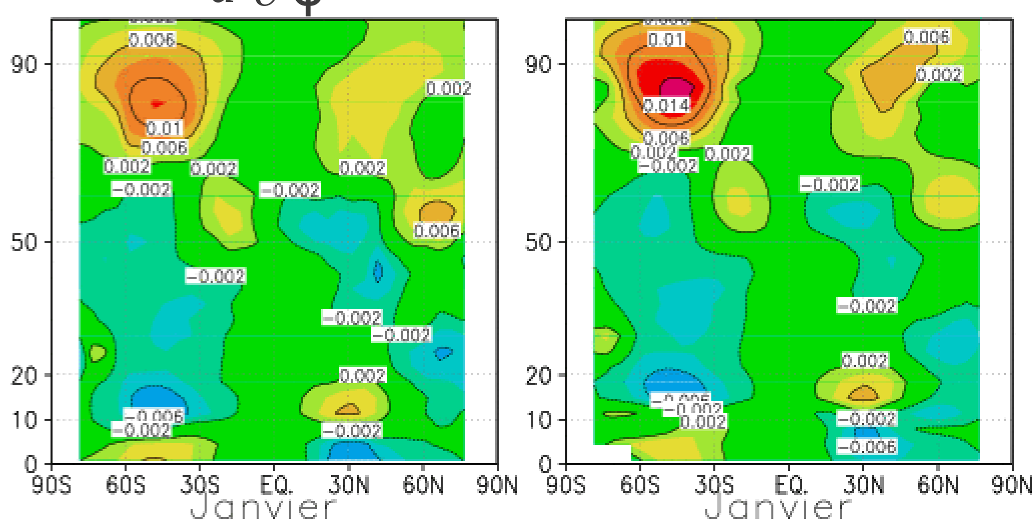
Thermal wind balance between the zonal mean zonal wind and the zonal mean temperature

Temperature $\overline{T}$ (K) Zonal wind $\overline{u}$ (m/s)



Thermal wind balance:

$$\frac{1}{a}\frac{\partial T}{\partial \phi} \qquad -\frac{fH}{R}\frac{\partial \overline{u}}{\partial z}$$



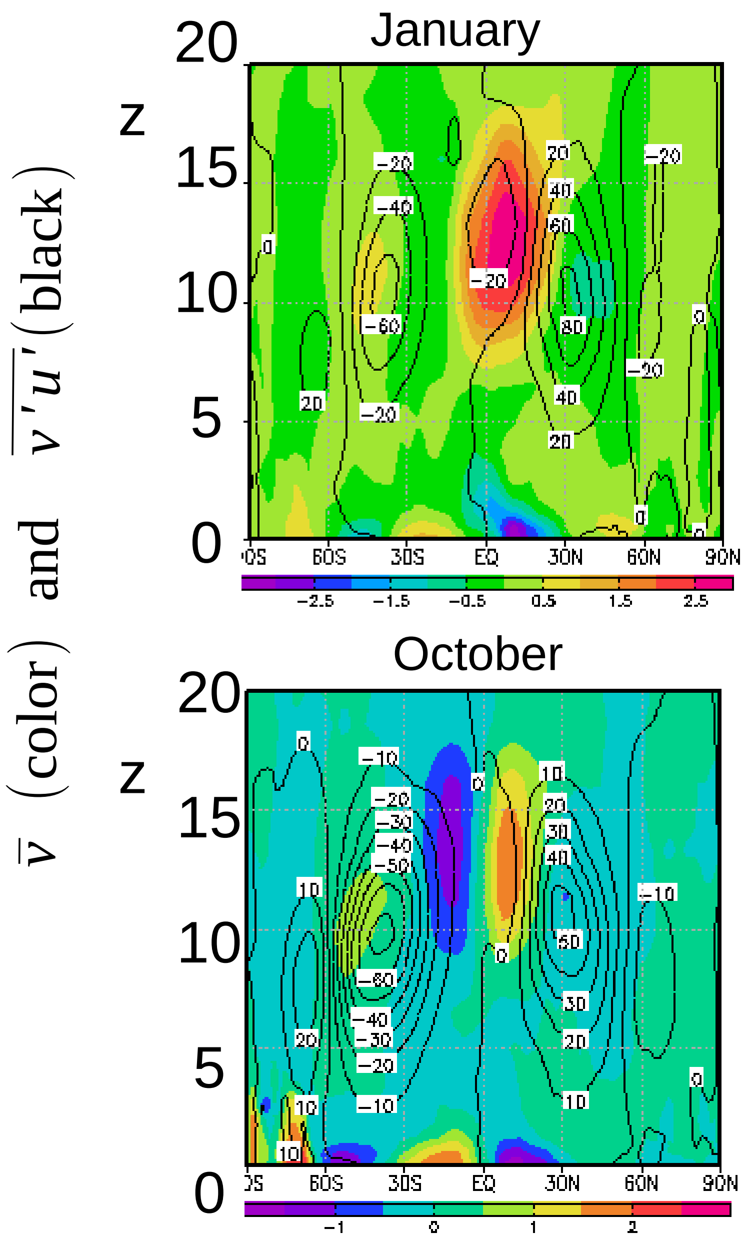
CIRA January climatologies

- Can we really separate the thermal and mechanical forcings?
- The thermal wind balance translates the strong links that exist in the midlatitude between dynamics and thermodynamics
- In January, the increase of T with latitude in the SH mesosphere permits the Easterly jet to close at the mesopause.
- Imagine now that one accelerates the jet at the mesopause to close the easterlies, the Temperature gradient in latitude should be positive below to satisfy the thermal wind balance.
- We will see that these mechanical forcings by the waves are extremely important, and drive the atmosphere out of radiative equilibrium.

a) <u>Tropospheric Hadley and Ferrel Cells, Eddy-driven jets</u>

Momentum fluxes

ERAI 1981-1985



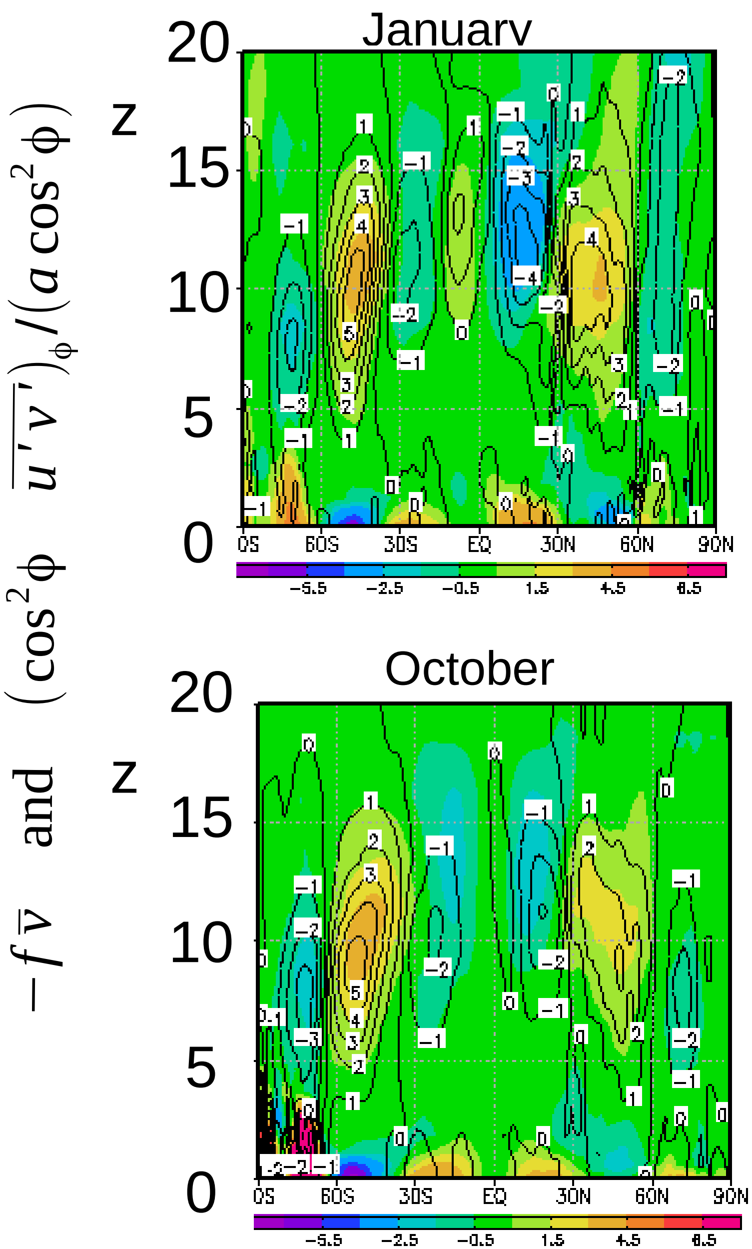
- The amplitude of the meridional wind is much less than that of the zonal wind (meridional circulations are slow)
- Note the flux of momentum due to the Eddies from the subtropics to the midlatitude in the upper tropopshere

a) Tropospheric Hadley and Ferrel Cells, Eddy-driven jets

Momentum fluxes divergence in good part equilibarated by the Coriolis torque:

ERAI 1981-1985

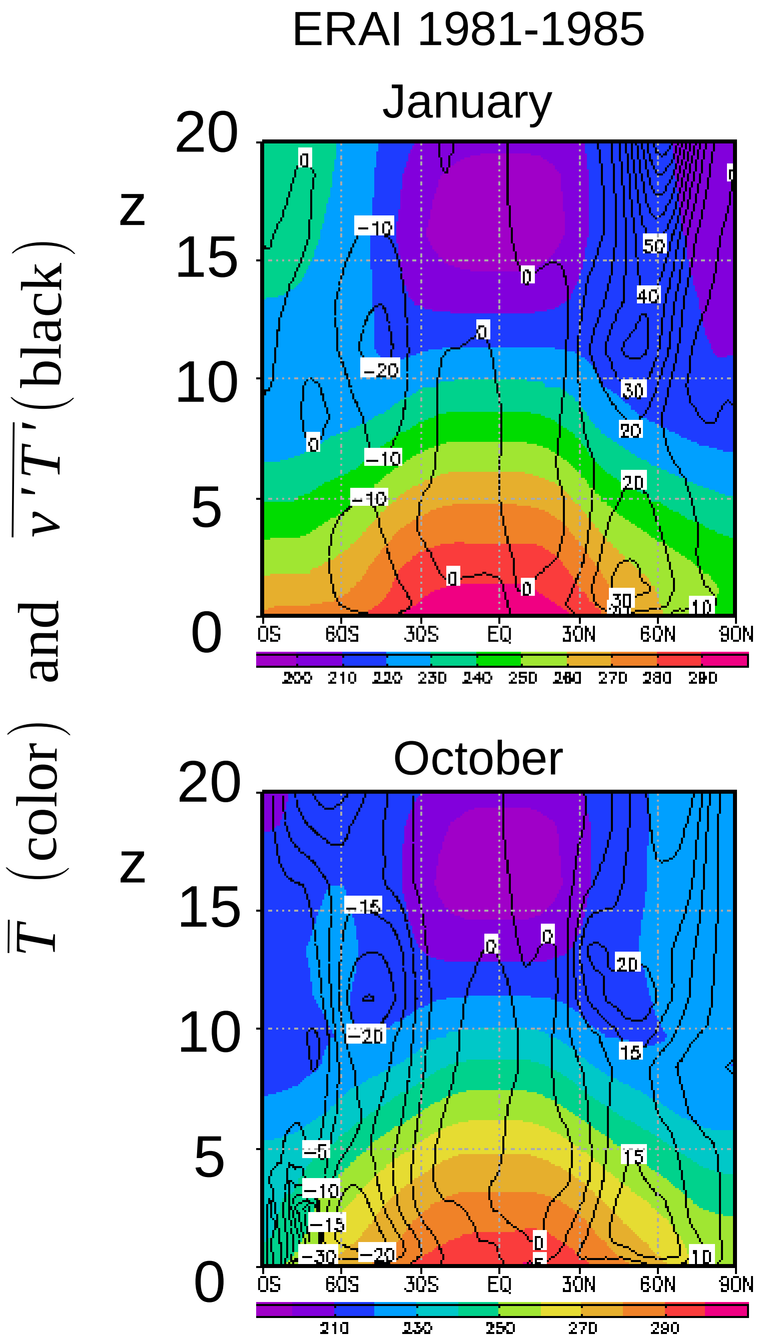
$$-\left(a\cos^2\phi\right)^{-1}\left(\overline{u'v'}\cos^2\phi\right)_\phi \approx -f\,\overline{v}$$



- The Hadley celll seems in good part Eady-driven.
- The Ferrel Cell is entirely related to the presence of the Eddies

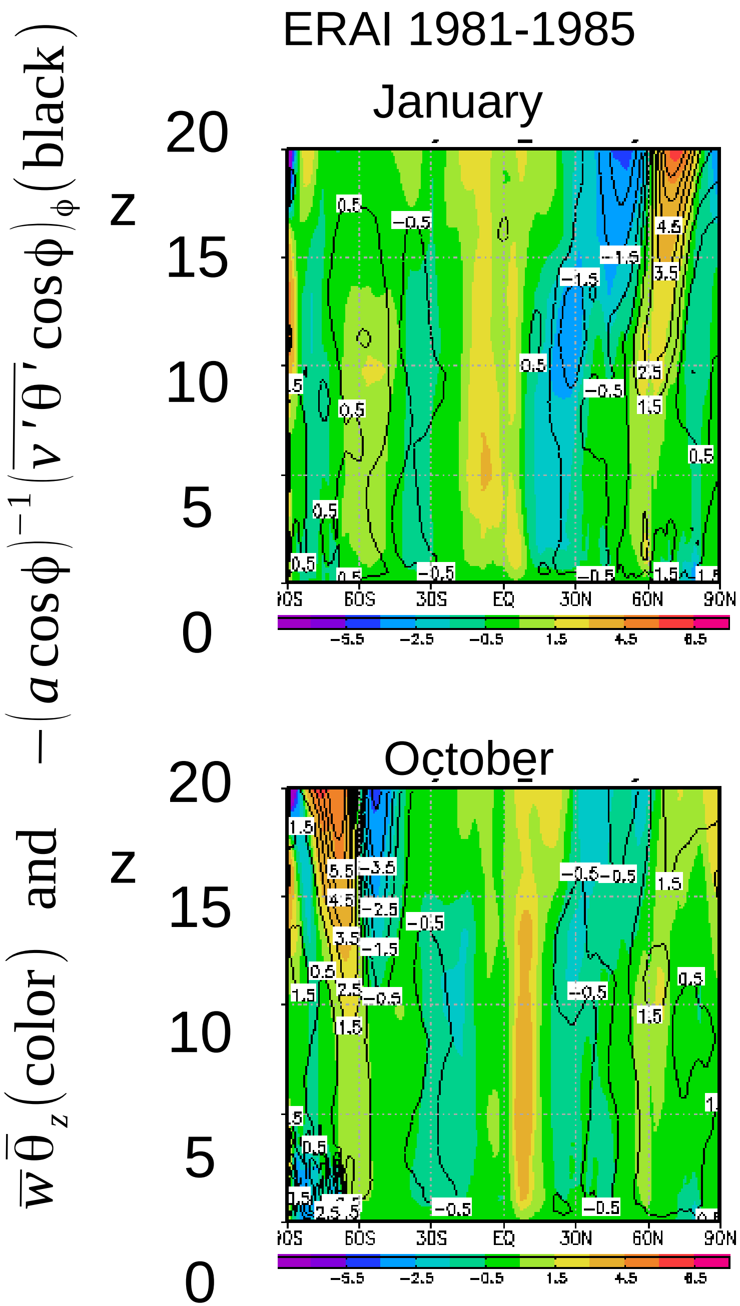
a) Tropospheric Hadley and Ferrel Cells, Eddy-driven jets

Thermal fluxes



- Substantial poleward flux of temperature at all altitudes in the midlatitudes
- Near the surface this warm the polar regions through diabatic effects
- But there are also substantial fluxs in the upper troposphere and lower stratosphere.
- In these regions the diabatic effects are not strong

a) Tropospheric Hadley and Ferrel Cells, Eddy-driven jets



Thermal fluxes

- On vertical velocity one sees ascent in the tropics as expected for a diabatic forcing
- The vertical motion in the midlatitudes are in part equilibrated by the Eddy heat flux divergence
- The relation becomes very strong in the midlatitude stratosphere
- Such an an adiabiatic mechanism can not explain a real ascent (mechanical waves do not heat in the absence of diabatic effects).
- Only a residual circulation that include such a balance can be more directly linked to diabatic heatings
- The appearance of eddy flux terms on both momentum and thermodynamic equations, their near cancellation by mean flow process, and the thermal wind balance constraint render the separate diagnostics rather inefficient to describe the net effect of the waves on the mean flow !

c) "Transformed Eulerian" mean formalisms

Wave-mean flow interaction equations in the **Transformed Eulerian mean formalism**

Residual mean Circulation :

$$\overline{w}^* = \overline{w} + \frac{1}{a\cos\phi}\frac{\partial}{\partial\phi}\left(\cos\phi\,\frac{\overline{v'\theta'}}{\overline{\theta}_z}\right) \qquad \overline{v}^* = \overline{v} - \frac{1}{\rho_0}\frac{\partial}{\partial z}\left(\rho_0\,\frac{\overline{v'\theta'}}{\overline{\theta}_z}\right)$$

Thermodynamics :

$$\overline{\theta}_t + a^{-1}\overline{v}^*\overline{\theta}_\phi + \overline{w}^*\overline{\theta}_z = \overline{Q} \quad -\frac{1}{\rho_0}\left(\rho_0\overline{v'\theta'}\frac{\overline{\theta}_\phi}{a\overline{\theta}_z} + \rho_0\overline{w'\theta'}\right)_z$$

Very small, at least well above the boundary layer

Angular momentum budget :

$$\overline{u}_t + \left[(a\cos\varphi)^{-1}(\overline{u}\cos\varphi)_\varphi - f\right]\overline{v}^* + \overline{u}_z\overline{w}^* = \overline{X} \quad + (\rho_0 a\cos\varphi)^{-1}\vec{\nabla}\cdot\vec{F}$$

$$\left(\vec{\nabla}\cdot\vec{F} = \frac{1}{a\cos\phi}\frac{\partial\cos\phi\,\overline{F}^\phi}{\partial\phi} + \frac{\partial\overline{F}^z}{\partial z}\right)$$

Geostrophic approximation: $f\overline{u} \approx -\frac{1}{a}\overline{\Phi}_\phi$

Hydrostatic approximation : $\overline{\Phi}_z = \frac{R\overline{T}}{H}$

Transformed Eulerian mean meridional circulation satisfies mass conservation :

$$\frac{1}{a\cos\varphi}(\overline{v}^*\cos\varphi)_\varphi + \frac{1}{\rho_0}(\rho_0\overline{w}^*)_z = 0$$

Eliasen Palm Flux:

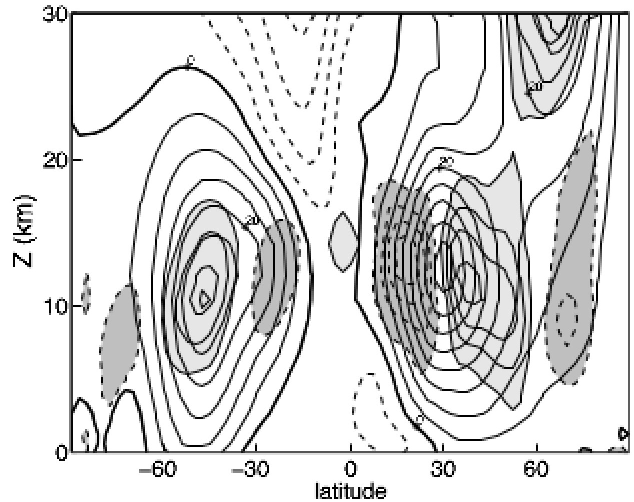
$$\overline{F}^\phi = \rho_0\,a\cos\phi\left(-\overline{v'u'} + \overline{v'\theta'}\frac{\overline{u}_z}{\overline{\theta}_z}\right) \qquad \overline{F}^z = \rho_0 a\cos\phi\left(\frac{\overline{v'\theta'}}{\overline{\theta}_z}\left(f - \frac{(\overline{u}\cos\phi)_\phi}{a\cos\phi}\right) - \overline{w'u'}\right)$$

b) Transformed Eulerian mean diagnostics

Forcing components of the Eady-driven jets in the troposphere

$$\frac{1}{a\cos\phi}\frac{\partial \cos\phi \overline{F^{\phi}}}{\partial \phi}$$

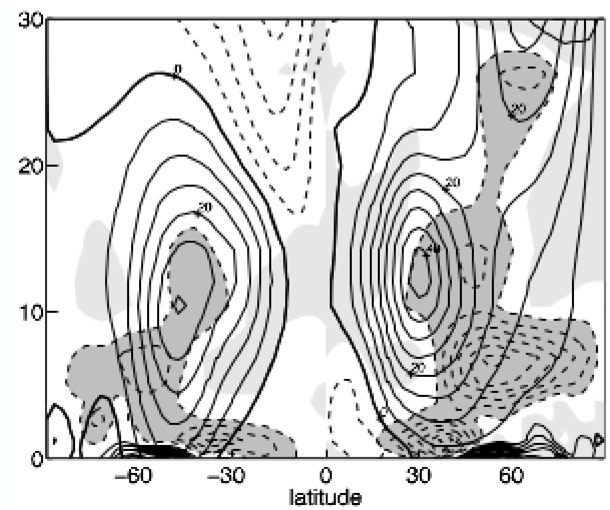
$\overline{F^{\phi}}$ essentially related to $\overline{v'u'}$



Decelerate the subtropical jet near the tropopause accelerates in the midlatitude (potentially yielding an upper level eddy driven jet)

$$\frac{\partial \overline{F^{z}}}{\partial z}$$

$\overline{F^{Z}}$ essentially related to $\overline{v'T'}$

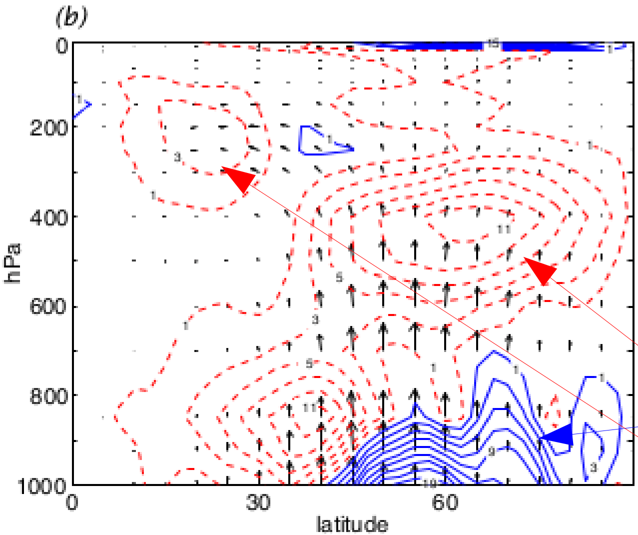


Decelerates in the midlatitude near the tropopause acceleration near the surface : « barotropisation » of the jet

Observed NH winter from ERA40 (Figs. 12.18, Vallis 2005)

b) Transformed Eulerian mean diagnostics

Application to the Eady-driven jets in the troposphere



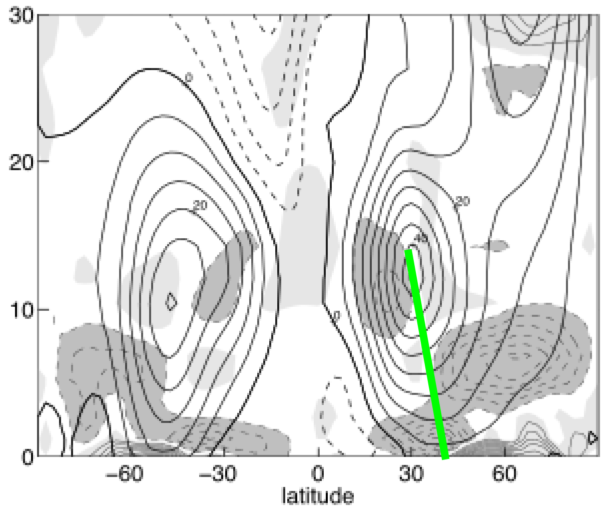
$\vec{\overline{F}}$ and $\dfrac{\vec{\nabla}\cdot\vec{\overline{F}}}{\rho_0 a\cos\phi}$

The dominant effect of the eddies is to make the subtropical jet more barotropic in the midlatitudes :

Dynamical acceleration at low levels In the NH midlatitudes,

compensated by

a decelaration near the tropopausse in the NH midlatitude and on the southern flank of the subtropical jet



$\overline{u}$ and $\dfrac{\vec{\nabla}\cdot\vec{\overline{F}}}{\rho_0 a\cos\phi}$

Tilt with altitude of the maxima in the westerlies

Observed NH winter from ERA40 (Figs. 12.16, 12.17, Vallis 2005)

c) **The middle atmosphere Brewer-Dobson circulation**

Rather than starting from angular momentum, we can look more directly at trace species: H_20, 0_3

Satellite measurements in the stratosphere

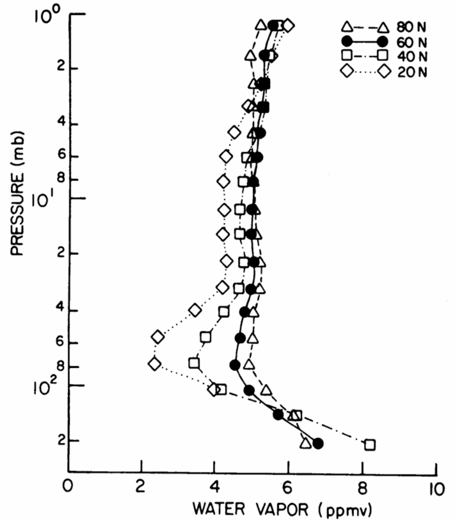
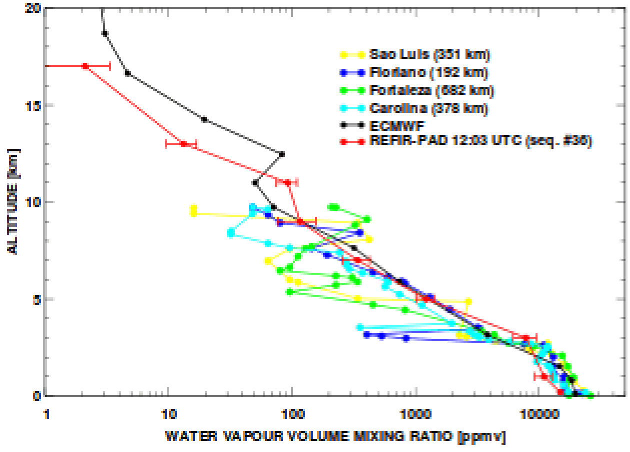


Fig. 1.5. Vertical profiles of water vapor mixing ratio at several latitudes measured by the LIMS instrument on the *Nimbus* 7 satellite for May 1–26, 1979. [From Remsberg *et al.* (1984b). American Meteorological Society.]

Balloon measurements in the troposphere:



- From lecture 1
- Water vapor (H_2O) presents Rapid decay with altitude, very weak values (almost uniform) in the stratosphere.
- Strong greenhouse impact in the troposphere.
- And minimum value at the equatorial tropopause

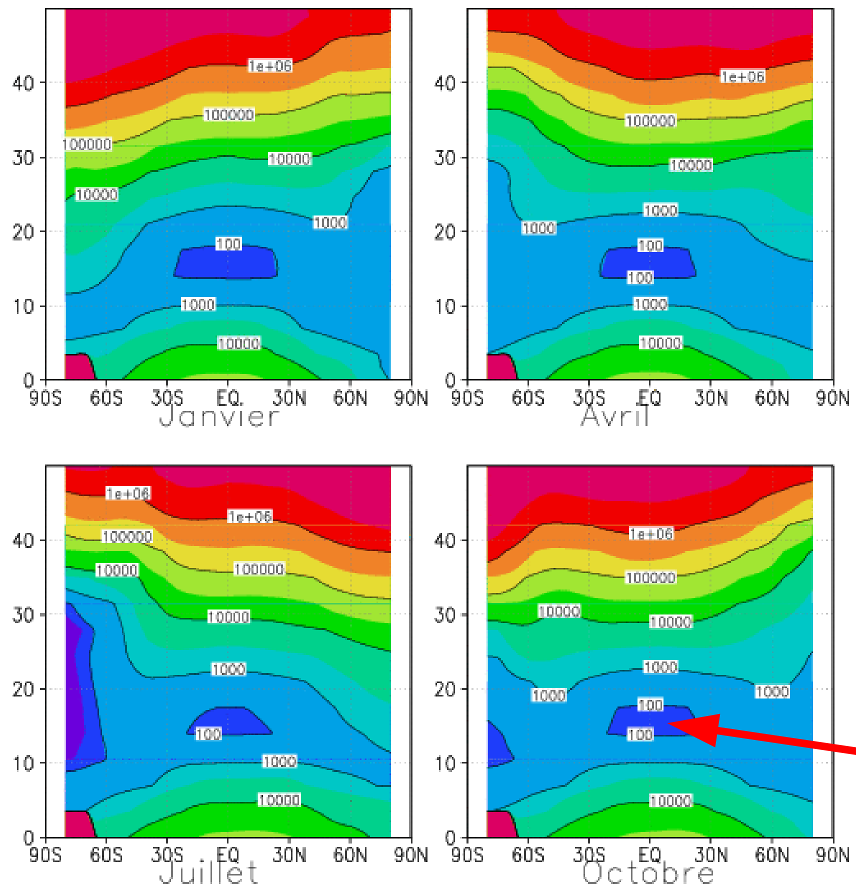
c) The middle atmosphere Brewer-Dobson circulation

Evidence on trace species: H_2O

Stratosphere is very poor in water vapor (few ppmv).

not related to the saturated values of the water vapor mixing ratios

Données CIRA, $\mathcal{V}_{v\,\mathrm{sat}}$



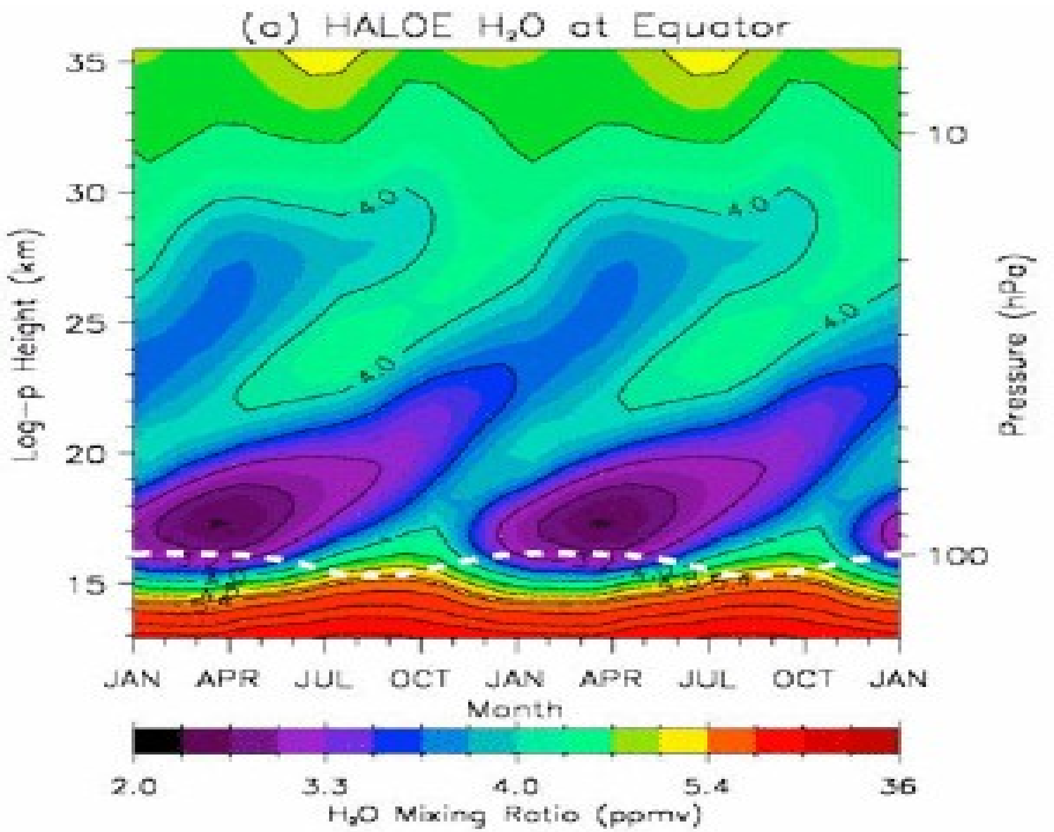
- Number of molecules per unit volume: n_v
- Volume mixing ratio (ppmv):
$$\mathcal{V}_v = n_v / n_A = p_v / p_A$$
- In the troposphere, $\mathcal{V}_v$ is around 10000ppmv, it is near the max value:
$$\mathcal{V}_{v\,\mathrm{sat}} = p_{v\,\mathrm{sat}} / p_A \quad \text{avec:}$$
$$p_{v\,\mathrm{sat}} \approx 1.E^{+5} \exp\left(13.7 - \frac{5120}{T}\right)$$
- In stratosphere: 2-6 ppmv
- The only region where such low values can be obtained durin all seasons is the equatorial tropopause
- Small scale processes are nevertheless still missing: 100ppmv is still too much

c) The middle atmosphere Brewer-Dobson circulation

Evidence on trace species: H_20

Satellites observations of H_20 (HALOE)

The « tape recorder »



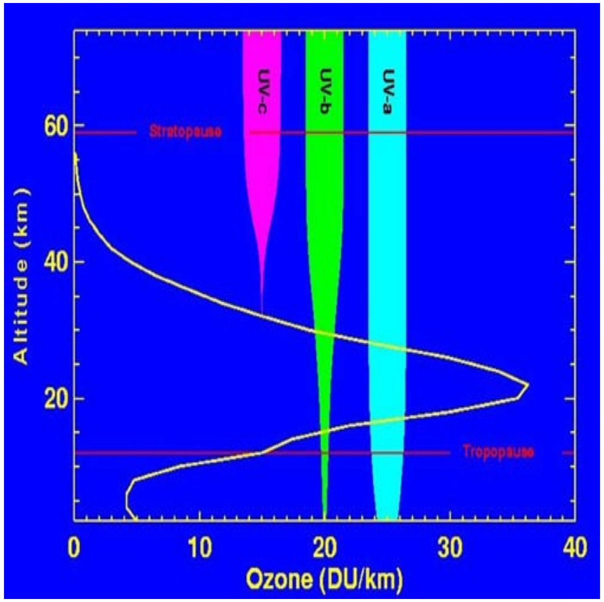
- Note the slow ascent of the tropical air, and which conserves quite well the value of $\mathcal{V}_v$ it acquires at the tropical tropopause.
- The values of $\mathcal{V}_v$ increases as the air rises via horizonbtal diffusion and because of the methane (CH_4) oxydation
- Park et al. JGR 2004

c) The middle atmosphere Brewer-Dobson circulation

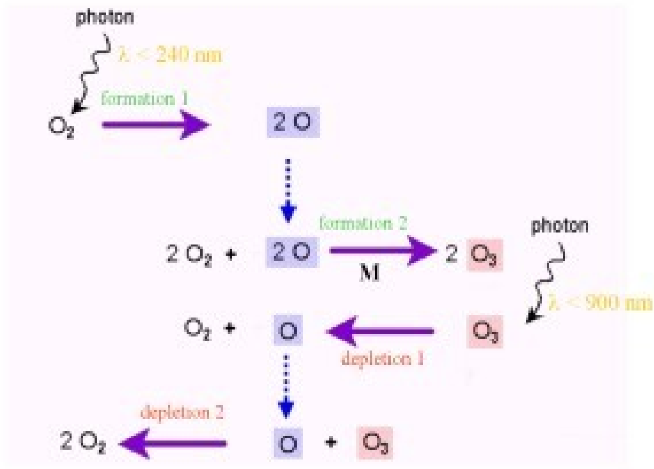
Evidence on trace species: O3

La production de l'Ozone (O_3)

Penetration Of the UV-a, UV-b, UV-c



Chapman cycle (1930):

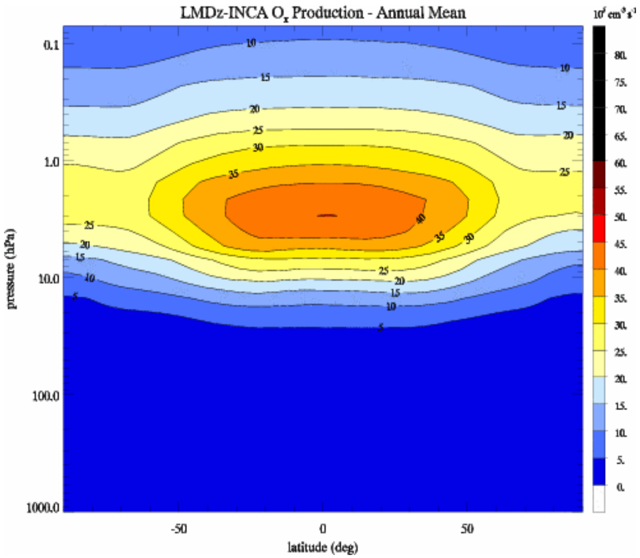


- **UV-c** are absorbed by l'$\mathbf{O_2}$ between 40km and 60km. This photolysis yields molecular oxygene free radicals **O**.
- **These free radicals combine rapidly with $\mathbf{O_2}$** to form Ozone $\mathbf{O_3}$.
- The Ozone absorbs the **UV-b** above 20km to give **O**
- In most cases **O** recombines back with avec $\mathbf{O_2}$ to give $\mathbf{O_3}$, there is chemical heating.
- The complete family **O+$\mathbf{O_3}$ ($\mathbf{O_x}$)** nevertheless has a relatively long lifetime in the low stratosphere.
- The Chapman cycle explains the $\mathbf{O_x}$ production and the heating by $\mathbf{O_3}$
- It does not explain well the $\mathbf{O_3}$ (Rôle of other free radicals than just O)

c) **The middle atmosphere Brewer-Dobson circulation**

Evidence on trace species: O_3

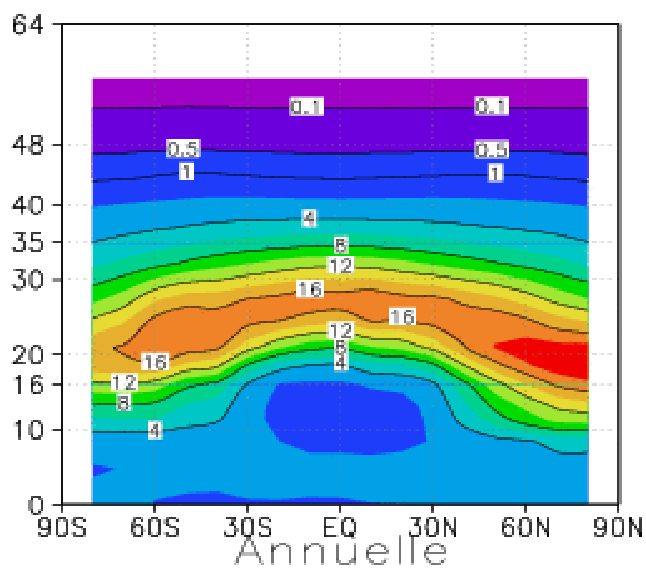
Annual production of Ox



The production of ozone (Ox)

Atmospheric photo-chemical model INCA,
operational in LMDz
(Thanks to D. Hauglustaine)

Ozone concentration (in DU/km)

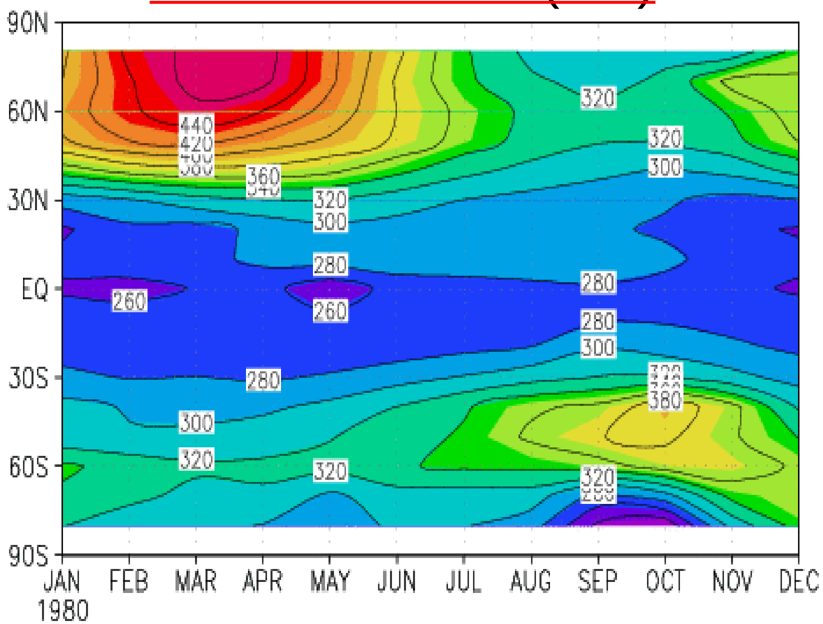


The ozone is not located where it is produced

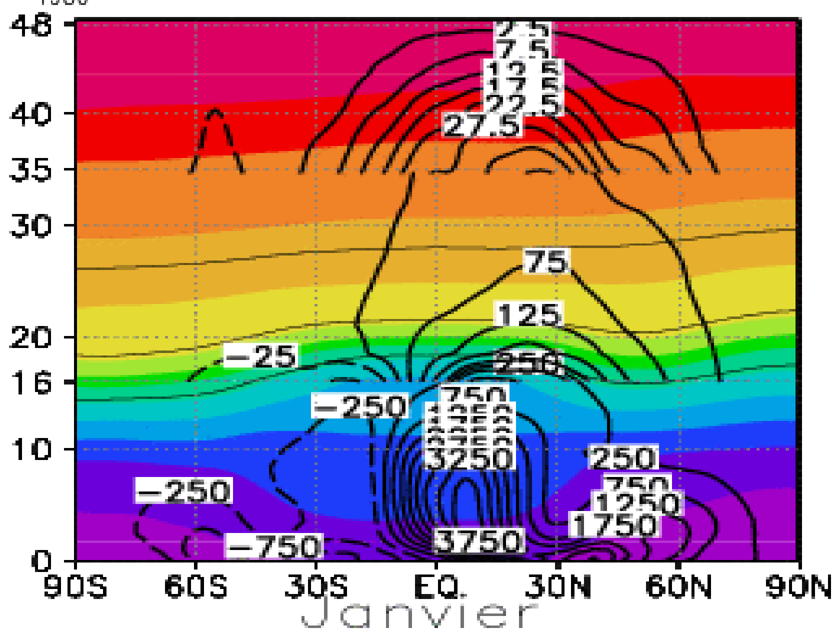
c) The middle atmosphere Brewer-Dobson circulation

Evidence on trace species: O_3

Annual cycle of the Ozone column in Dobson Units (DU)



- The air filled with newly produced O3 travels from the zones of Ozone production (near and above the tropical stratopause) and transported toward the mid and high latitudes
- Note in April there is a strong concentration of O_3 at the North Pole! The Ozone has cumulated during winter
- Note also the O_3 deficit near the south pole in October: this is the Ozone hole, relted to weak horizontal mixing and to a Brewer Dobson in the southern hemisphere winter not reaching the South pole



$$\frac{\partial \Psi^*}{\partial z} = -\rho_0 \cos \phi v^*$$

- "pseudo" Lagrangian streamfunction Ψ^*:

  it transports the air from the upper stratosphere to the lower stratosphere and polar regions

c) **The middle atmosphere Brewer-Dobson circulation**

Limit of the Eulerian formalism

The Eulerian formalism can not explains the transport of 03 from the equatorial stratopause to the lower stratosphere in the polar regions: illustration here using streamfunctions.

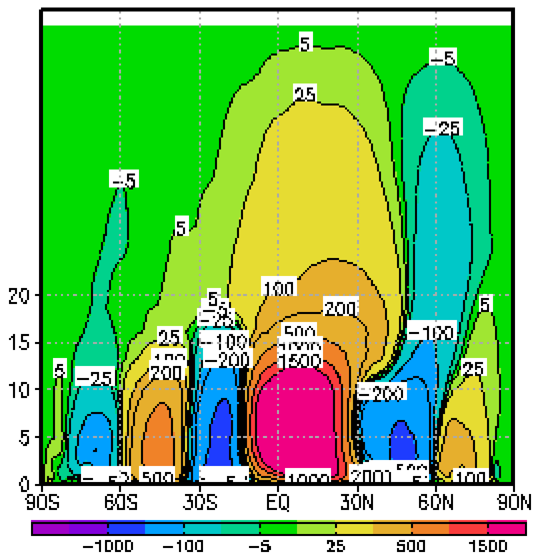
One needs to take into account Lagrangian averaging, or here its approximation, the transformed Eulerian mean

ERA40, January 1981–1985

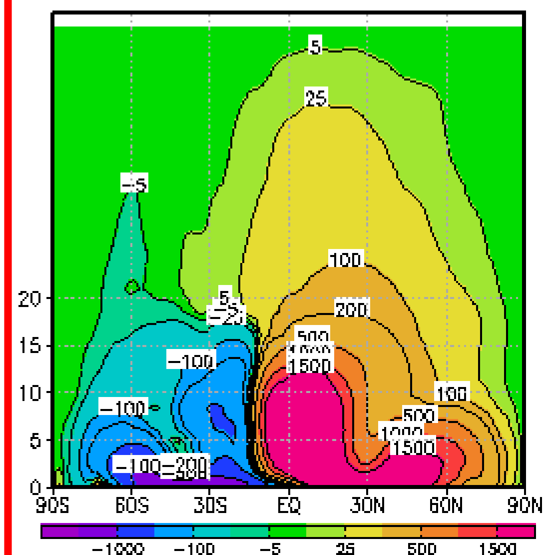
Eulerian mean streamfunction

$$\frac{\partial \overline{\Psi}}{\partial z} = -\rho_0 \cos\phi\, \overline{v}$$

$$\frac{1}{a\cos\phi}\frac{\partial \overline{\Psi}}{\partial \phi} = +\rho_0\, \overline{w}$$



Transformed eulerian mean streamfunction



$$\frac{\partial \overline{\Psi^*}}{\partial z} = -\rho_0 \cos\phi\, \overline{v^*}$$

$$\frac{1}{a\cos\phi}\frac{\partial \overline{\Psi^*}}{\partial \phi} = +\rho_0\, \overline{w^*}$$

Residual mean circulation (definition):

$$\overline{v^*} = \overline{v} - \frac{1}{\rho_0}\frac{\partial}{\partial z}\left(\rho_0 \frac{\overline{v'\theta'}}{\overline{\theta}_z}\right)$$

$$\overline{w^*} = \overline{w} + \frac{1}{a\cos\phi}\frac{\partial}{\partial \phi}\left(\cos\phi \frac{\overline{v'\theta'}}{\overline{\theta}_z}\right)$$

c) The middle atmosphere Brewer-Dobson circulation

The middle atmosphere is not in radiative equilibrium

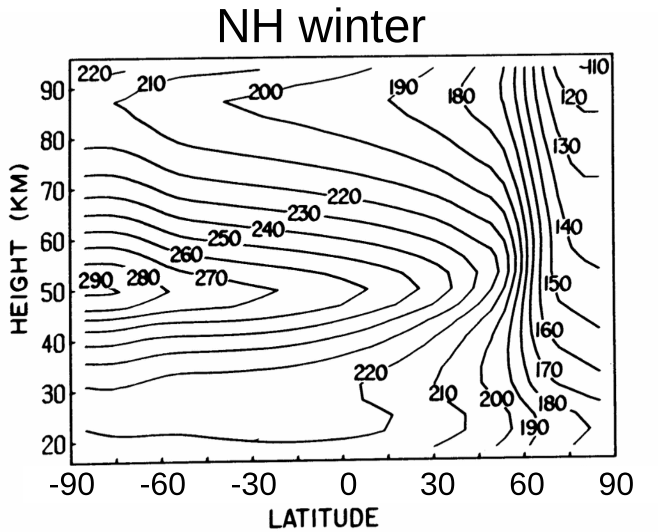


Fig. 2.34. Radiative equilibrium temperature distribution for northern (left) summer solstice. [From Wehrbein and Leovy (1982), with permission.]

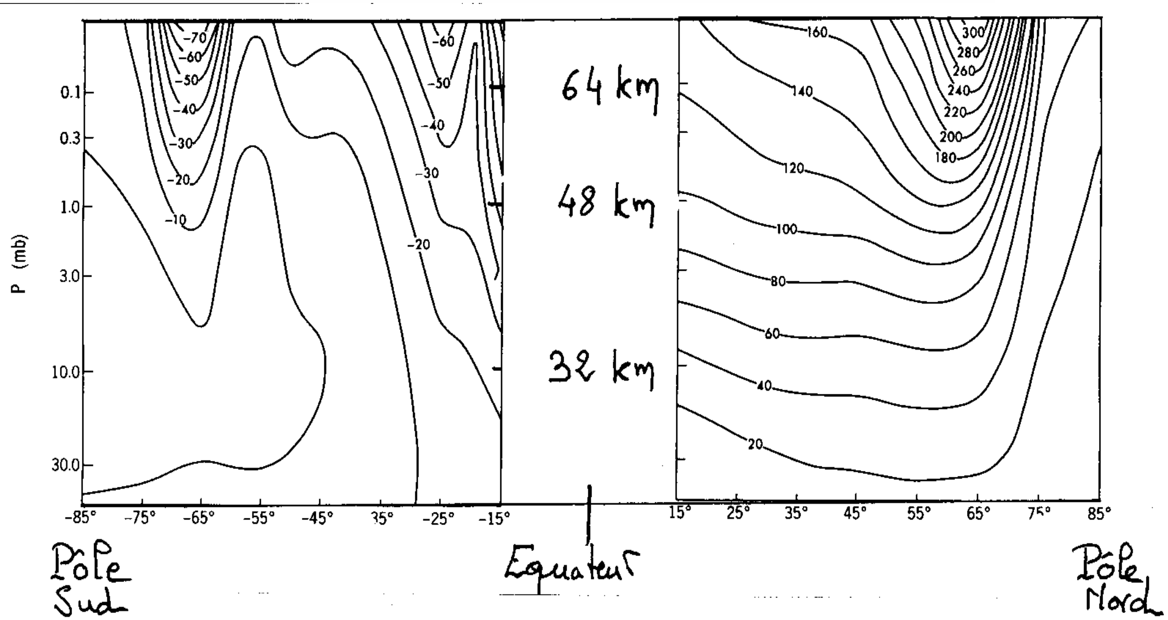


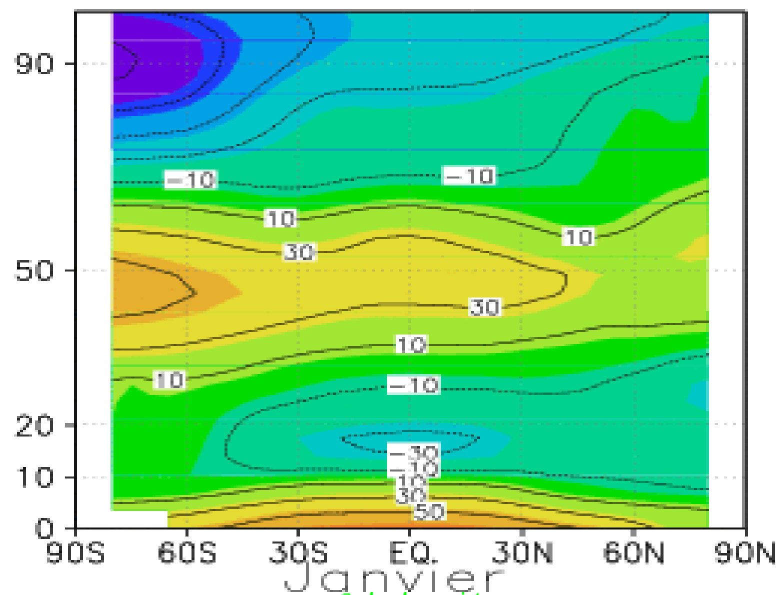
Fig. 7.1. Zonal gradient wind u_{gr} that is in thermal-wind balance with the temperature field T_r of Fig. 1.2 and equals the observed climatological zonal wind at 100 mb. (a) Northern Hemisphere (winter), (b) Southern Hemisphere (summer). (Courtesy of Dr. S. B. Fels.)

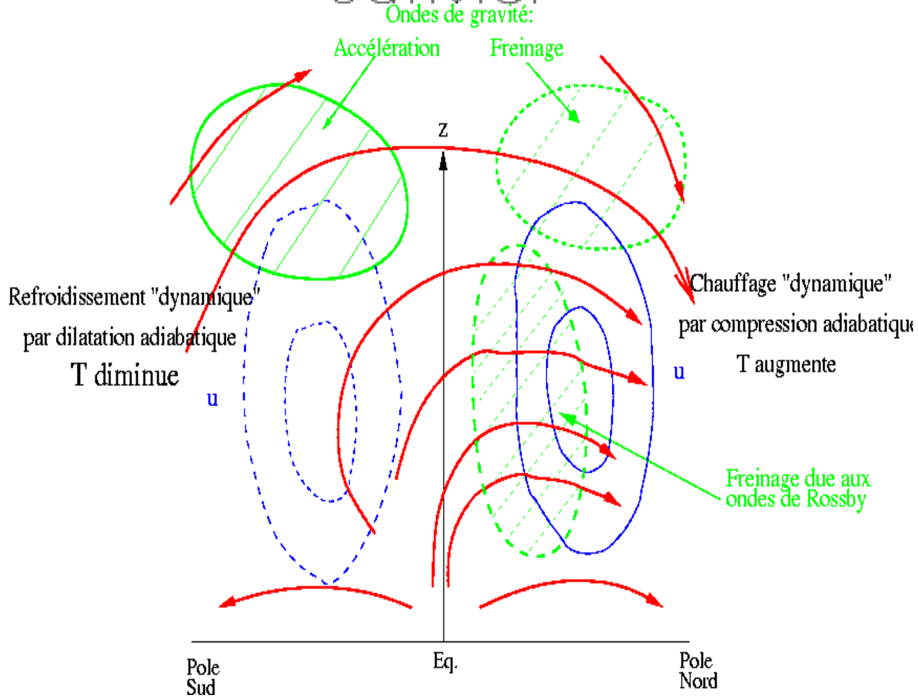
- Temperature and winds resulting from radiative equilibrium and thermal wind balance.
- In January, T decreases from summer pole to the winter pole as expected
- The winds are eastward in winter and westward in summer
- The winds are much to strong, and the jets do not close when approaching the mesopause

c) The middle atmosphere Brewer-Dobson circulation

The middle atmosphere is not in radiative equilibrium

January Temperature (CIRA)





- Temperature is far from the radiative equilibrium seen before:
- They are warmer in at the winter pole, colder at the summer pole.
- During solstices:
  - In the upper mesosphere (70-90km) T increases from the winter pole to the summer pole !!!
  - At the mesopause (90km) over the summer pole is the coldest region of the neutral atmosphere!

Interpretation via an essentially mechanically driven circulation: the Brewer Dobson circulation.