Specatmos

Spectroscopie et atmosphères : mesures et modèles

École thématique CNRS

Atmospheric Physics

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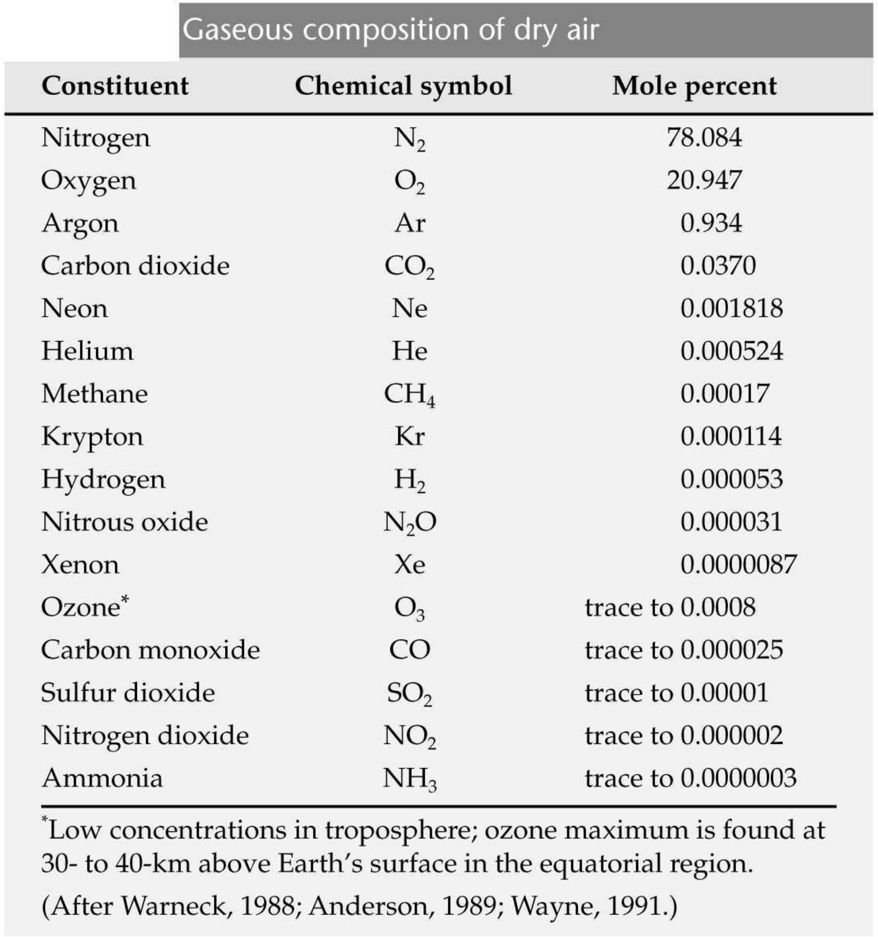
Earth‘s atmosphere and circulation

Atmospheric composition

Composition



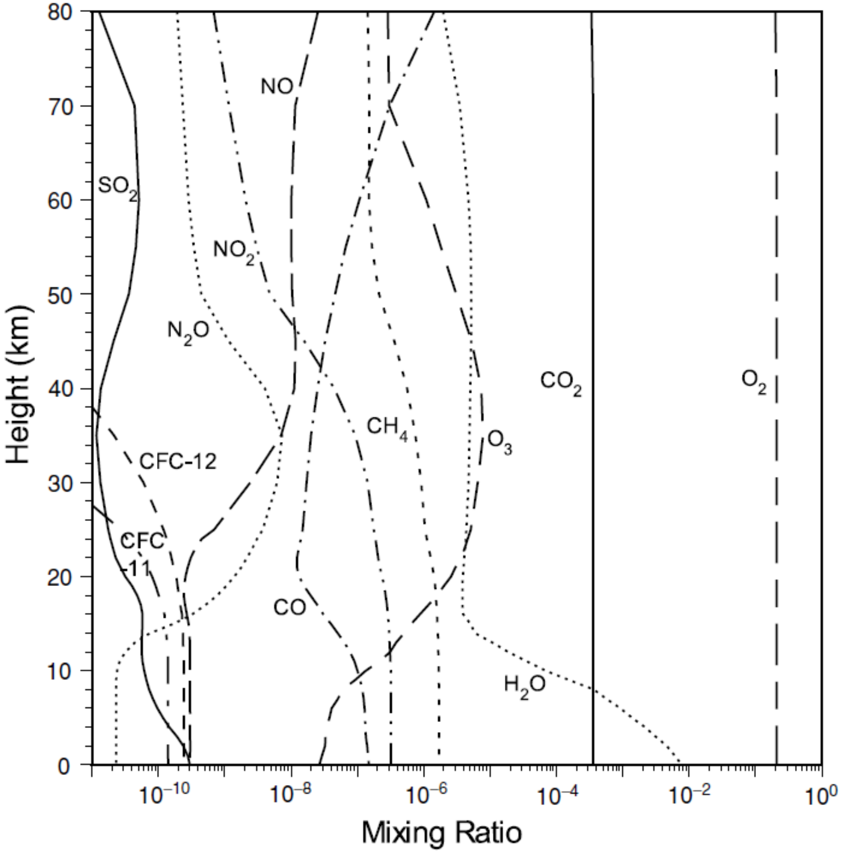
Gaseous composition of dry air

Constituent	Chemical symbol	Mole percent
Nitrogen	N_2	78.084
Oxygen	O_2	20.947
Argon	Ar	0.934
Carbon dioxide	CO_2	0.0370
Neon	Ne	0.001818
Helium	He	0.000524
Methane	CH_4	0.00017
Krypton	Kr	0.000114
Hydrogen	H_2	0.000053
Nitrous oxide	N_2O	0.000031
Xenon	Xe	0.0000087
Ozone*	O_3	trace to 0.0008
Carbon monoxide	CO	trace to 0.000025
Sulfur dioxide	SO_2	trace to 0.00001
Nitrogen dioxide	NO_2	trace to 0.000002
Ammonia	NH_3	trace to 0.0000003

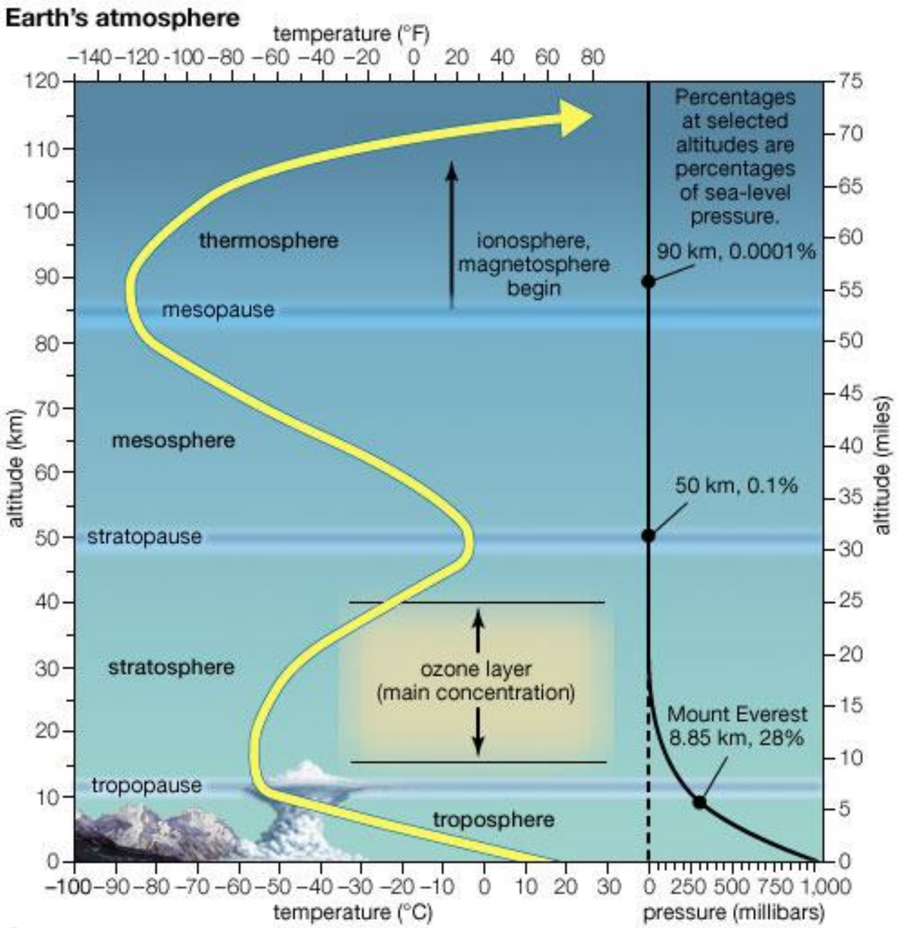
*Low concentrations in troposphere; ozone maximum is found at 30- to 40-km above Earth's surface in the equatorial region.

(After Warneck, 1988; Anderson, 1989; Wayne, 1991.)

Vertical distribution of constituents



Temperature & pressure latitude-altitude distribution



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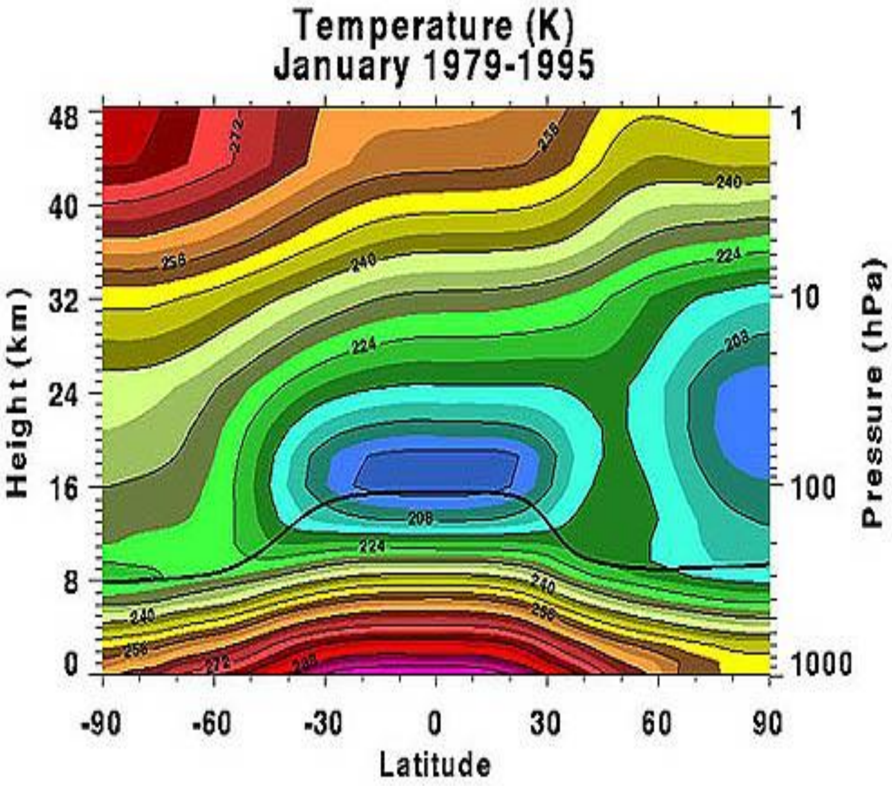


P. Newman (NASA), E. Nash (ARC), R. Nagatani (NCEP CPC)

80% of atmospheric air masses below 15km

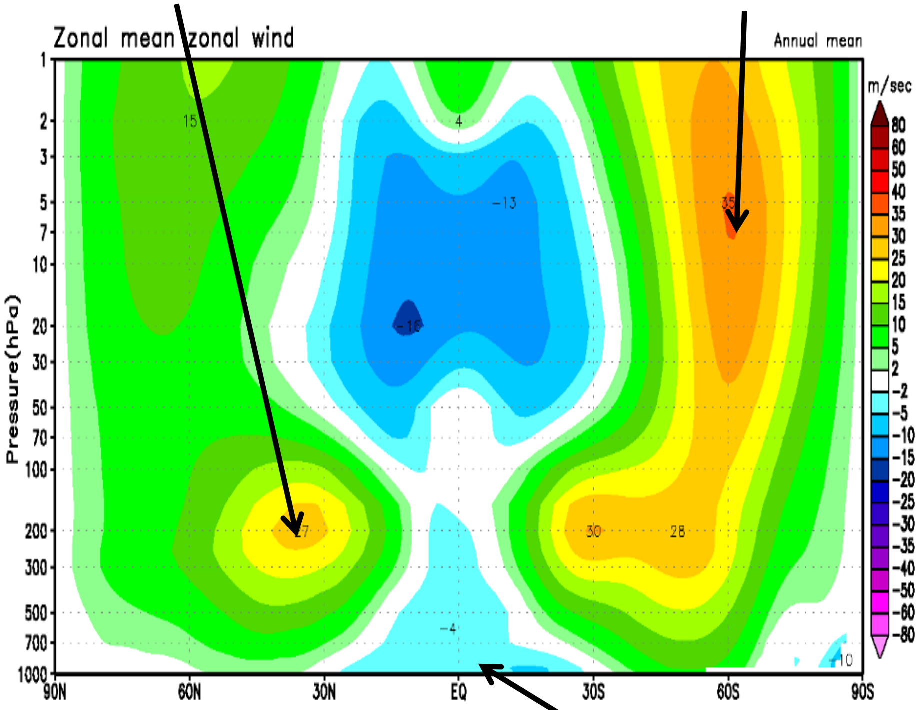
Large scale circulation

Annual zonal wind zonal mean (<U>)

Subtropical jet

Stratospheric jet

Pressure

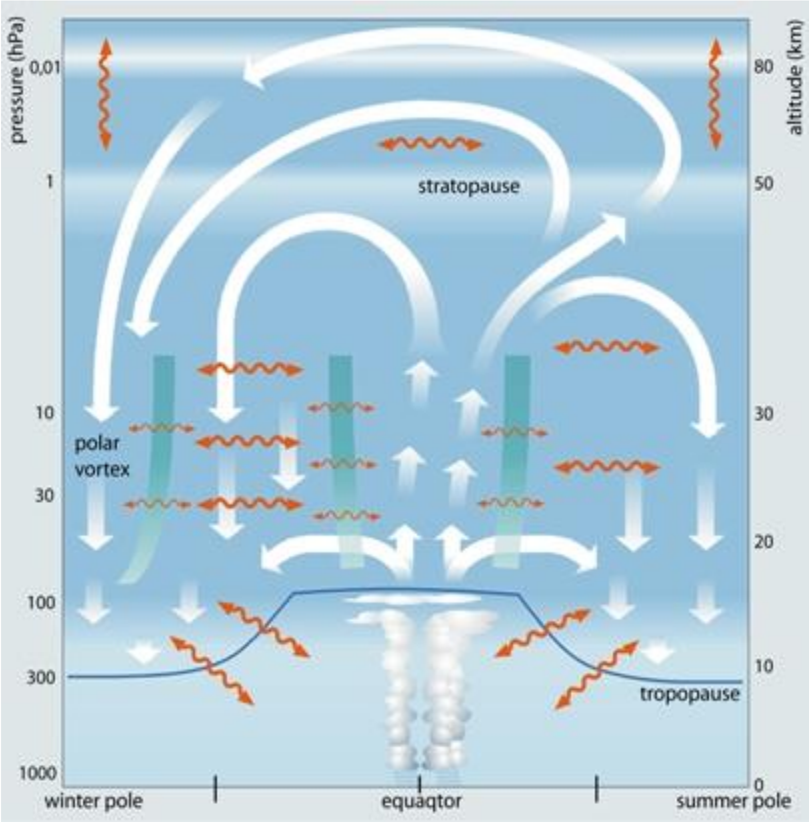


Latitude

Trade winds

(http://ds.data.jma.go.jp/gmd/jra/atlas/eng/indexe_isobar13.htm)

Stratosphere meridional overturning circulation



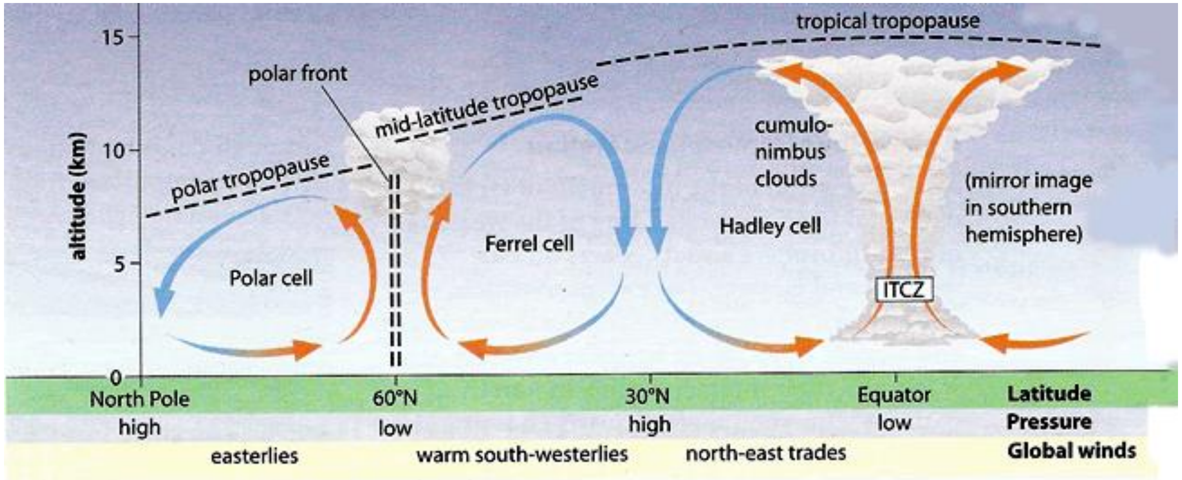
Latitude

(univ. Frankfurt)

Clear asymmetry between both hemisphere

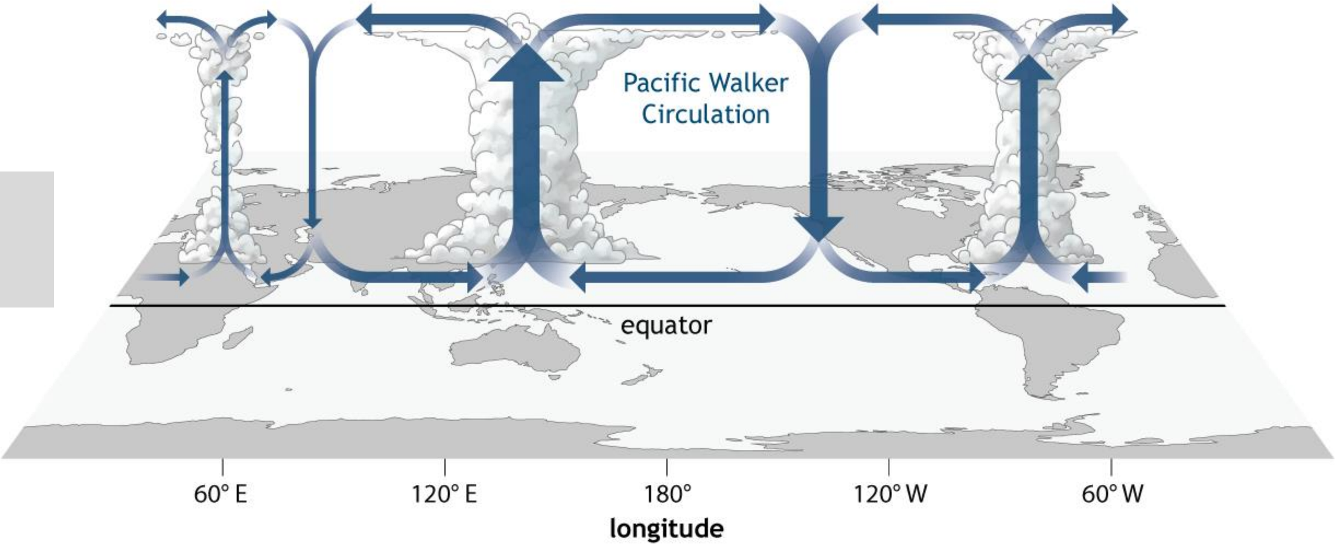
Large scale circulation (2)

Meridional/height circulation



Zonal/height circulation

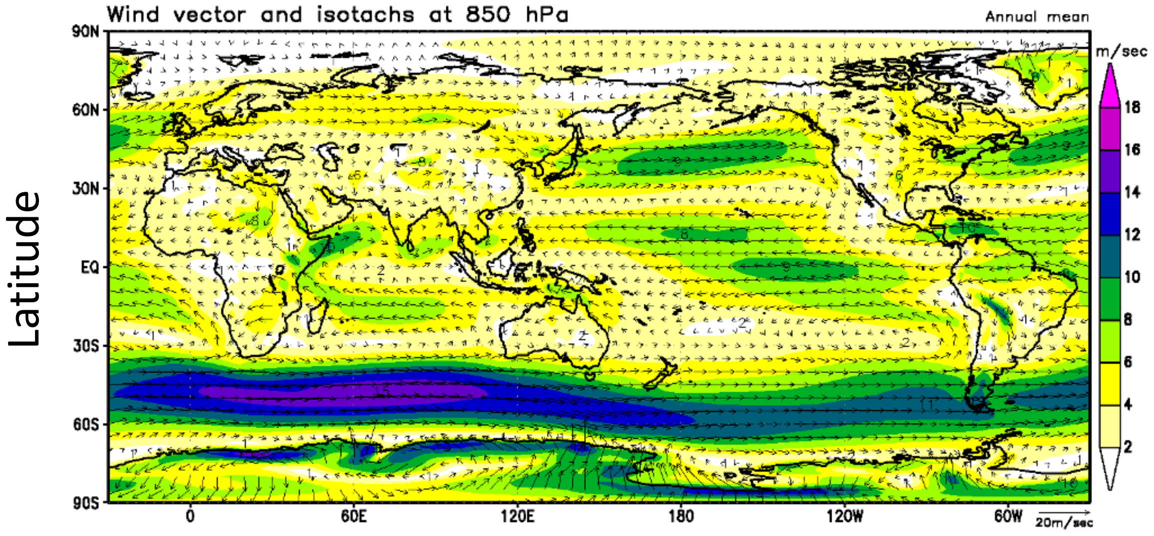
Neutral conditions



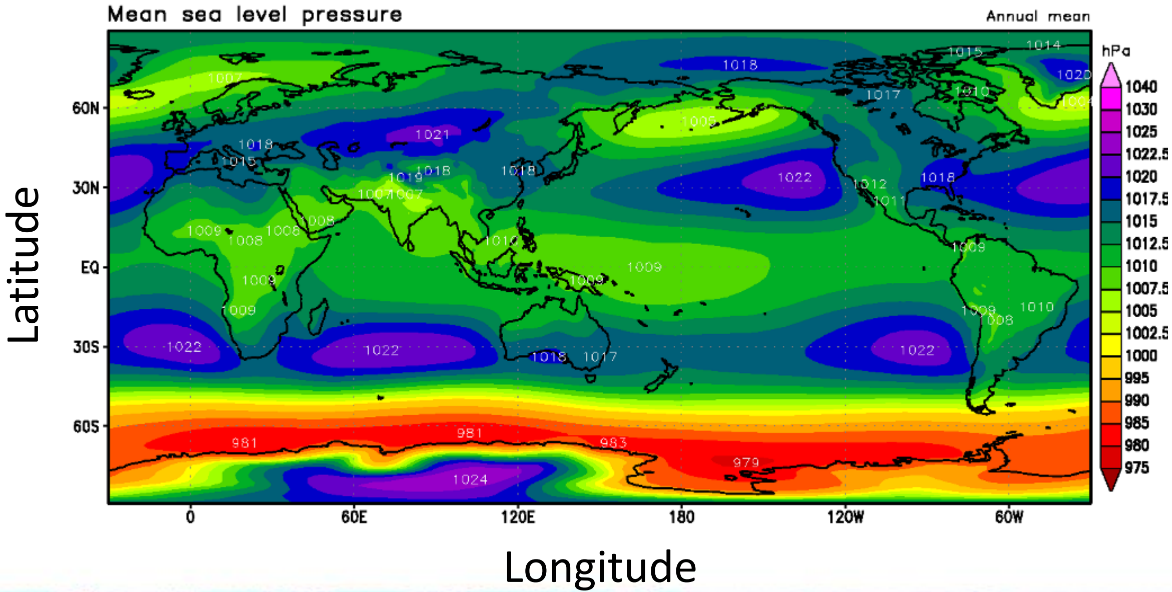
NOAA Climate.gov

Regional circulation

Annual mean wind direction and velocity at 850 hPa (~1.5 km)



Annual mean sea level pressure (hPa)



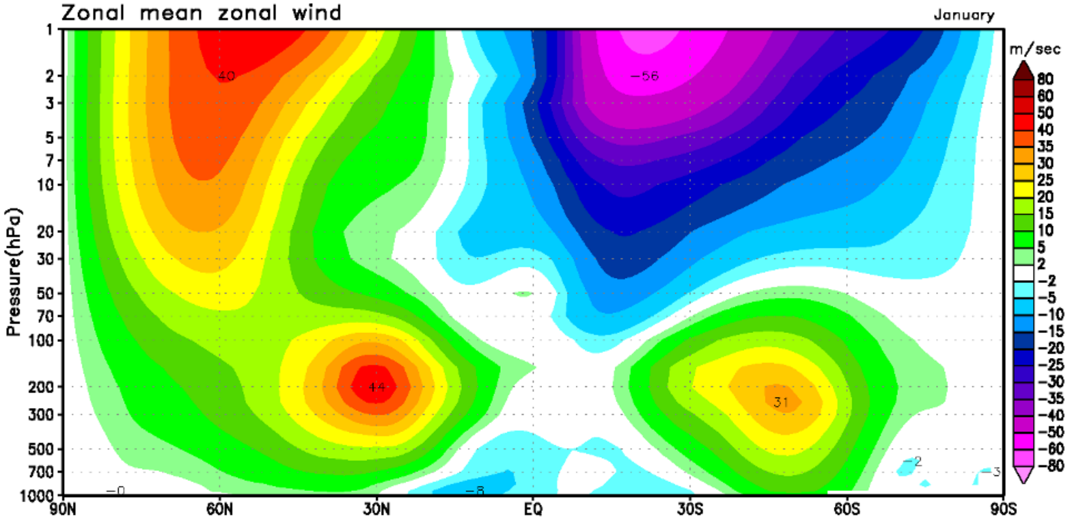
(http://ds.data.jma.go.jp/gmd/jra/atlas/eng/indexe_isobar13.htm)

Seasonality

Mean zonal wind zonal mean

January

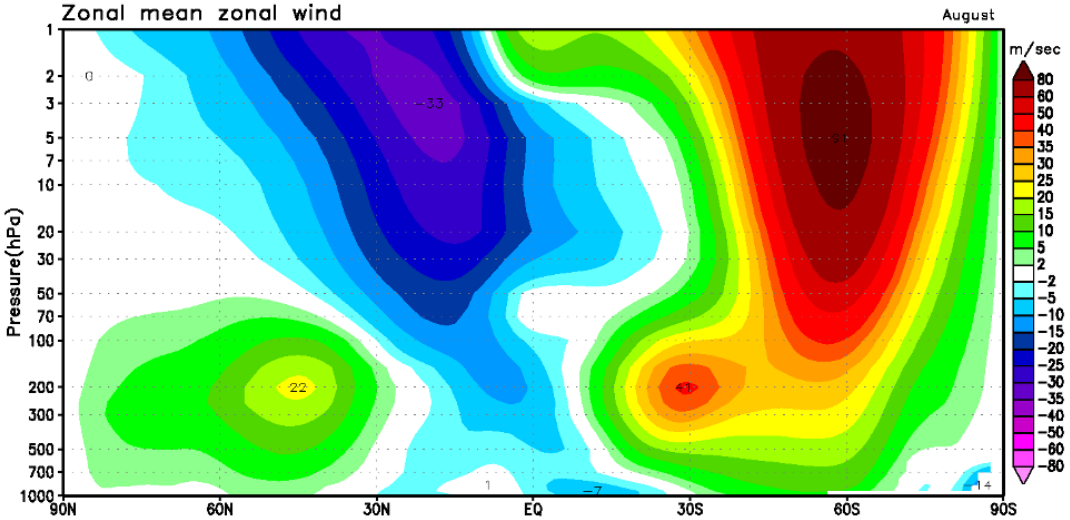
Pressure



Clear asymmetry between both hemispheres at different seasons

August

Pressure



(http://ds.data.jma.go.jp/gmd/jra/atlas/eng/indexe_isobar13.htm)

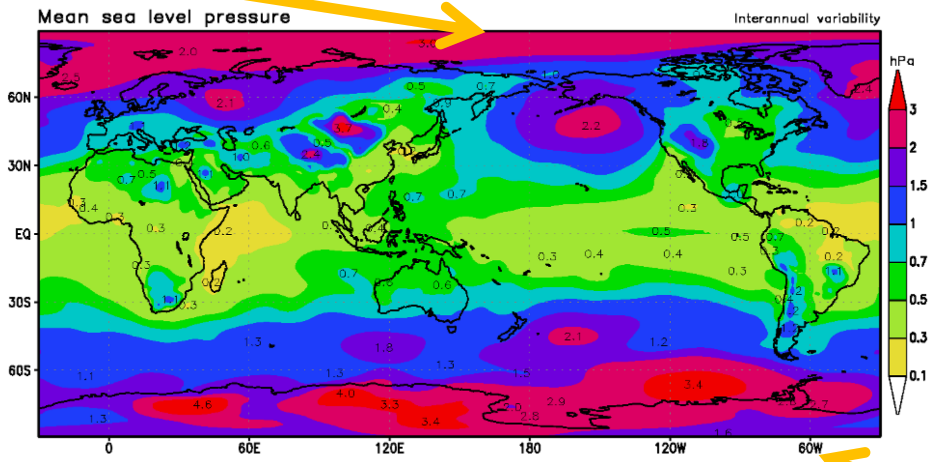
Variability

AO
(Arctic Oscillation)

Interannual standard deviation

SLP (hPa)

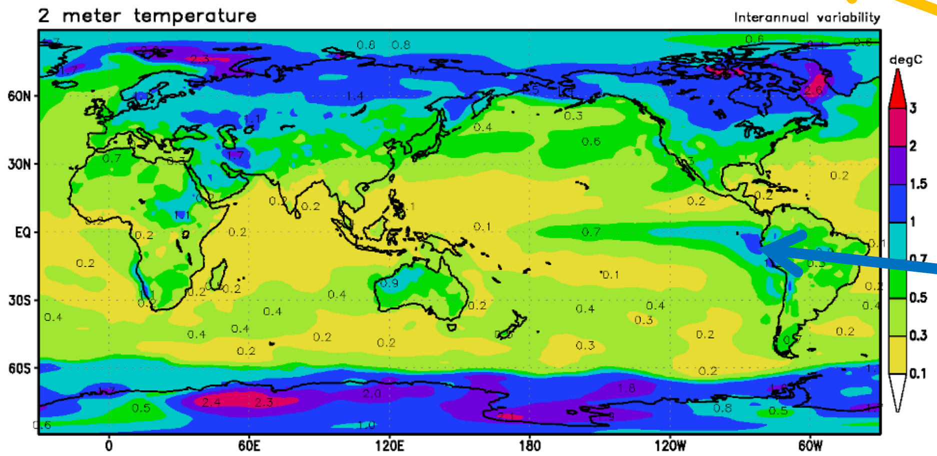
latitude



AAO
(Antarctic
Oscillation)

2m Temp

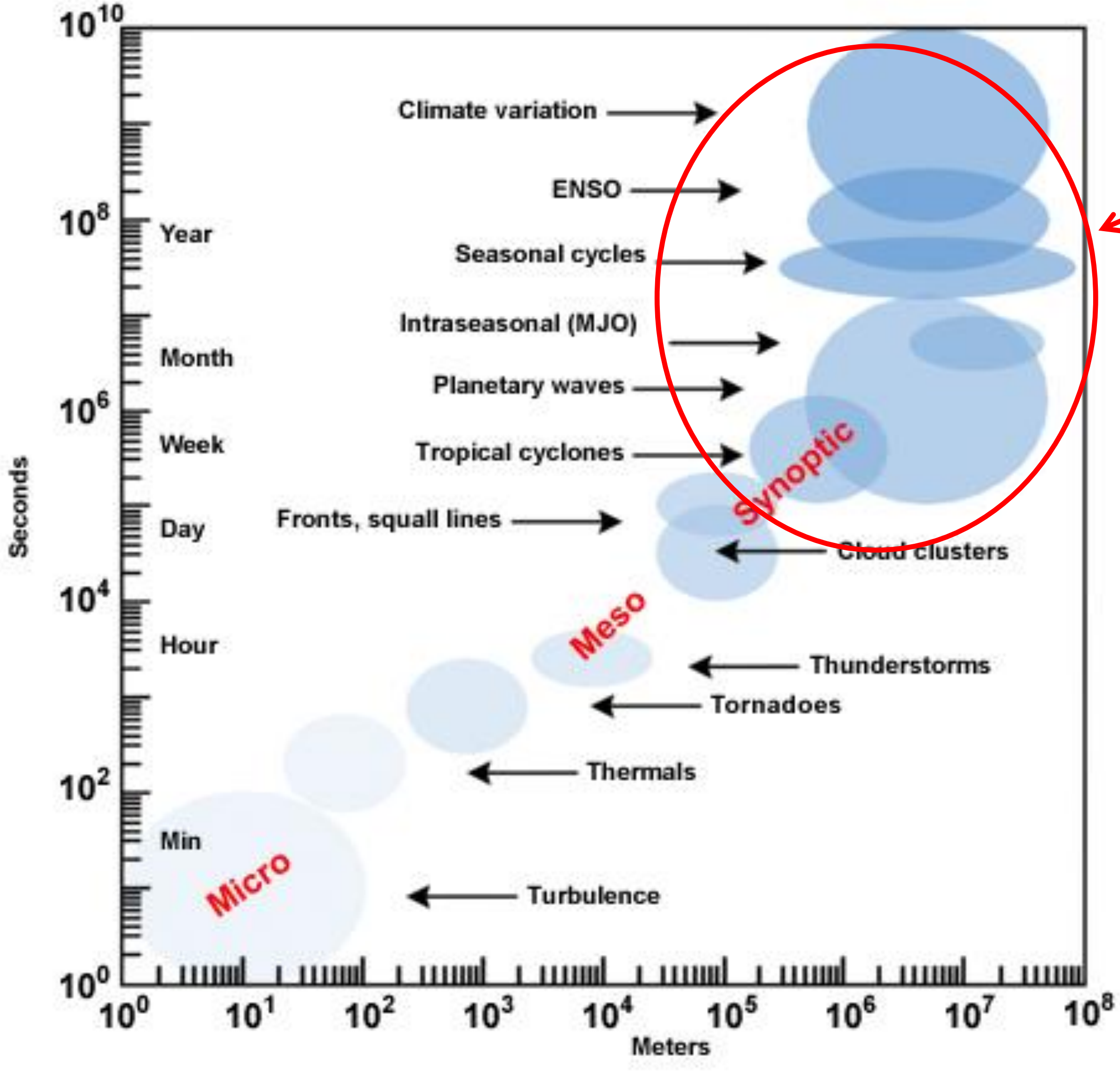
latitude



ENSO

longitude

Scales



Focus of the lecture

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What drive the atmosphere dynamical processes ?

Outline

I. Radiations

II. Thermodynamical processes

III. Dynamics

IV. Stratosphere

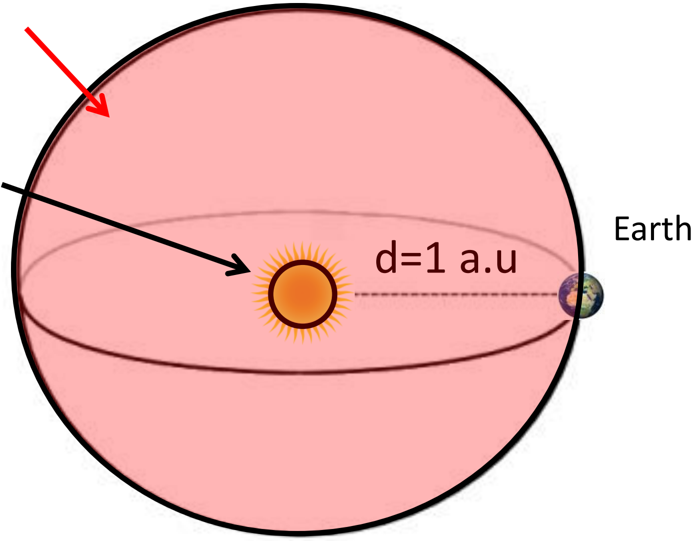
I. Earth‘s radiative balance

Energy source: the Sun

The sun can be considered as a „Black body“:

$$\Sigma = 4\pi d^2$$

$$\Sigma_s = 4\pi r_s{}^2$$



Energy flux (Stefan‘s law): $F_s = \sigma T_s{}^4$

T_s: Sun temperature
Rs: Sun radius

σ =5.67 10^{-8} W/m²/K⁴

Energy conservation: $\Sigma_s * F_s = \Sigma * F$

$$S = F_{TOA} = \sigma T_s^4 \left(\frac{r_s}{d}\right)^2 \approx 1360.8\, Wm^{-2}$$

(TOA: Top of the Atmosphere)

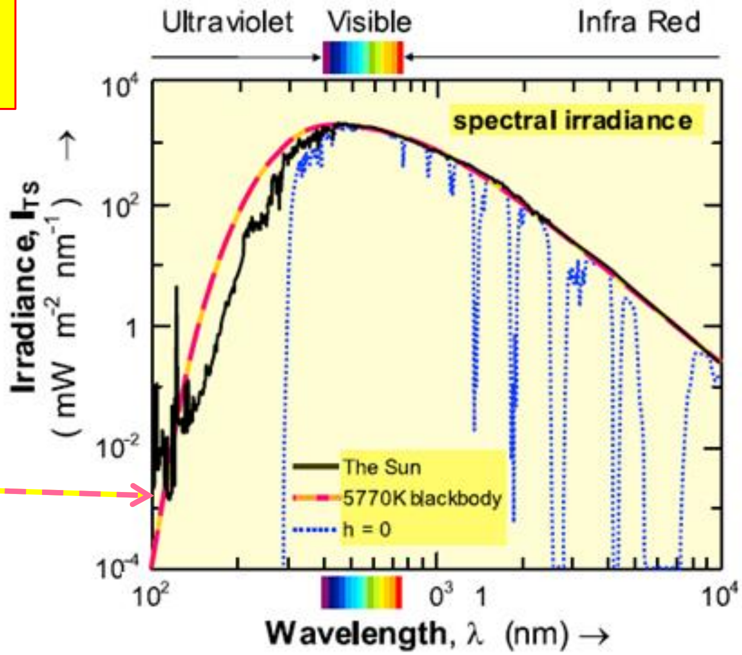
Solar constant (TSI: Total Solar Irradiance)

Spectral Solar Irradiance (SSI):

$$F(\lambda, T) = \frac{2\pi hc^2}{\lambda^5} \frac{1}{e^{hc/(\lambda kT)-1}}$$

(w/m²/m)

h: Planck constant, c: speed of light, k: Boltzmann constant



Gray et al., 2010

Earth‘s radiative balance: model without atmosphere

Energy balance:

Part of the SW are directly
reflected back to space

$$\pi a^2 (S - S\alpha) = 4\pi a^2 \sigma T_{earth}^4$$

a: Earth‘s radius
α: albedo

$$\boxed{\frac{S}{4}(1-\alpha) = \sigma T_{earth}^4}$$

$\alpha_{earth}=0.3$

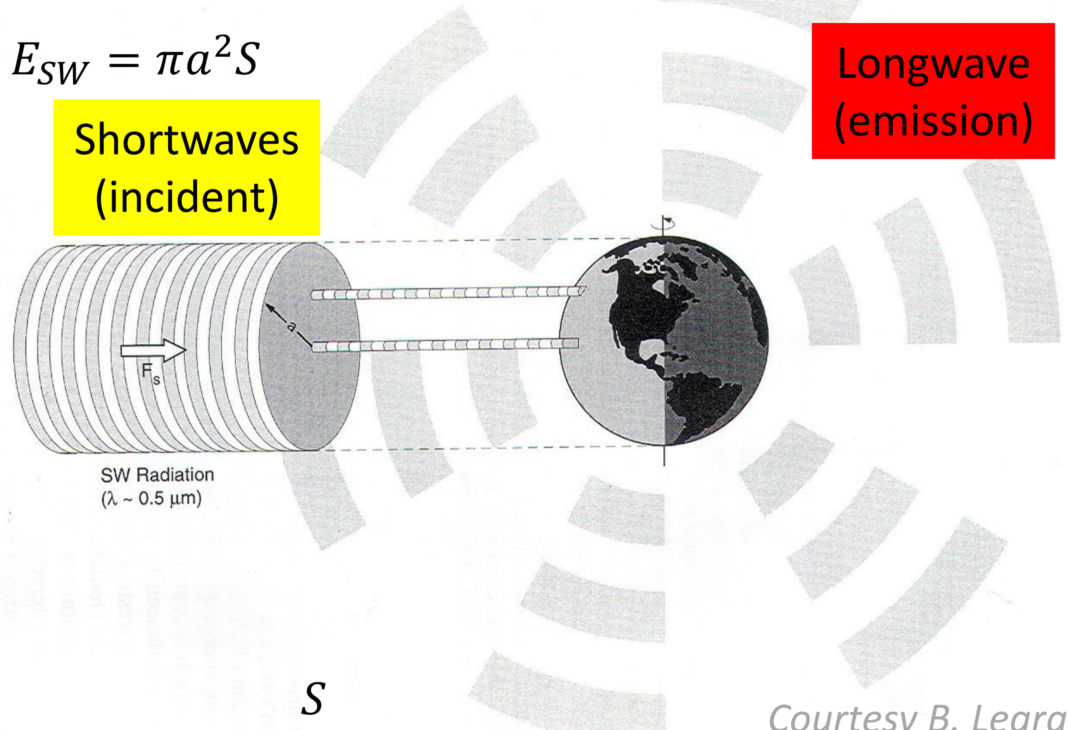
$T_{earth} = -18°C$

$$E_{LW} = 4\pi a^2 \sigma T_{earth}^4$$

$$E_{SW} = \pi a^2 S$$

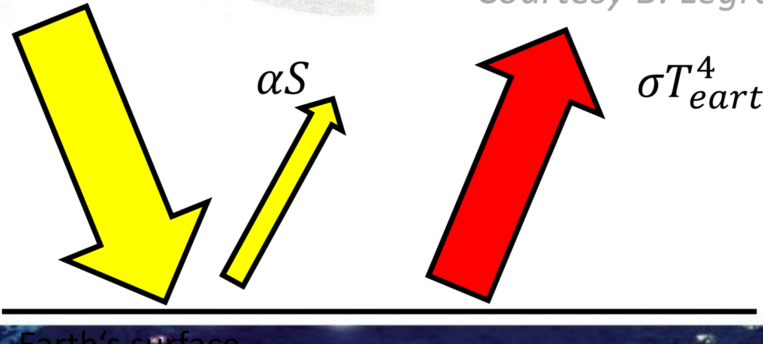
Shortwaves
(incident)

Longwave
(emission)



Courtesy B. Legras

S



Earth‘s radiative balance: single layer model

Atmosphere is transparent to visible

Energy balance surface: $\pi a^2(S - S\alpha) - 4\pi a^2\sigma T^4_{surface} + 4\pi a^2\sigma T^4_{atmosphere} = 0$

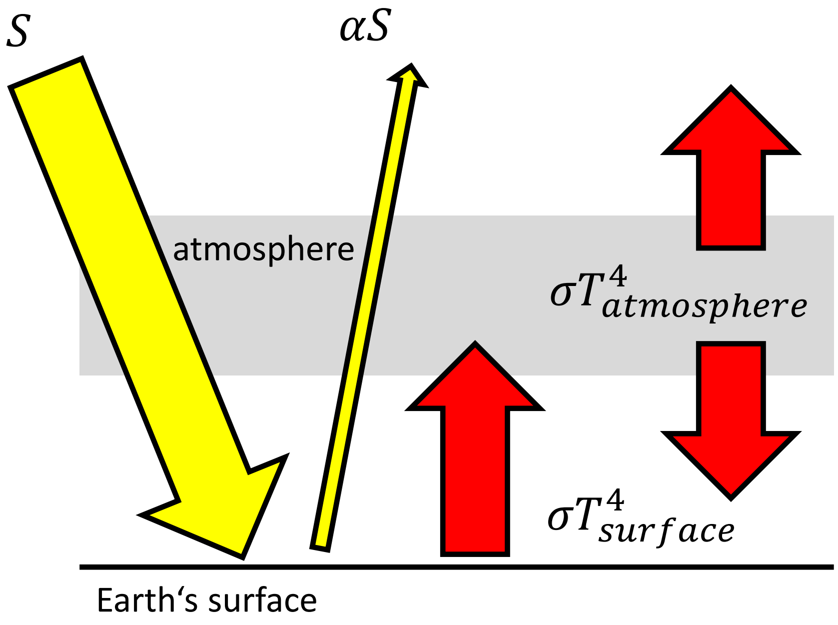
Energy balance atmos: $4\pi a^2\sigma T^4_{surface} = 2 * 4\pi a^2\sigma T^4_{atmosphere}$

σT^4_{eq}: (temperature without atmosphere)

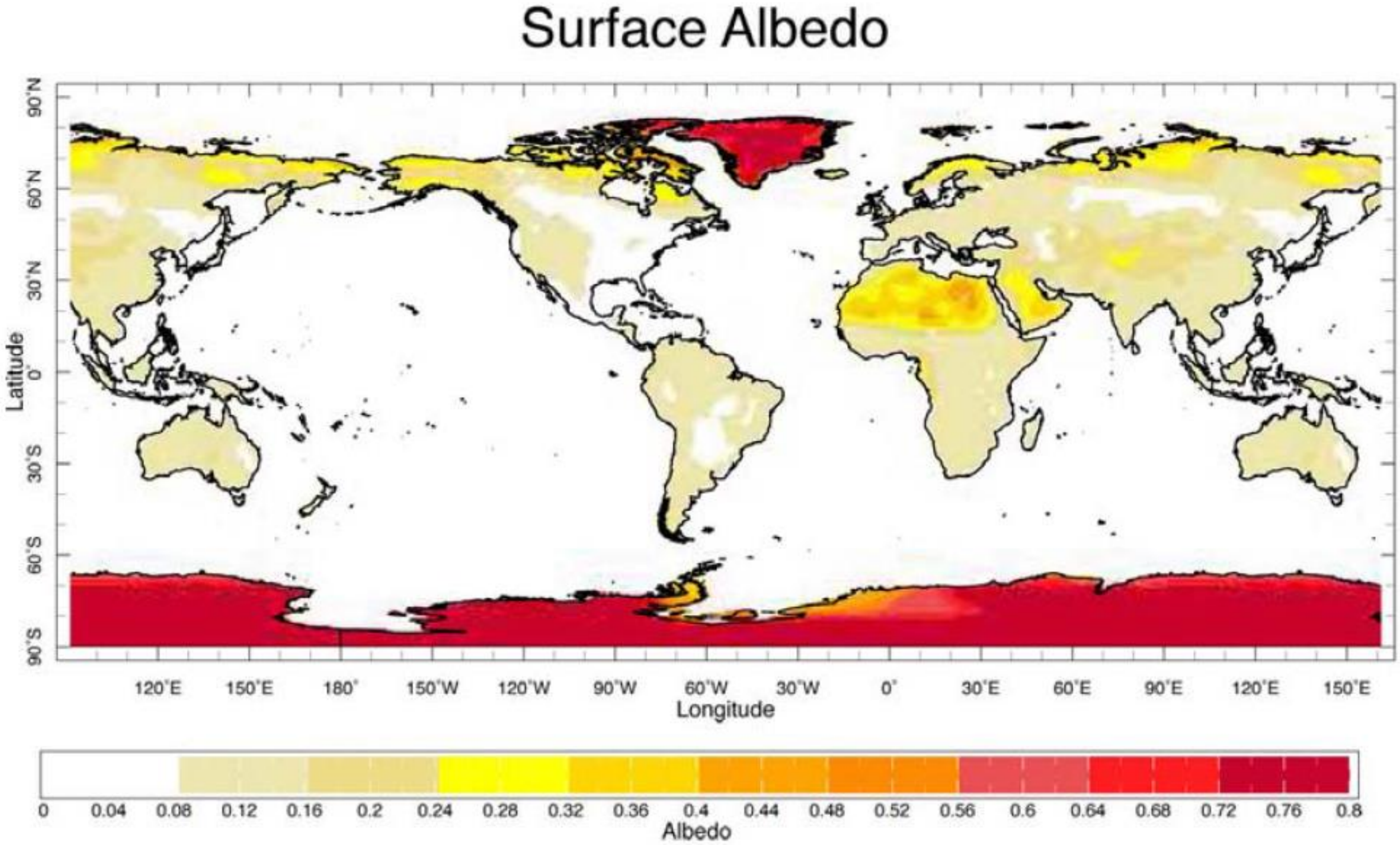
$$\frac{S}{4}(1-\alpha) - \frac{1}{2}\sigma T^4_{surface} = 0$$

$$T^4_{surface} = 2T^4_{eq}$$

$$T_{surface} = 15°C$$



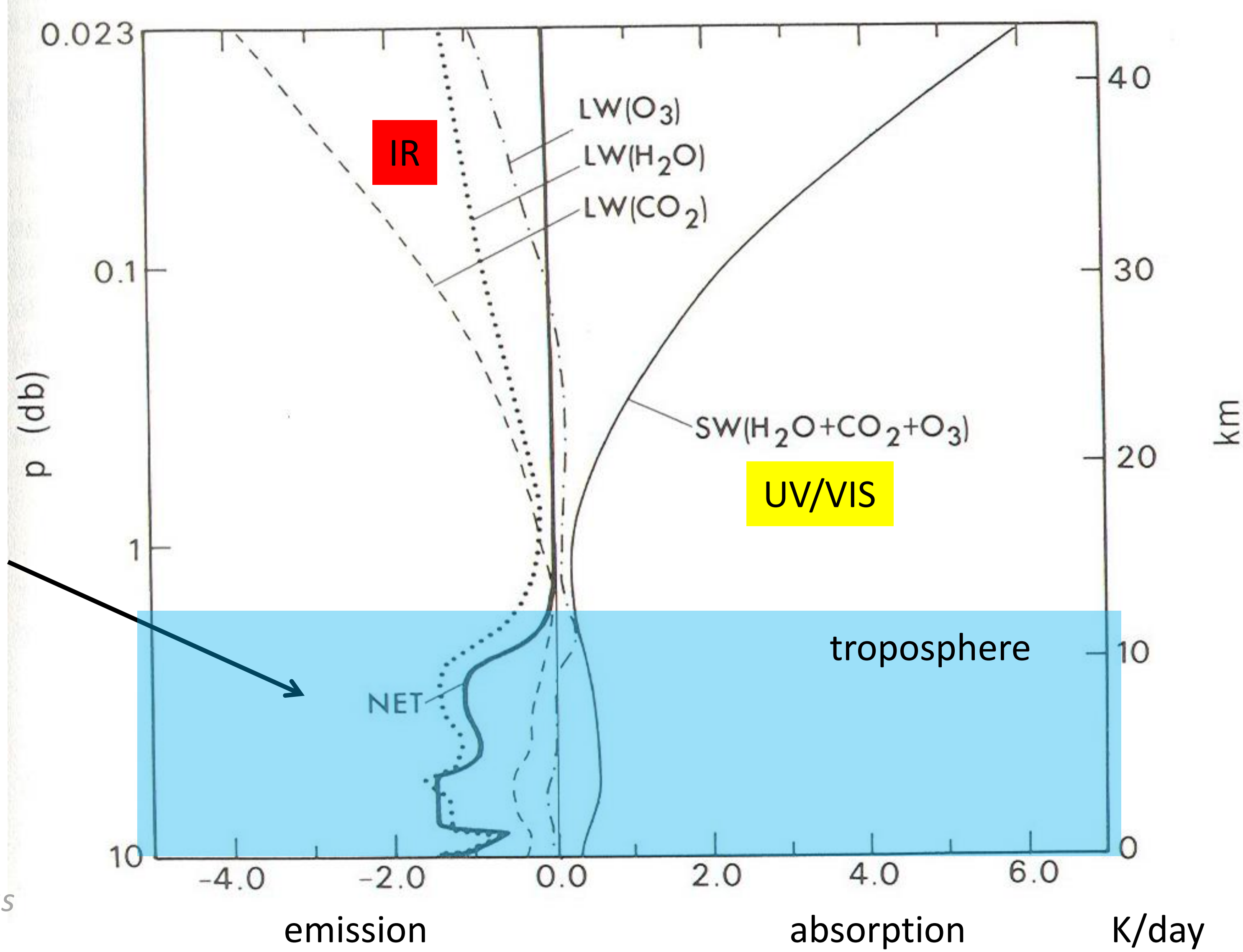
Earth‘s albedo



Earth‘s radiative balance: vertical profile

Stratosphere in radiative balance (UV absorption by O_3 and IR emission by CO_2).

Troposphere must be balanced.

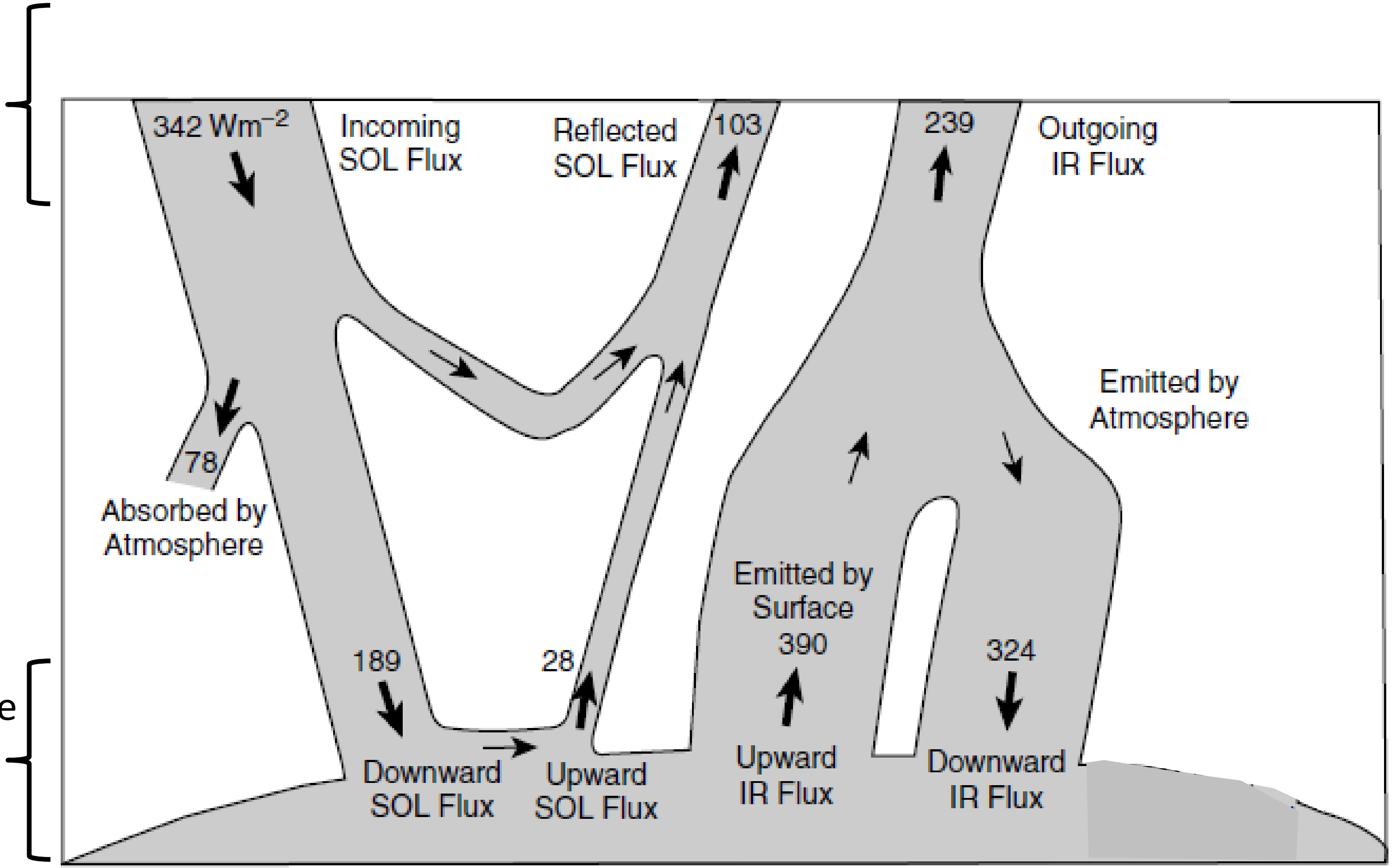


Courtesy B. Legras

Earth‘s energy budget

TOA: balanced

Troposphere radiatively unbalanced

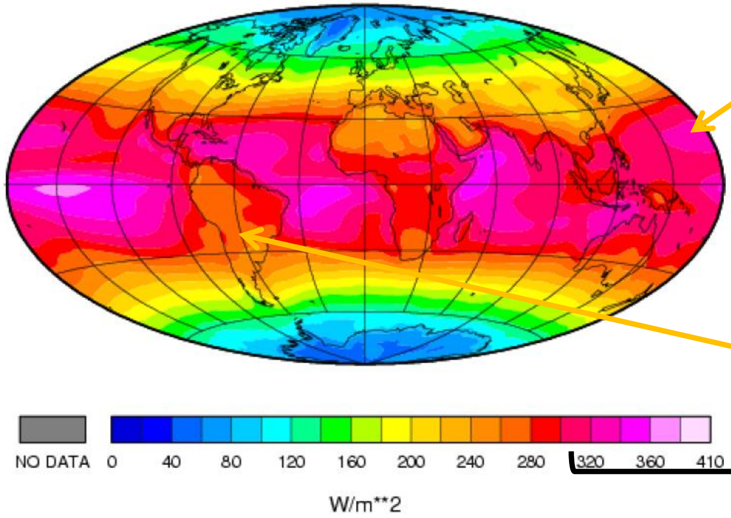


Liou, 2002

Earth‘s radiative balance

UV/VIS

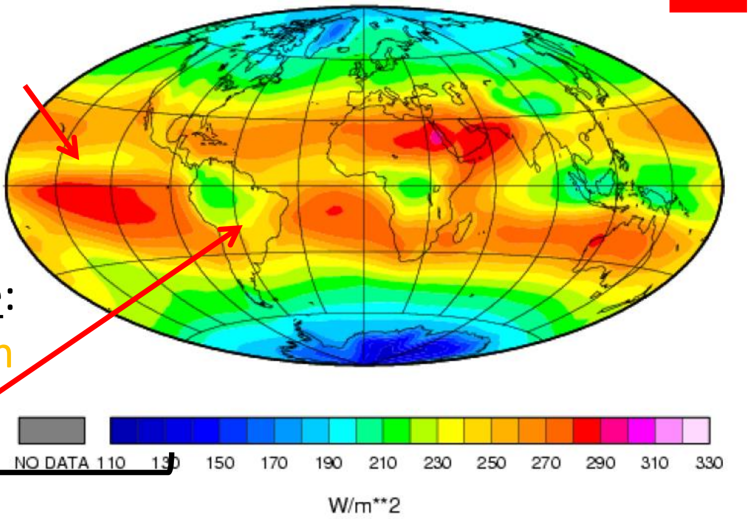
Absorbed Shortwave Radiation
1985-1986



Ocean: **strong** absorption and emission

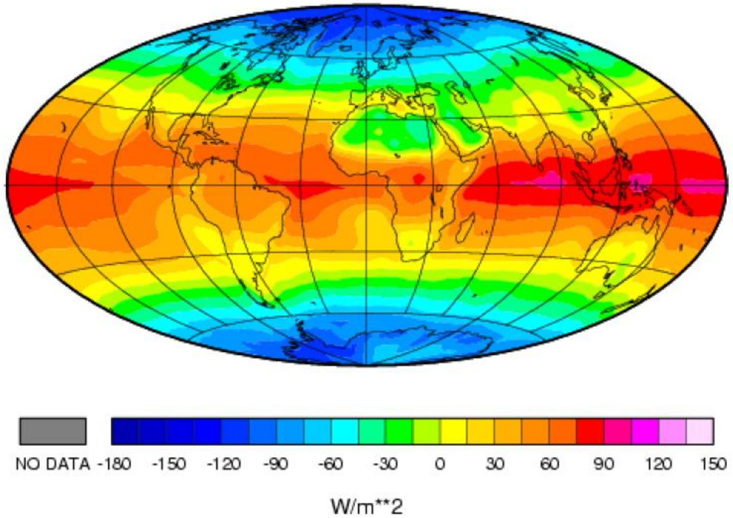
IR

Outgoing Longwave Radiation
1985-1986

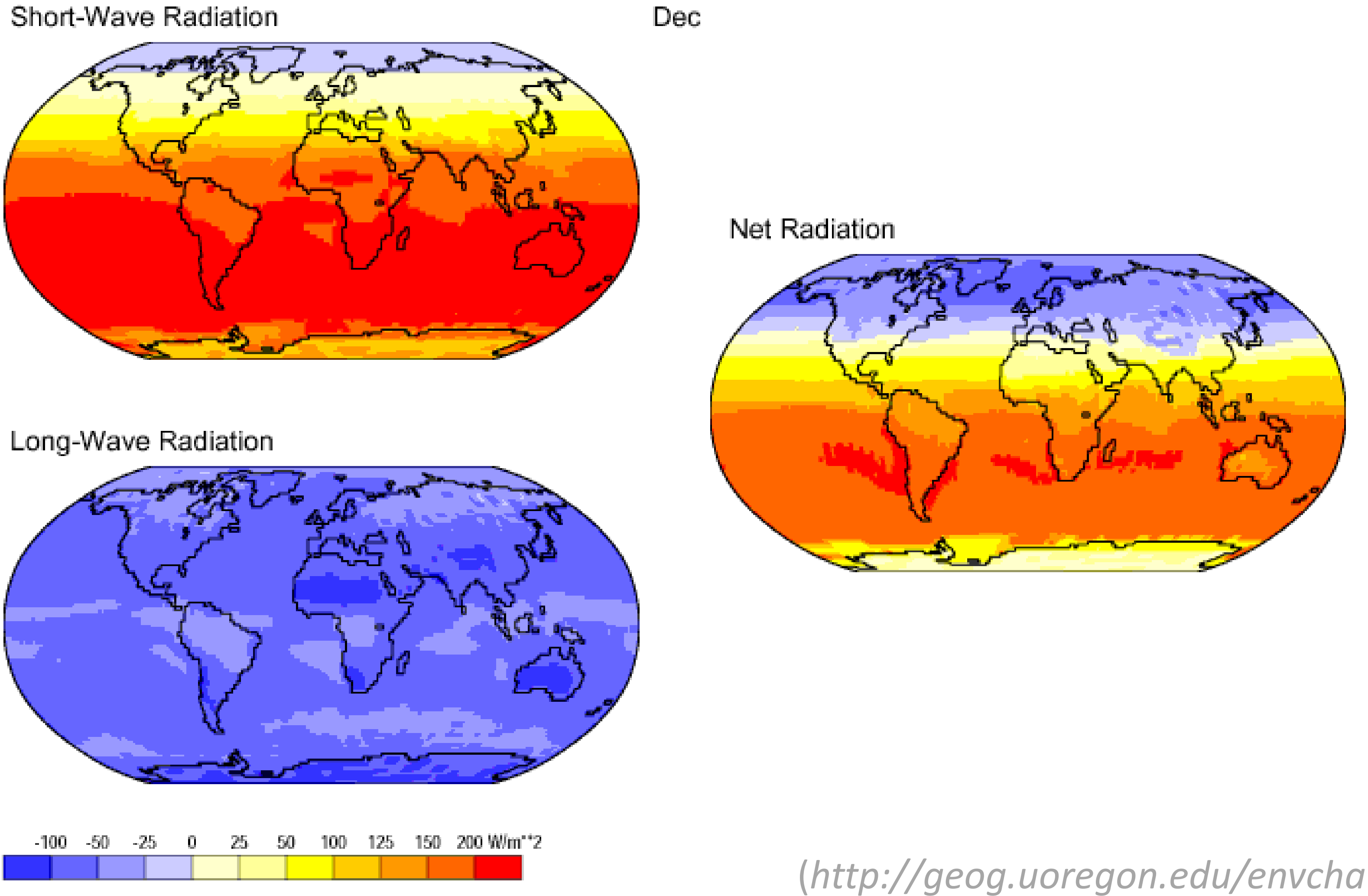


Convective zone: **weak** absorption and emission

Net Radiation
1985-1986



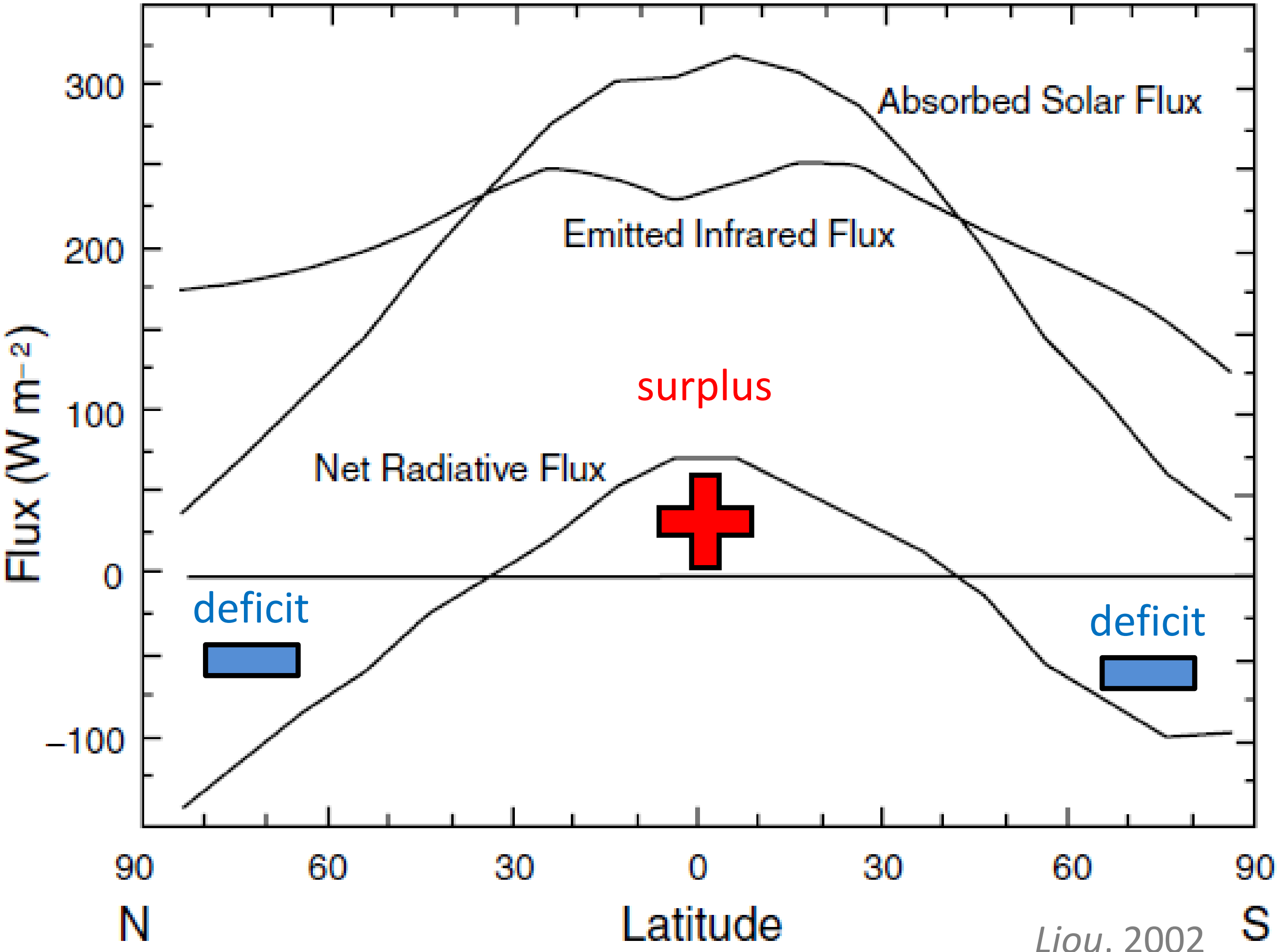
Earth‘s radiative balance: seasonality



Data: NCEP/NCAR Reanalysis Project, 1959-1997 Climatologies
Animation: Department of Geography, University of Oregon, March 2000

(http://geog.uoregon.edu/envchange/clim_animations/)

Earth‘s distribution



Liou, 2002

Driver of the atmosphere‘s general circulation

II. Thermodynamical processes

Pressure-altitude equation

For dry air (⇔ no water vapor):

Gas constant: 8.314 J/(K.mol)

Gas law: $\boxed{p = \rho_{air} R T}$,with $R = R^* / M_{air}$

$$\cong 0.21 \times M_{O_2} + 0.78 \times M_{N_2} + 0.01 \times M_{Ar} = 28.9\ g/mol$$

Hydrostatic equation: $\boxed{{dp}/{dz} + \rho g = 0}$

$$dp/p = \frac{-gM_{air}}{R^*T} dz$$

integrate

$$\boxed{p(z) = p_0 e^{\left(-\int_{z_0}^{z} \frac{gM_{air}}{R^*T(z)} dz\right)}}$$

Under isothermal conditions (T_0=255 K): $p(z) = p_0 e^{(-z/H)}$, with $H=R^*T_0/(gM_{air})$=7.4 km

H (scale height) ⇔ pressure divided by 2 every 5 km (*Hln(2)=5km*)

Standard Atmosphere Approximation (ICAO)

Ground conditions: P_0=1013 hPa, T_0=15°C

Troposphere (0-11 km): $T=T_0+\Gamma z$ with Γ=-6.5°C/km

$$T(z) = T_0 + \Gamma z \qquad {dp}/_{p} = \frac{-gM_{air}}{R^*T(z)}dz$$

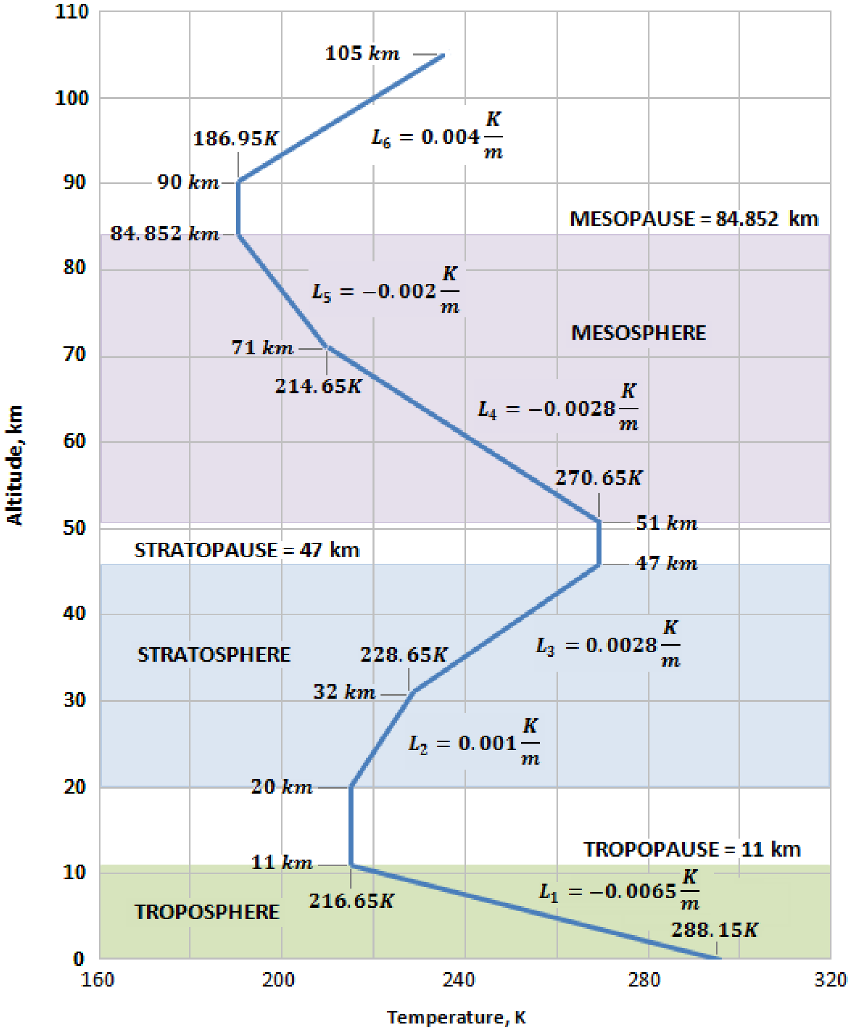
$$d(T_0 + \Gamma z\,) = \Gamma dz$$

$$ {dp}/_{p} = \frac{-gM_{air}}{\Gamma R^*}\frac{d(T_0+\Gamma z)}{(T_0+\Gamma z)}$$

$$\boxed{\frac{P}{P_0} = \left(\frac{T_0+\Gamma z}{T_0}\right)^{-\frac{M_a g}{\Gamma R}}}$$

Γ ⇔ Environmental laspe rate

Definition of standard temperatures at different altitudes



Dry adiabatic lapse rate

First law of thermodynamics: $dU = \delta Q + \delta W$

$$\delta Q = TdS = C_v dT + pdV$$

For adiabatic transformation: $\delta Q = 0 = C_v dT + pdV$

$$pV = nR^*T \rightarrow pdV + Vdp = nR^*dT$$

$$0 = (nR^* + C_v)dT - Vdp$$

$$C_p dT = Vdp$$

$$dp = -\rho_{air} g dz = -{}^{m_{air}}/_{V}\, gdz$$

$$V = m_{air} RT/p$$

$$c_p = C_p/m_{air}$$

$$\frac{dT}{dz} = -\frac{g}{c_p}$$

9.81 m/s²

1006 J/(kg.K)

$$dp\,/p = c_p/R \times dT/T$$

$$\int_{standard}^{T,p}$$

$$\Gamma' = \frac{dT}{dz} \approx -10K/km$$

Adiabatic laspe rate

$$\theta = T_0 = T\left(\frac{p_{standard}}{p}\right)^{R/c_p}$$

Potential temperature⇔ Temperature that the air parcel would acquire if adiabatically brought to a standard pressure

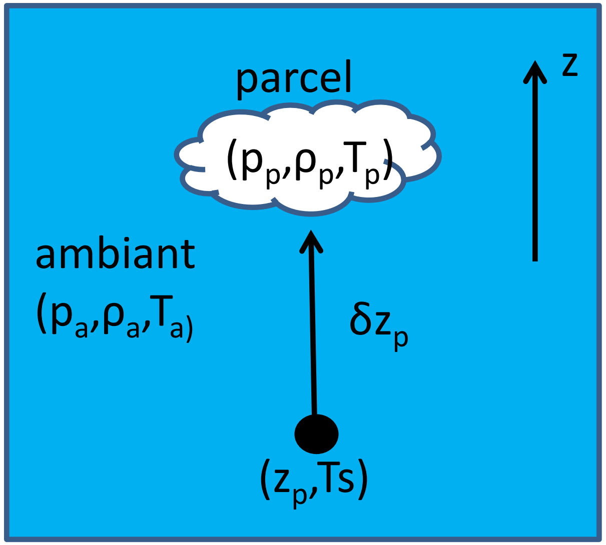
Vertical stability

Vertical momentum equation for the displaced air parcel:

$$\text{(z)} \quad \rho_p \frac{d^2\delta z_p}{dt^2} = -\rho_p g - \frac{\partial p_a}{\partial z}$$

... combined with hydrostatic equation applied to ambiant air:

$$\rho_a g + \frac{\partial p_a}{\partial z} = 0$$



Gas law & $p_a=p_p$

$$\Rightarrow \quad a_p = \frac{d^2\delta z_p}{dt^2} = g\frac{\rho_a - \rho_p}{\rho_p} = g\frac{T_p - T_a}{T_a}$$

With
$$\left.\begin{aligned} T_p &= T_s + \delta T_p \\ T_a &= T_s + \delta T_a \end{aligned}\right\}$$

$$a_p = g\frac{\delta T_p - \delta T_a}{T_a}$$

$$\theta = T\left(\frac{p_0}{p}\right)^{\kappa}$$
$$\frac{\delta\theta}{\theta} = \frac{\delta T}{T} - \kappa\frac{\delta p}{p}$$
$$\delta\theta_p = 0$$

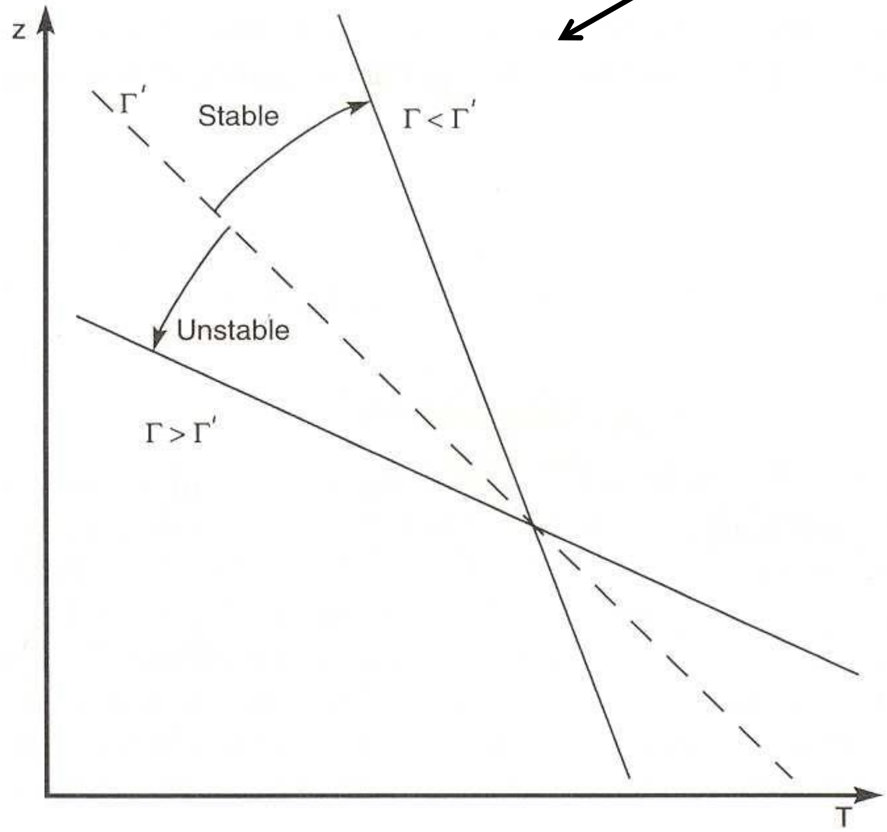
$$\boxed{a_p = \frac{g}{T_a}\left[\left(\frac{\partial T}{\partial z}\right)_p - \left(\frac{\partial T}{\partial z}\right)_a\right]\delta z_p}$$

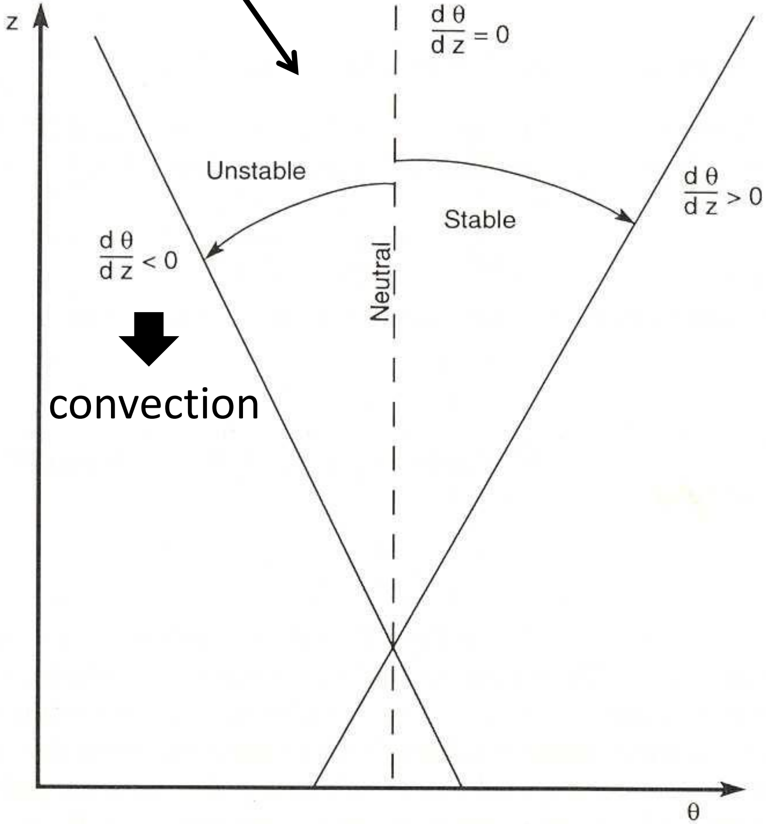
$\uparrow$ Γ'(adiabatic) $\uparrow$ Γ

$$\boxed{a_p = -g\frac{\delta\theta_a}{\theta_a} = -\frac{g}{\theta_a}\left(\frac{\partial\theta_a}{\partial z}\right)\delta z_p}$$

Vertical stability: application

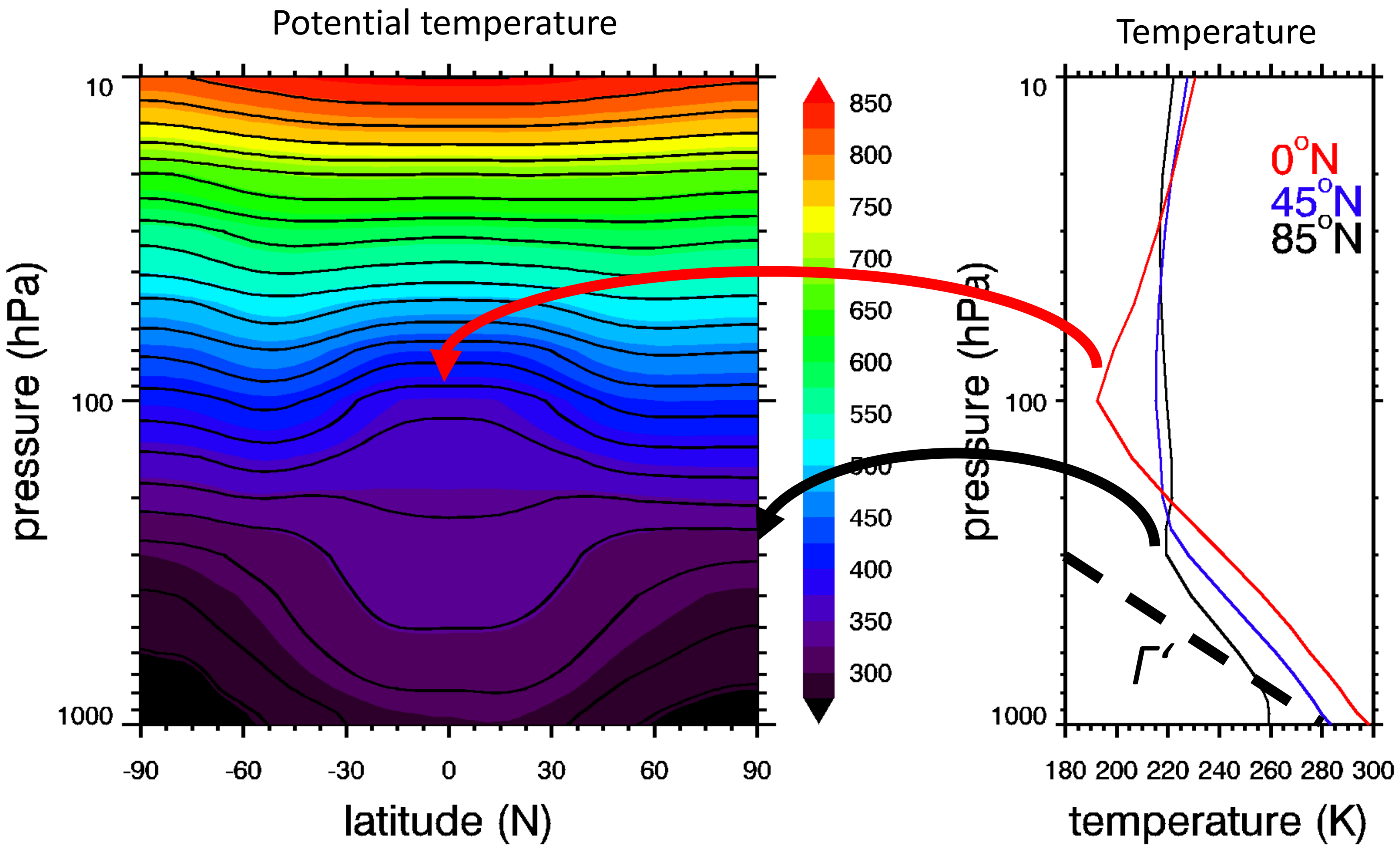
$$\frac{d^2\delta z_p}{dt^2} = \frac{g}{T_a}[\Gamma' - \Gamma_a]\delta z_p = -\frac{g}{\theta_a}\left(\frac{\partial\theta_a}{\partial z}\right)\delta z_p$$



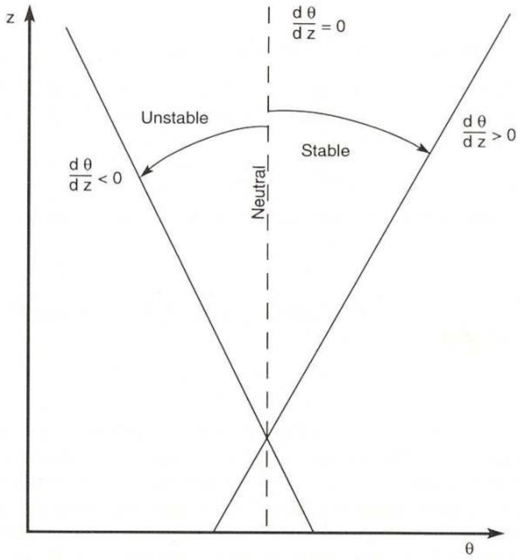


convection

Potential temperature mean vertical profile



Atmospheric boundary layer diurnal cycle



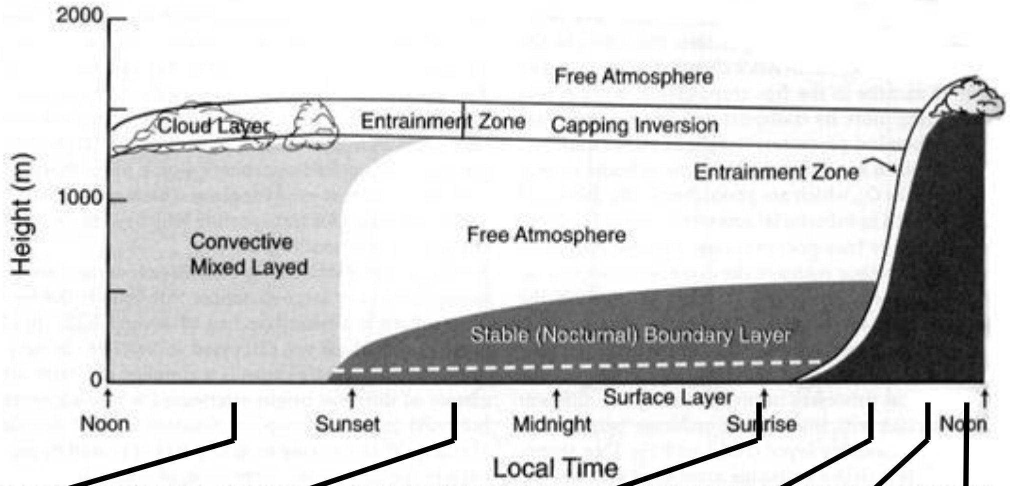
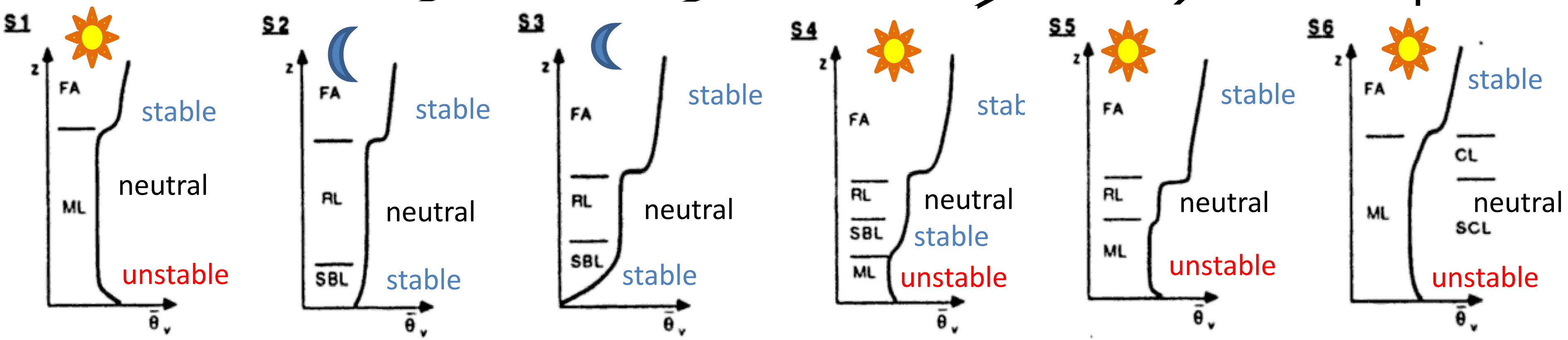


Figure 2.7. Typical diurnal evolution of the planetary boundary layer under a high pressure system. (adapted from Stull, Boundary Layer Meteorology, Kluwer, 1988)



Brünt-Vaissala frequency : acceleration along z axis

$$\underbrace{\frac{d^2\delta z_p}{dt^2} = -\frac{g}{\theta_a}\left(\frac{\partial \theta_a}{\partial z}\right)\delta z_p = -N^2\delta z_p}_{\text{Harmonic oscillator !}}$$

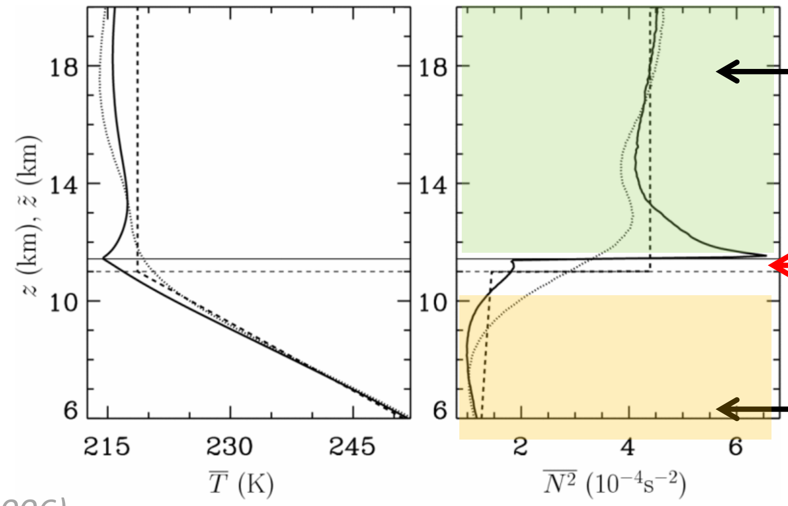
, with $\boxed{N^2 = \frac{g}{\theta_a}\left(\frac{\partial \theta_a}{\partial z}\right)}$

N is the Brünt-Vaissala frequency

if $\frac{\partial \theta_a}{\partial z} < 0$ ⇔ <u>unstable</u>: Convection

if $\frac{\partial \theta_a}{\partial z} > 0$ then $N \in \mathbb{R}$ ⇔ <u>stable</u>: the parcel oscillates with the frequency N or the period *T=2π/N*

Radiosounding at midlatitudes from 1998 to2002



T~5 min ⇔ more stable than the troposphere

Tropopause: very high static stability ⇔ very few vertical transport

T~10 min

(Birner, 2006)

Atmospheric water vapor

Water vapour partial pressure

- Water vapour mass mixing ratio:

$$r = \frac{m_v}{m_{dry\ air}} = \frac{M_v}{M_{dry\ air}} \frac{P_v}{P_{dry\ air}}$$

$$\boxed{r = 0.622 \frac{P_v}{P - P_v}}$$

with $P = P_{dry\ air} + P_v$

- Saturated air:

$$\boxed{\frac{1}{P_{v,sat}} \frac{dP_{v,sat}}{dT} = \frac{L(T)}{R_v T^2}}$$

with $R_v = R^*/M_v$

L : specific latent heat of vaporization

Clausius-Clapeyron

$$P_{v,sat} = 6.112 e^{\left(17.67 \frac{T-273.15}{T-29.65}\right)}$$

@1000 hPa, 20°C : r_{sat}=14.5 g/kg

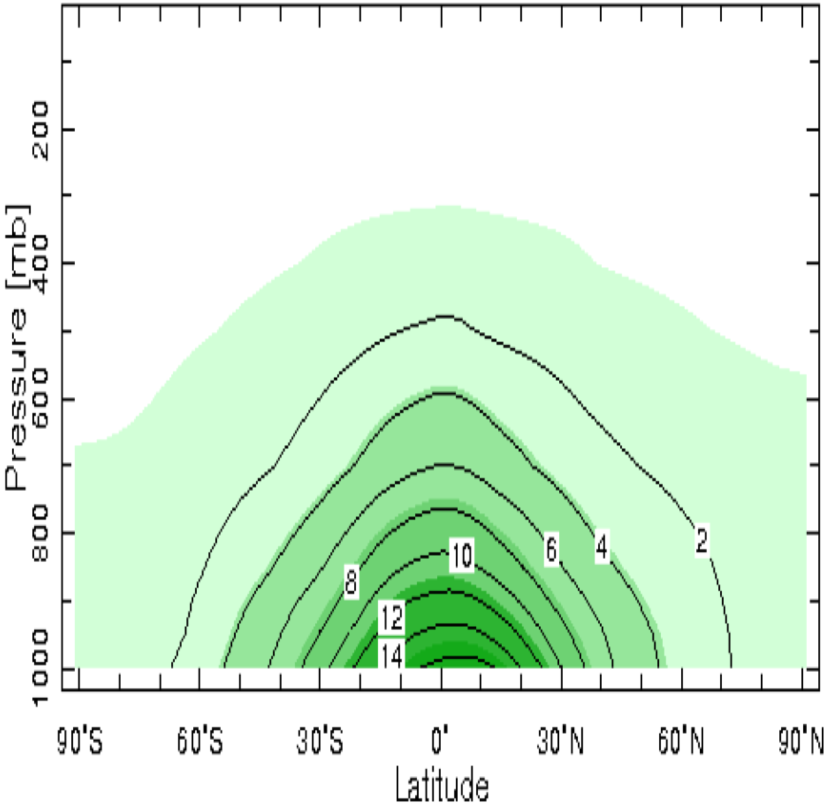
@500 hPa, -30°C : r_{sat}=0.47 g/kg

@100 hPa, -80°C : r_{sat}=0.003 g/kg

Clausius-Clapeyron law limits the air water vapor content by fixing the saturated pressure as a function of T and P.

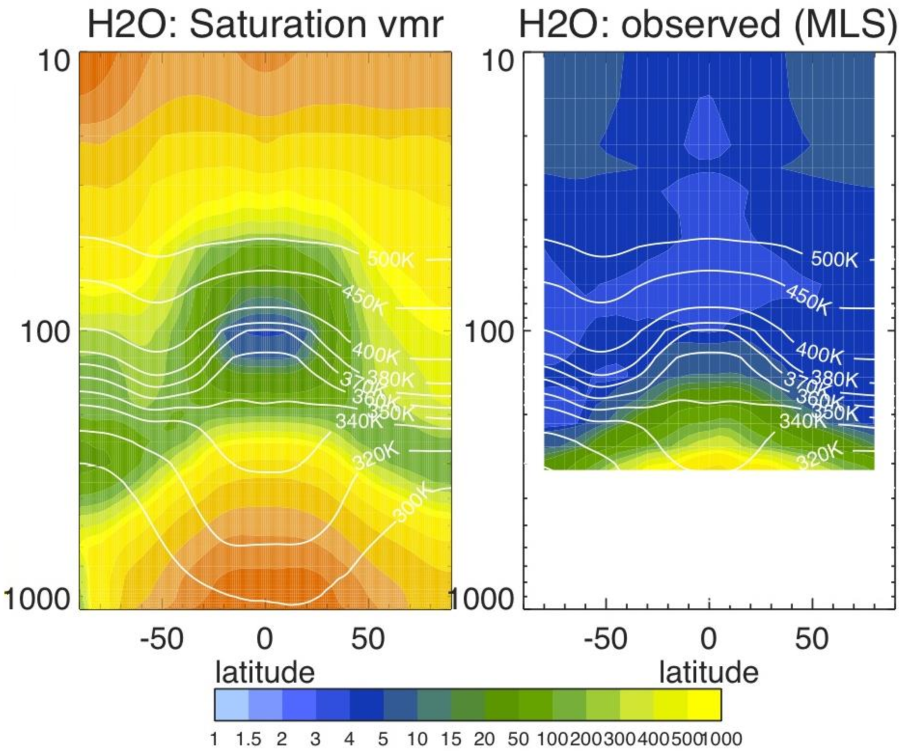
Atmospheric water vapor (2)

Mean water vapor mass mixing ratio distribution (g/kg)
(troposphere)



Courtesy F. Codron

Water vapor volume mixing
(stratosphere)



! The stratosphere is extremely dry !

Saturated lapse rate

Dry adiabatic conditions: $\delta Q = 0 = C_p dT + g dz$

If the water vapor in the air parcel is condensing during the ascent: $\delta Q = -L dr$

Pseudo-adiabatic hypothesis:
The water vapour (liquid or vapour) is negleted but the latent heat release is considered.

(r decreases if condensation:
r mass vapor mixing ration)

$$0 = C_p dT + g dz + L dr$$

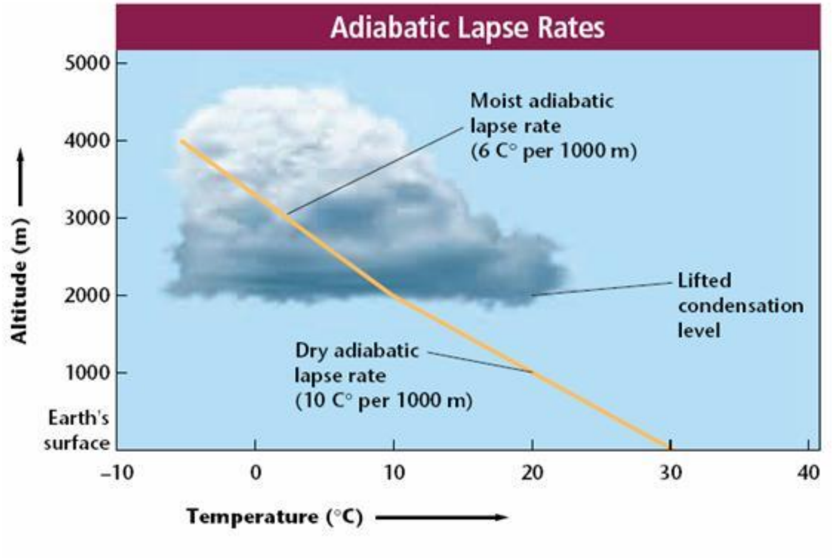
for saturated air: $r = r_{sat}(T, p)$, thus $dr_{sat}(T, p) = \frac{\partial r_{sat}}{\partial T} dT + \frac{\partial r_{sat}}{\partial p} dp$

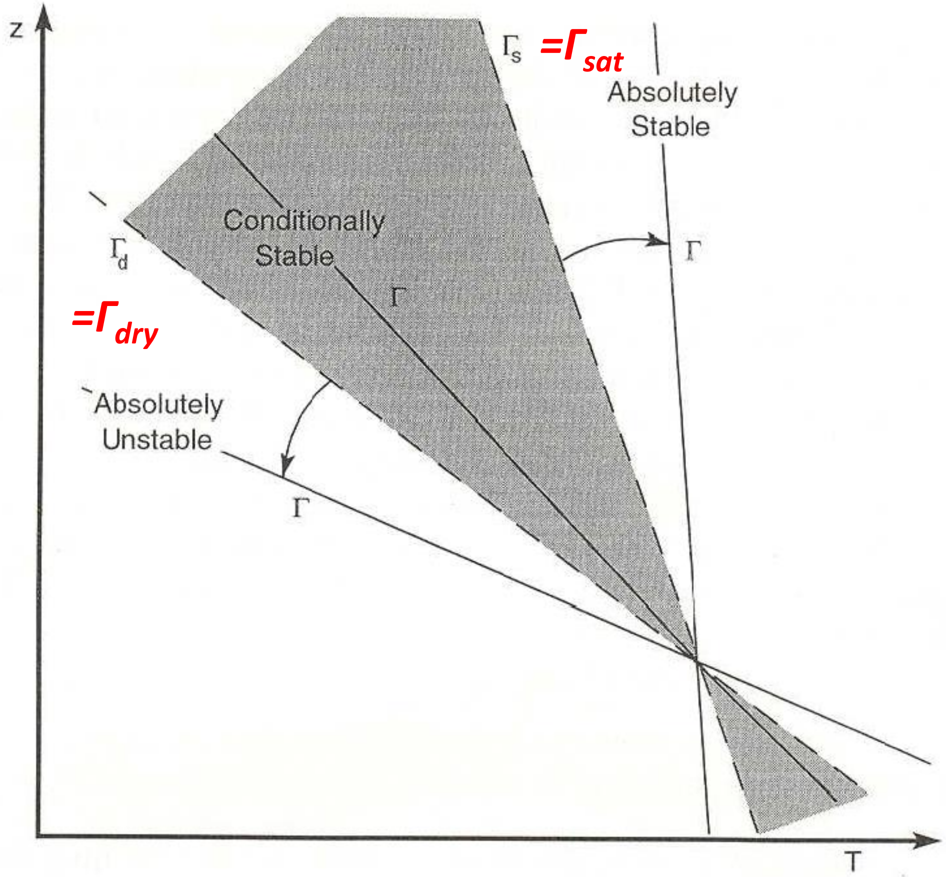
$$\left(C_p + L \frac{\partial r_{sat}}{\partial T}\right) = -g\left(1 - \rho L \frac{\partial r_{sat}}{\partial p}\right)\frac{dz}{dT}$$

$$\boxed{\Gamma_{sat} = \frac{dT}{dz_{sat}} = \Gamma_{dry} \frac{\left(1 - \rho L \frac{\partial r_{sat}}{\partial p}\right)}{\left(1 + \frac{L}{C_p}\frac{\partial r_{sat}}{\partial T}\right)}}$$

$\Gamma_{sat} > \Gamma_{dry}$

Conditional stability





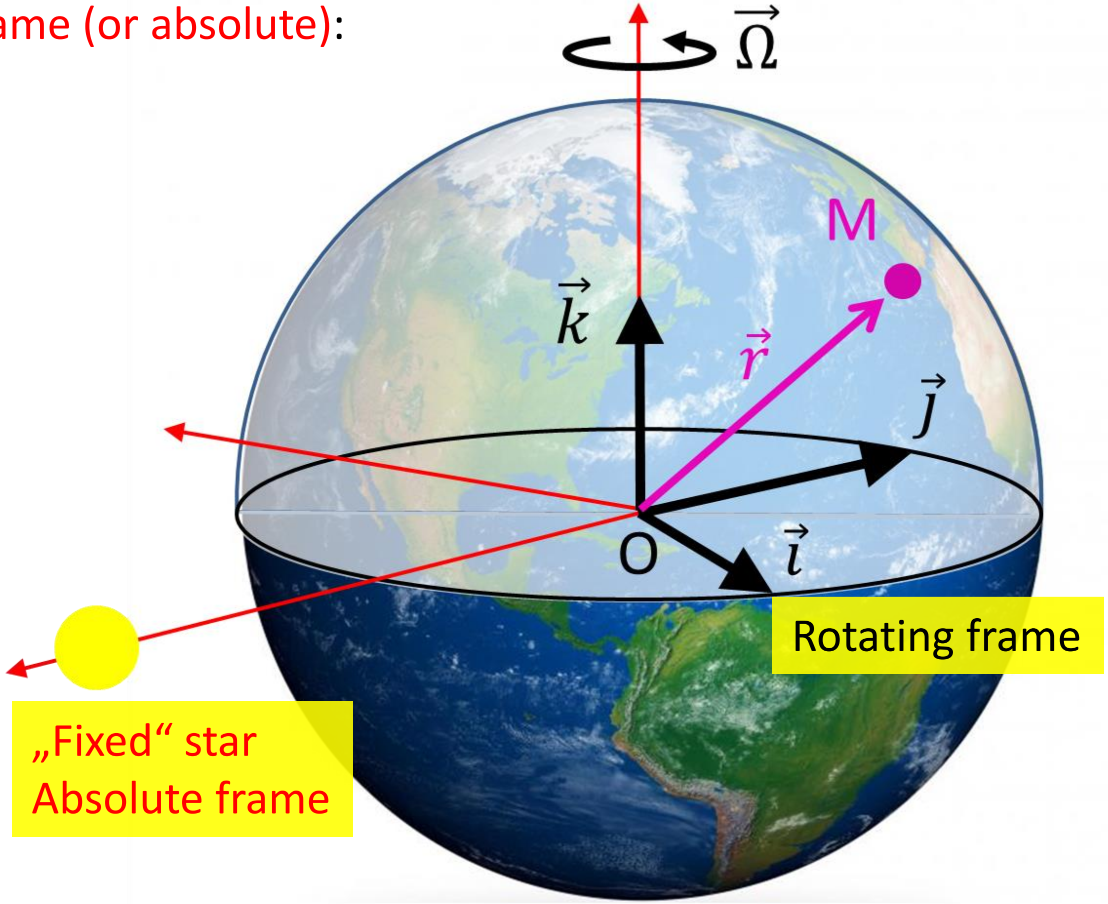
III. Atmospheric Circulation

Frames of reference

Newton‘s Second Law in a inertial frame (or absolute):

$$\overrightarrow{a_a} = \sum \frac{\vec{F}}{m}$$

We want to express this in a reference frame which rotates with the Earth



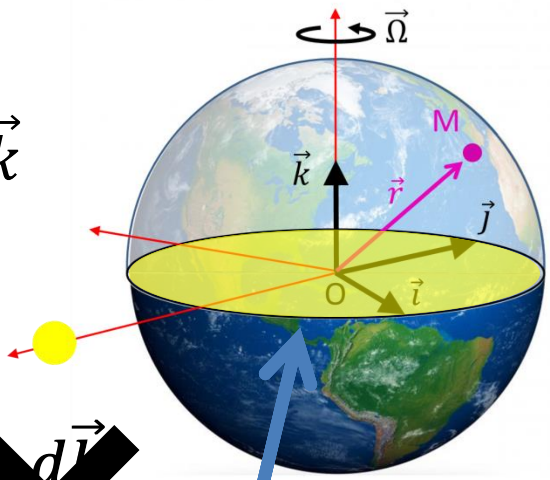
Rotating frame

„Fixed“ star
Absolute frame

Absolute acceleration

$$\overrightarrow{a_a} = \frac{D\overrightarrow{V_a}}{Dt} = \frac{D^2\vec{r}}{Dt^2} \quad \text{with} \quad \vec{r} = \overrightarrow{OM} = x\vec{\imath} + y\vec{\jmath} + z\vec{k}$$

(D/Dt to distinguish the total derivative in the absolute frame)



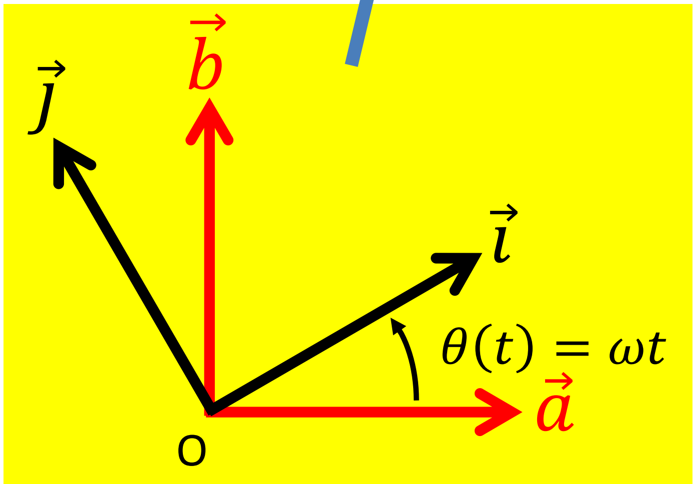
$$\overrightarrow{V_a} = \frac{D\vec{r}}{Dt} = \frac{dx}{dt}\vec{\imath} + x\frac{d\vec{\imath}}{dt} + \frac{dy}{dt}\vec{\jmath} + y\frac{d\vec{\jmath}}{dt} + \frac{dz}{dt}\vec{k} + \cancel{z\frac{d\vec{k}}{dt}}$$

$$\overrightarrow{V_a} = \overrightarrow{V_r} + \boxed{x\frac{d\vec{\imath}}{dt} + y\frac{d\vec{\jmath}}{dt}}$$

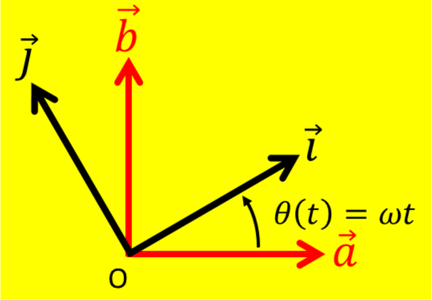
?

Absolute velocity

Relative velocity



Absolute acceleration (2)



$$\Rightarrow \begin{cases} \vec{i} = \cos\theta\,\vec{a} + \sin\theta\,\vec{b} \\ \vec{j} = -\sin\theta\,\vec{a} + \cos\theta\,\vec{b} \end{cases} \Rightarrow \begin{cases} d\vec{i}/dt = \omega\vec{j} \\ d\vec{j}/dt = -\omega\vec{i} \end{cases}$$

<u>We thus find</u>

$$\boxed{x\frac{d\vec{i}}{dt} + y\frac{d\vec{j}}{dt}} = x\omega\vec{j} - y\omega\vec{i}$$

<u>and</u> $\vec{\Omega} = \begin{pmatrix} 0 \\ 0 \\ \omega \end{pmatrix}, \vec{r} = \begin{pmatrix} x \\ y \\ z \end{pmatrix}$ <u>thus</u> $\boxed{\vec{\Omega} \times \vec{r}} = x\omega\vec{j} - y\omega\vec{i}$

<u>Finally:</u>

$$\overrightarrow{V_a} = \overrightarrow{V_r} + x\frac{d\vec{i}}{dt} + y\frac{d\vec{j}}{dt} = \overrightarrow{V_r} + \vec{\Omega} \times \vec{r}$$

Absolute velocity = Relative velocity + Entrainment velocity

Absolute acceleration (3)

$$\overrightarrow{a_a} = \frac{D\overrightarrow{V_a}}{Dt} = \frac{D\overrightarrow{V_r}}{Dt} + \frac{D}{Dt}\left(\vec{\Omega} \times \vec{r}\right) = \frac{D\overrightarrow{V_r}}{Dt} + \frac{D\vec{\Omega}}{Dt} \times \vec{r} + \vec{\Omega} \times \frac{D\vec{r}}{Dt}$$

$$\overrightarrow{a_a} = \overrightarrow{a_r} + \vec{\Omega} \times \overrightarrow{V_r} + D\vec{\Omega}/Dt \times \vec{r} + \vec{\Omega} \times \left(\overrightarrow{V_r} + \vec{\Omega} \times \vec{r}\right)$$

$$\overrightarrow{a_a} = \overrightarrow{a_r} + \underbrace{2\vec{\Omega} \times \overrightarrow{V_r}}_{\substack{\text{Coriolis} \\ \text{acceleration}}} + \underbrace{D\vec{\Omega}/Dt \times \vec{r}}_{\substack{\text{Euler} \\ \text{acceleration} \\ =0}} + \underbrace{\vec{\Omega} \times \left(\vec{\Omega} \times \vec{r}\right)}_{\substack{\text{Centripetal} \\ \text{acceleration}}} = \sum \vec{F}/m$$

In practice

(V_r=V)

$$\boxed{\frac{d\vec{V}}{dt} = -2\vec{\Omega} \times \vec{V} - \vec{\Omega} \times \left(\vec{\Omega} \times \vec{r}\right) + \sum \vec{F}/m}$$

Coriolis
„apparent“ force

Centrifugal
„apparent“ force

**Fundamental forces
???**

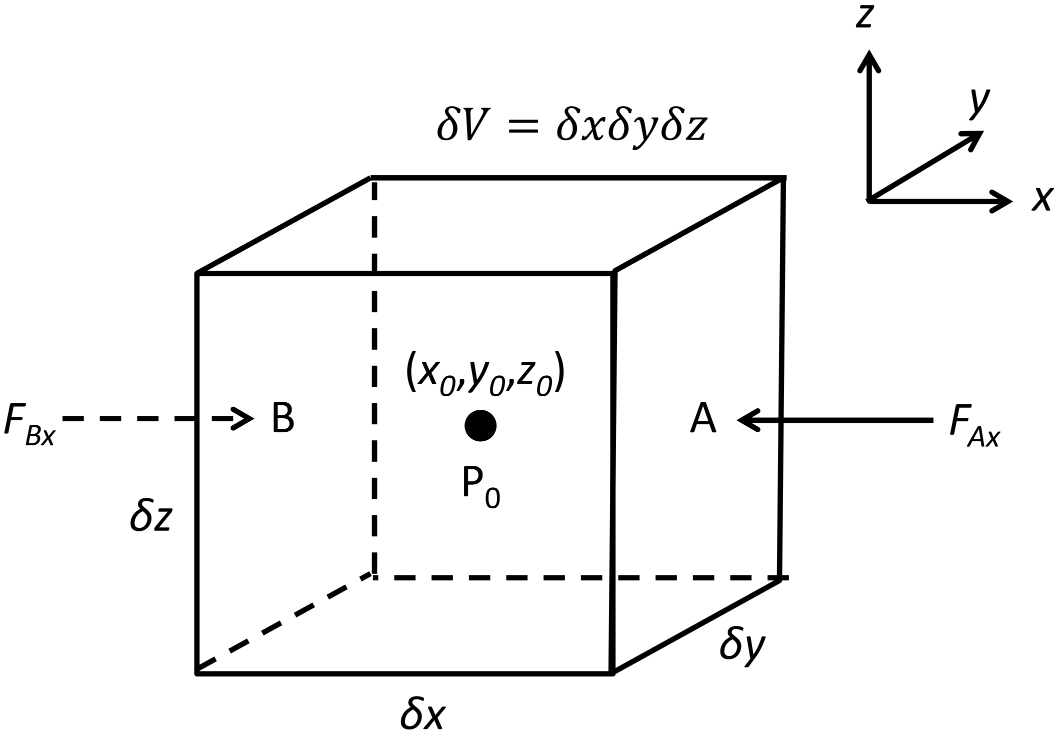
Fundamental forces: Pressure gradient

$$\begin{cases} p_A = p_0 + \dfrac{\partial p}{\partial x}\dfrac{\delta x}{2} \\ p_B = p_0 - \dfrac{\partial p}{\partial x}\dfrac{\delta x}{2} \end{cases}$$

$$d\vec{F} = p \cdot d\vec{S}$$

$$\begin{cases} F_{Ax} = -\left(p_0 + \dfrac{\partial p}{\partial x}\dfrac{\delta x}{2}\right)\delta y \delta z \\ F_{Bx} = +\left(p_0 - \dfrac{\partial p}{\partial x}\dfrac{\delta x}{2}\right)\delta y \delta z \end{cases}$$

$$\delta V = \delta x \delta y \delta z$$



$$m = \rho \delta V = \rho \delta x \delta y \delta z$$

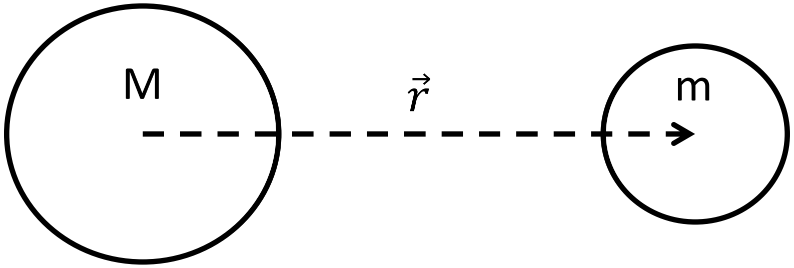
$$F_x = F_{Ax} + F_{Bx} = -\frac{\partial p}{\partial x}\delta x \delta y \delta z$$

$$\frac{F_x}{m} = -\frac{1}{\rho}\frac{\partial p}{\partial x}$$

$$\boxed{\frac{\vec{F}}{m} = -\frac{1}{\rho}\vec{\nabla} p}$$

Fundamental forces (2)

- Gravitational force & gravity



$$\vec{F_g} = -\frac{GMm}{r^2}\left(\frac{\vec{r}}{|\vec{r}|}\right)$$

$$\frac{\vec{F_g}}{m} \equiv \vec{g^*} = -\frac{GM}{r^2}\left(\frac{\vec{r}}{|\vec{r}|}\right)$$

a: Earth‘s radius
z: altitude
G: gravitational constant

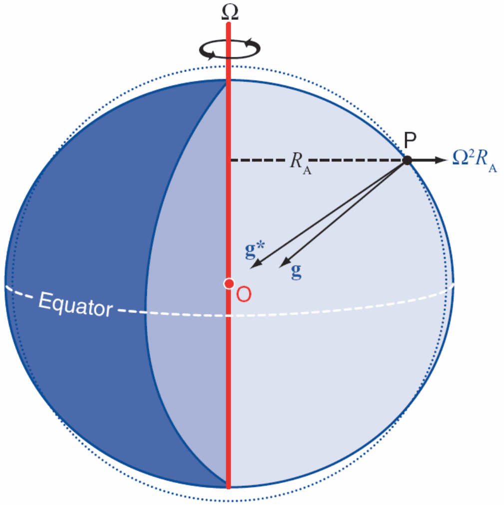
$$\vec{g^*} = -\frac{GM}{(a+z)^2}\left(\frac{\vec{r}}{|\vec{r}|}\right) = \frac{\vec{g_0^*}}{(1+z/a)^2}$$,with $$\vec{g_0^*} = -\frac{GM}{(a)^2}\left(\frac{\vec{r}}{|\vec{r}|}\right)$$

In practice: $\vec{g_0^*} = \vec{g^*}$

Gravity = Gravitational + Centrifugal forces

$$\boxed{\vec{g} \equiv \vec{g^*} + \Omega^2\vec{R_A} \approx 9.81\ m/s^2}$$

$$\mathbf{a} \times (\mathbf{b} \times \mathbf{c}) = (\mathbf{a} \cdot \mathbf{c})\mathbf{b} - (\mathbf{a} \cdot \mathbf{b})\mathbf{c}$$



- Friction

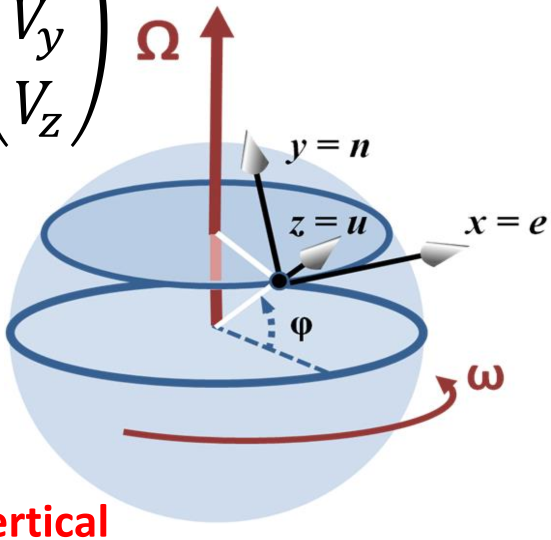
$\vec{F_f} = -\alpha\vec{V} \leftarrow$ Wind vector

Coriolis effect

$$\vec{a_c} = -2\vec{\Omega} \times \vec{V} \quad \text{with} \quad \vec{\Omega} = \omega \begin{pmatrix} 0 \\ \cos\varphi \\ \sin\varphi \end{pmatrix}, \vec{V} = \begin{pmatrix} V_x \\ V_y \\ V_z \end{pmatrix}$$

$$\Downarrow$$

$$\vec{a_c} = 2\omega \begin{pmatrix} V_y \sin\varphi - \xcancel{V_z \cos\varphi} \\ -V_x \sin\varphi \\ \xcancel{V_x \cos\varphi} \end{pmatrix}$$



In atmospheric dynamics, the **vertical velocity is small** and the **vertical component of the Coriolis acceleration is small compared to gravity**

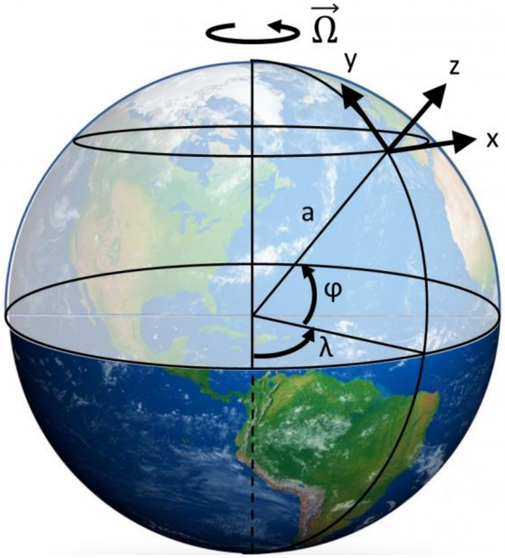
$$\boxed{\vec{a_c} = 2\omega \sin\varphi \begin{pmatrix} V_y \\ -V_x \end{pmatrix} = f \begin{pmatrix} V_y \\ -V_x \end{pmatrix} = -f\vec{k} \times \vec{V_h}}$$

Ballistic missile fired eastward at 43°N ($f=10^{-4}\ s^{-1}$). If the missile travels 1000 km at $V_{x0}=1000$ m/s, how much is the missile deflected ?

$$\frac{dV_y}{dt} = -fV_{x0} \xrightarrow{\int} V_y = -fV_{x0}t \xrightarrow{\int} \delta y = -\frac{fV_{x0}t^2}{2} = -50km \text{ (southward)}$$

Momentum equation: Spherical coordinates (λ,φ,z)

$$\boxed{\frac{d\overrightarrow{V}}{dt} = \vec{g} - 2\overrightarrow{\Omega} \times \overrightarrow{V} - \frac{1}{\rho}\vec{\nabla} p - \overrightarrow{F_f}}$$



We note: $\vec{V} = u\vec{\imath} + v\vec{\jmath} + w\vec{k}$

In spherical coordinate, some **additional curvature terms** appear

$$\begin{cases} \frac{du}{dt} - \frac{uv\tan\varphi}{a} + \frac{uw}{a} = -\frac{1}{\rho}\frac{\partial p}{\partial x} + 2\Omega v\sin\varphi - 2\Omega w\cos\varphi + F_{fx} & \text{(x)} \\ \frac{dv}{dt} - \frac{u^2\tan\varphi}{a} + \frac{vw}{a} = -\frac{1}{\rho}\frac{\partial p}{\partial y} - 2\Omega u\sin\varphi + F_{fy} & \text{(y)} \\ \frac{dw}{dt} - \frac{u^2+v^2}{a} = -\frac{1}{\rho}\frac{\partial p}{\partial z} - g + 2\Omega u\cos\varphi + F_{fz} & \text{(z)} \end{cases}$$

! Very difficult to handle since nonlinear...

Momentum equation: horizontal scale analysis

... however, not all terms are important, it depends on the scale that is considered !

For synoptic scale at mid-latitudes (φ=45°):

- U~10m/s, W~1cm/s, L~10^6m, D~10^4m, $\Delta_H P/\rho$~$10^3 \mathrm{m^2/s^2}$, $T=L/U$~10^5s, a~6400 km.
- $2\Omega\sin(\varphi)$~ $2\Omega\cos(\varphi)$~ $10^{-4}\mathrm{s}^{-1}$

Thus, considering the x&y-equation (and negleting friction):

(x) $$\frac{du}{dt} - \frac{uv\tan\varphi}{a} + \frac{uw}{a} = -\frac{1}{\rho}\frac{\partial p}{\partial x} + 2\Omega v\sin\varphi - 2\Omega w\cos\varphi$$

(y) $$\frac{dv}{dt} - \frac{u^2\tan\varphi}{a} + \frac{vw}{a} = -\frac{1}{\rho}\frac{\partial p}{\partial y} - 2\Omega u\sin\varphi$$

(scale)	$\frac{U^2}{L}$	$\frac{U^2}{a}$	$\frac{UW}{a}$	$\frac{\Delta p}{\rho L}$	$f_0 U$	$f_0 W$
($\mathrm{m/s^2}$)	10^{-4}	10^{-5}	10^{-8}	10^{-3}	10^{-3}	10^{-6}

Coriolis force and pressure gradient force in approximate balance ⬄ geostrophic approximation

Geostrophic balance

(x)-geostrophic $\frac{1}{\rho}\frac{\partial p}{\partial x} = 2\Omega v_g \sin\varphi$

(y)-geostrophic $\frac{1}{\rho}\frac{\partial p}{\partial y} = -2\Omega u_g \sin\varphi$

$$\boxed{v_g = \frac{1}{\rho f}\frac{\partial p}{\partial x} \quad u_g = -\frac{1}{\rho f}\frac{\partial p}{\partial y}}$$

$f = 2\Omega \sin\varphi$

Coriolis parameter

~ 80% of motion can be explained by the geostrophic approximation

General expression: $-2\vec{\Omega} \times \vec{V_g} - \frac{1}{\rho}\overrightarrow{\nabla_h p} = 0$

$$f\vec{k} \times \vec{V_g} = -\frac{1}{\rho}\overrightarrow{\nabla_h p}$$

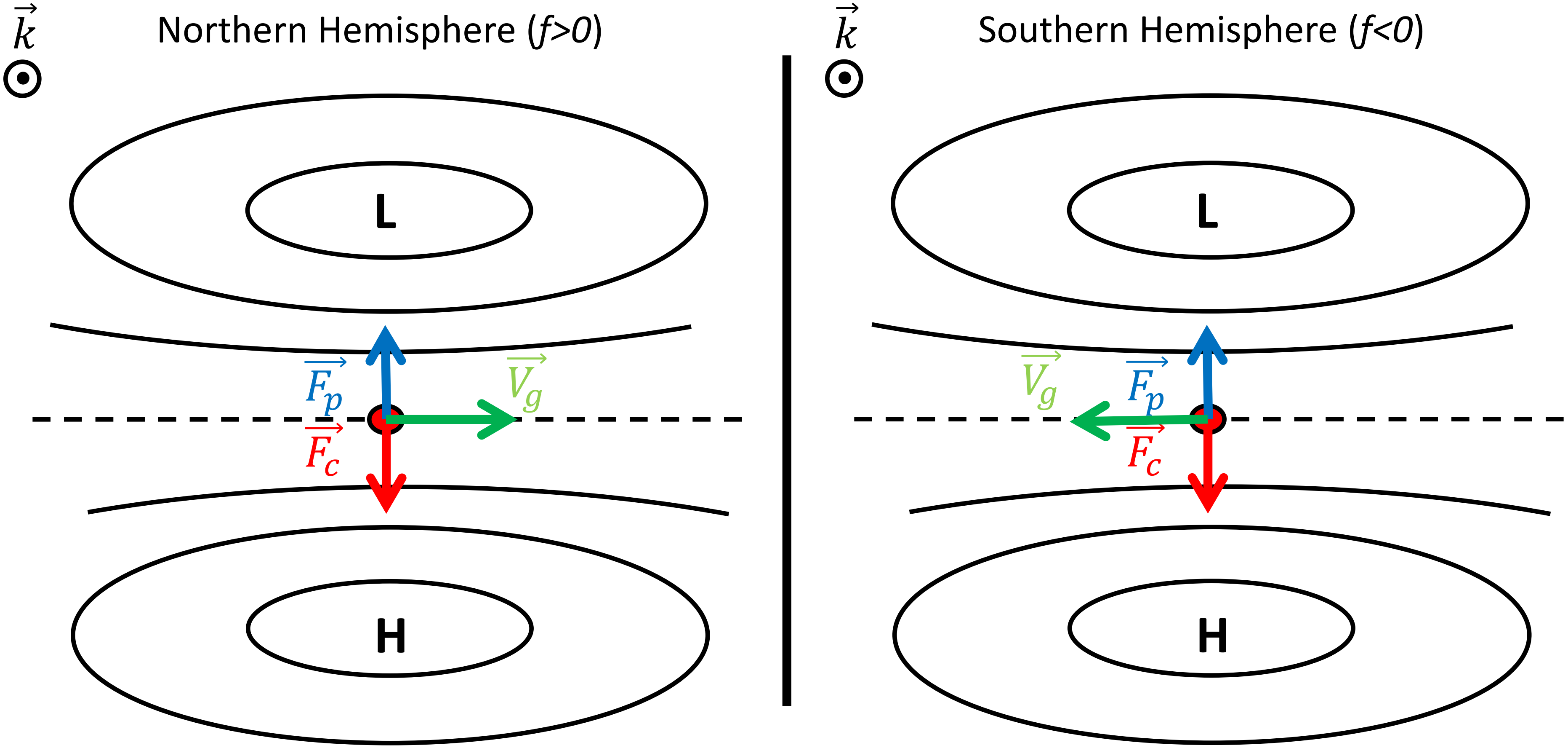
$\times \vec{k}$ *and* $(\vec{A} \times \vec{B}) \times \vec{C} = (\vec{A} \cdot \vec{C}) \cdot \vec{B} - (\vec{B} \cdot \vec{C}) \cdot \vec{A}$

$$\boxed{\overrightarrow{V_g} = \frac{1}{\rho f}\vec{k} \times \overrightarrow{\nabla_h p}}$$

The geostrophic wind blows along the isobares

Buys-Ballot law

$$\overrightarrow{V_g} = \frac{1}{\rho f}\vec{k} \times \overrightarrow{\nabla_h}p$$



"In the Northern Hemisphere, if a person stands with his back to the wind, the atmospheric pressure is low to the left, high to the right"

Validity of geostrophic approximate

To obtain prediction equation, acceleration term has to be retained

$$\begin{cases} \frac{du}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial x} + fv = f(v - v_g) \\ \frac{dv}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial y} - fu = -f(u - u_g) \end{cases}$$

Acceleration = Pressure gradient + Coriolis force

Validity of the geostrophic approximation: Acceleration/Coriolis ⇔ Rossby number

R_0 small
Significant Coriolis effect

$$R_0 = \frac{U^2/_L}{f_0 U} = \frac{U}{f_0 L}$$

R_0 big
No Coriolis effect

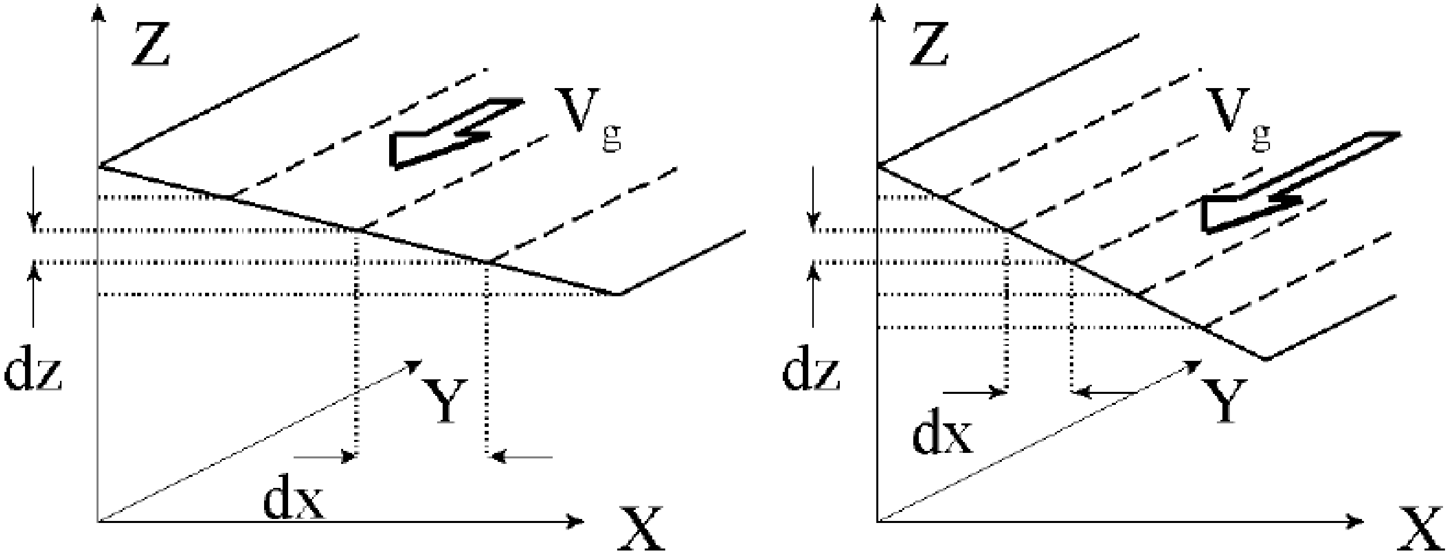
The smallness of the Rossby number is a measure of the validity of the geostrophic approximation

Thermal wind

Starting from geostrophic wind: $u_g = -\frac{1}{\rho f}\left(\frac{\partial p}{\partial y}\right)_z$ $v_g = \frac{1}{\rho f}\left(\frac{\partial p}{\partial x}\right)_z$

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

On a constant pressure surface: $u_g = -\frac{g}{f}\left(\frac{\partial z}{\partial y}\right)_p$ $v_g = \frac{g}{f}\left(\frac{\partial z}{\partial x}\right)_p$



The magnitude of the geostrophic wind is determined by the tilt of pressure surface

Thermal wind

By differentiating to pressure:

$$\frac{\partial}{\partial p}u_g = \frac{\partial}{\partial p}\left(-\frac{g}{f}\frac{\partial z}{\partial y}\right) = -\frac{g}{f}\frac{\partial}{\partial y}\left(\frac{\partial z}{\partial p}\right)$$

$$\times p \qquad \frac{\partial z}{\partial p} = -\frac{1}{\rho g} = -\frac{1}{g}\frac{RT}{p}$$

$$p\frac{\partial}{\partial p}u_g = \frac{\partial u_g}{\partial \ln(p)} = -\frac{R}{f}\left(\frac{\partial T}{\partial y}\right)_p \text{,and similarly } p\frac{\partial}{\partial p}v_g = \frac{\partial v_g}{\partial \ln(p)} = \frac{R}{f}\left(\frac{\partial T}{\partial x}\right)_p$$

Vectorial form:

$$\boxed{\frac{\partial \overrightarrow{V_g}}{\partial \ln(p)} = -\frac{R}{f}\vec{k} \times \overrightarrow{\nabla_p}T}$$

Vertical variation of geostrophic wind depends on temperature horizontal gradient

By integrating from p_0 to p_1 ($p_0 > p_1$) we obtain the thermal wind equation:

$$\boxed{u_T = -\frac{R}{f}\left(\frac{\partial \langle T\rangle}{\partial y}\right)_p \ln\left(\frac{p_0}{p_1}\right)} \quad \text{and} \quad \boxed{v_T = \frac{R}{f}\left(\frac{\partial \langle T\rangle}{\partial x}\right)_p \ln\left(\frac{p_0}{p_1}\right)}$$

<*T*>: mean temperature in the [p_0,p_1] layer

! The thermal wind equation is a relationship for the vertical wind shear !

Thermal wind: application to the jet stream

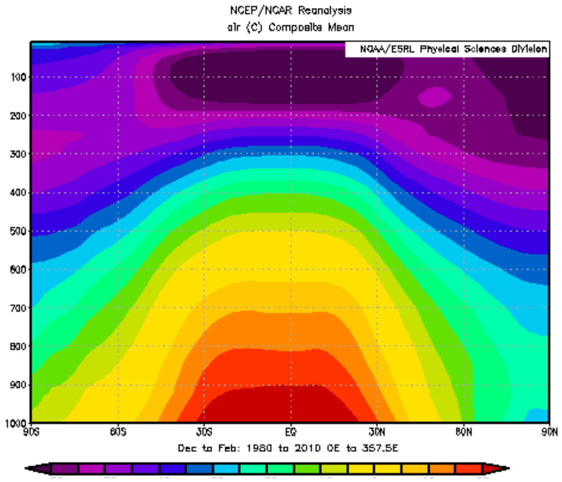
$$u_T = -\frac{R}{f}\left(\frac{\partial\langle T\rangle}{\partial y}\right)_p \ln(\frac{p_0}{p_1})$$

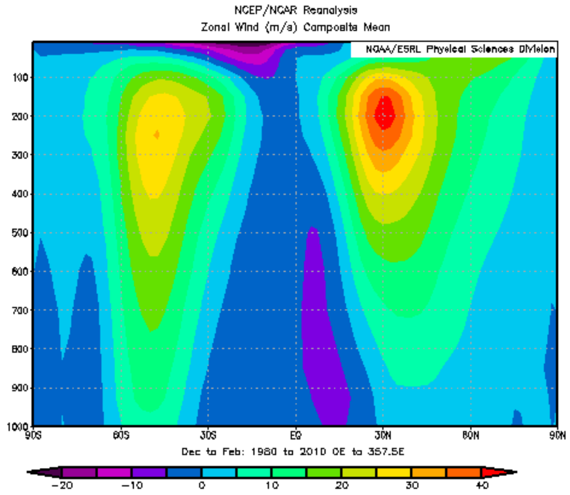
NCEP reanalysis: 1980-2000

Temperature

Zonal wind

DJF



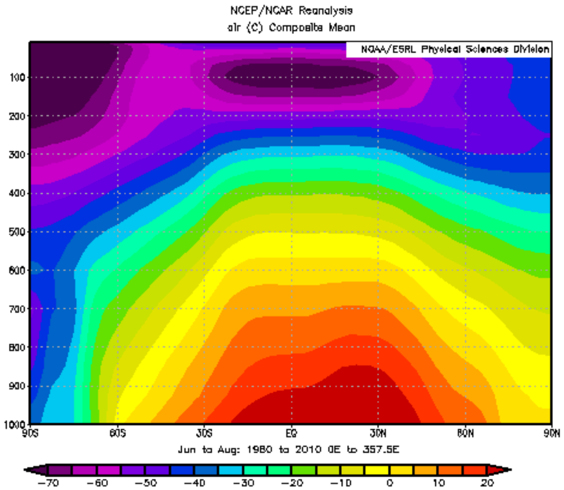


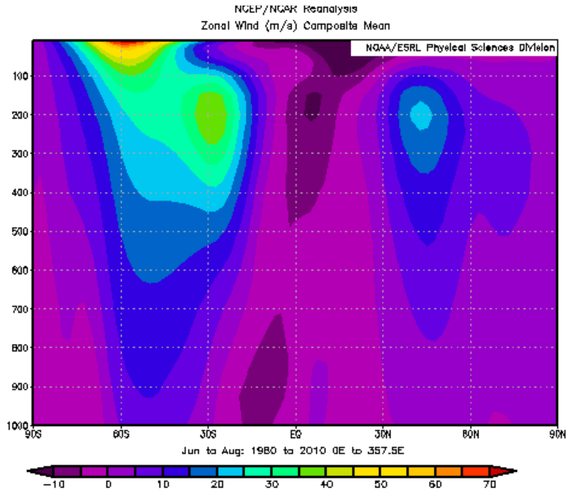
pressure

easterlies

westerlies

JJA



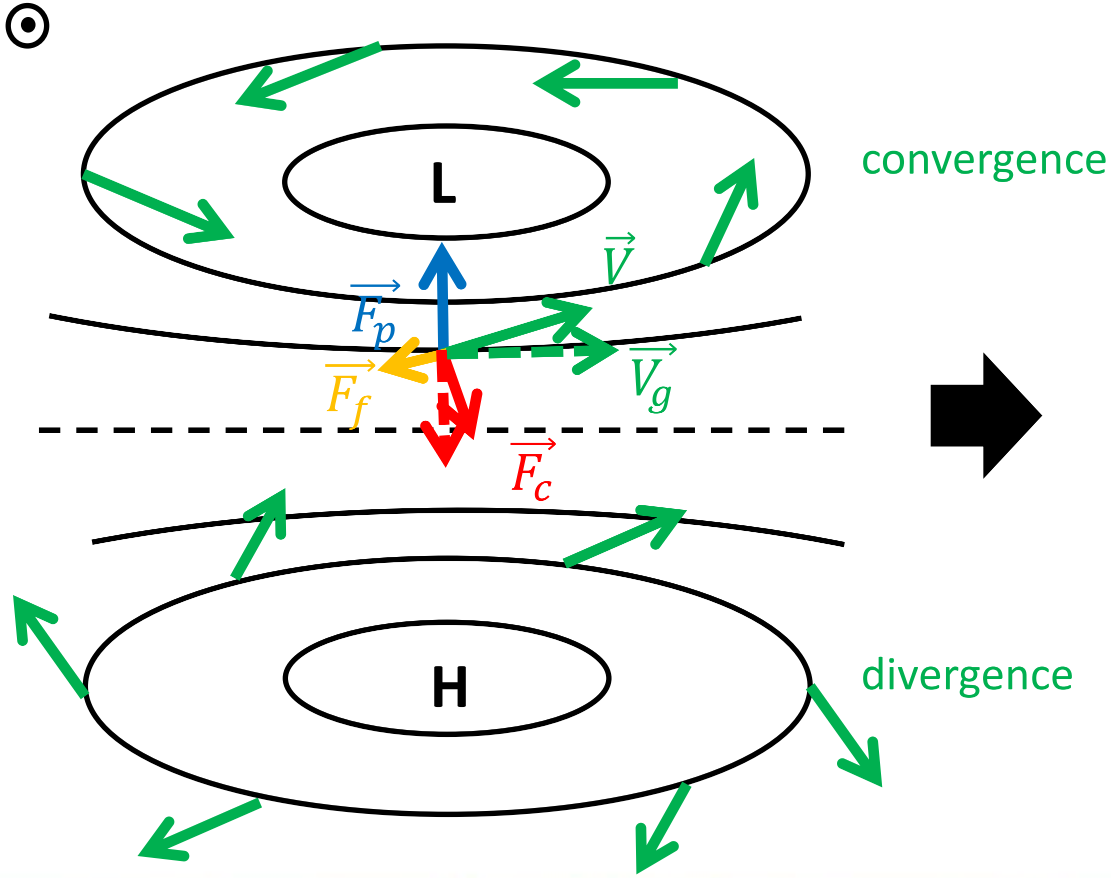


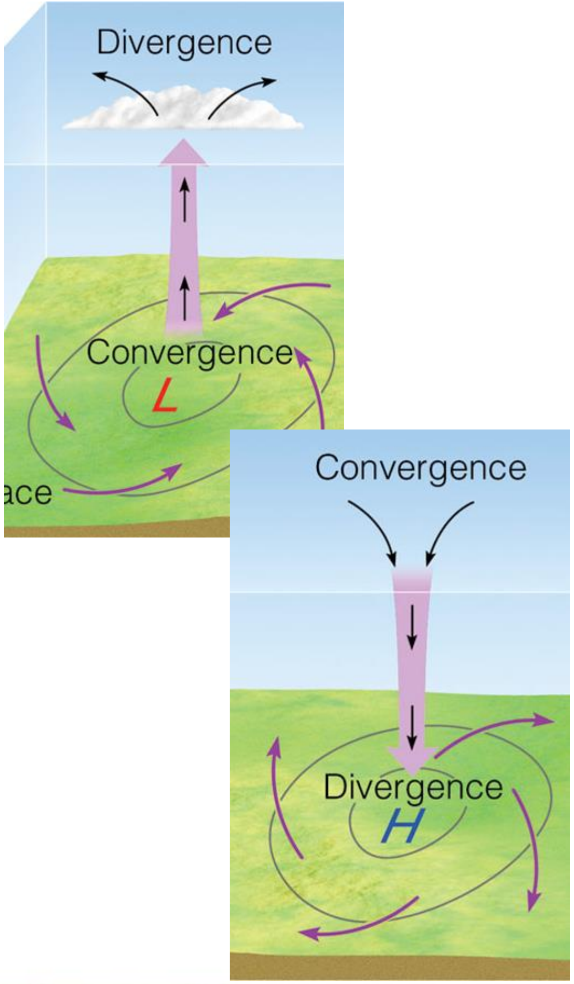
pressure

Influence of friction: boundary layer

$$f\vec{k}\times\vec{V}+\vec{F_f}=-\frac{1}{\rho}\overrightarrow{\nabla_h}p \quad \text{,with} \quad \vec{F_f}=-\alpha\vec{V}$$

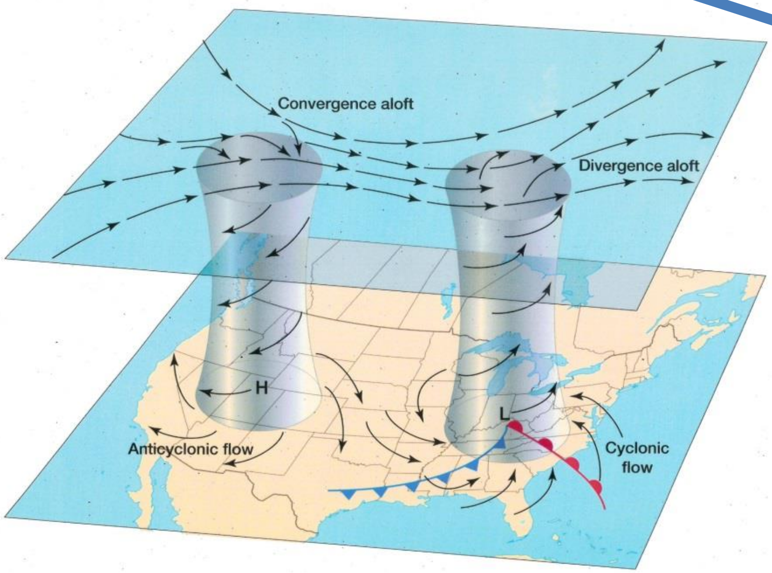
$\vec{k}$ Northern Hemisphere (f>0)

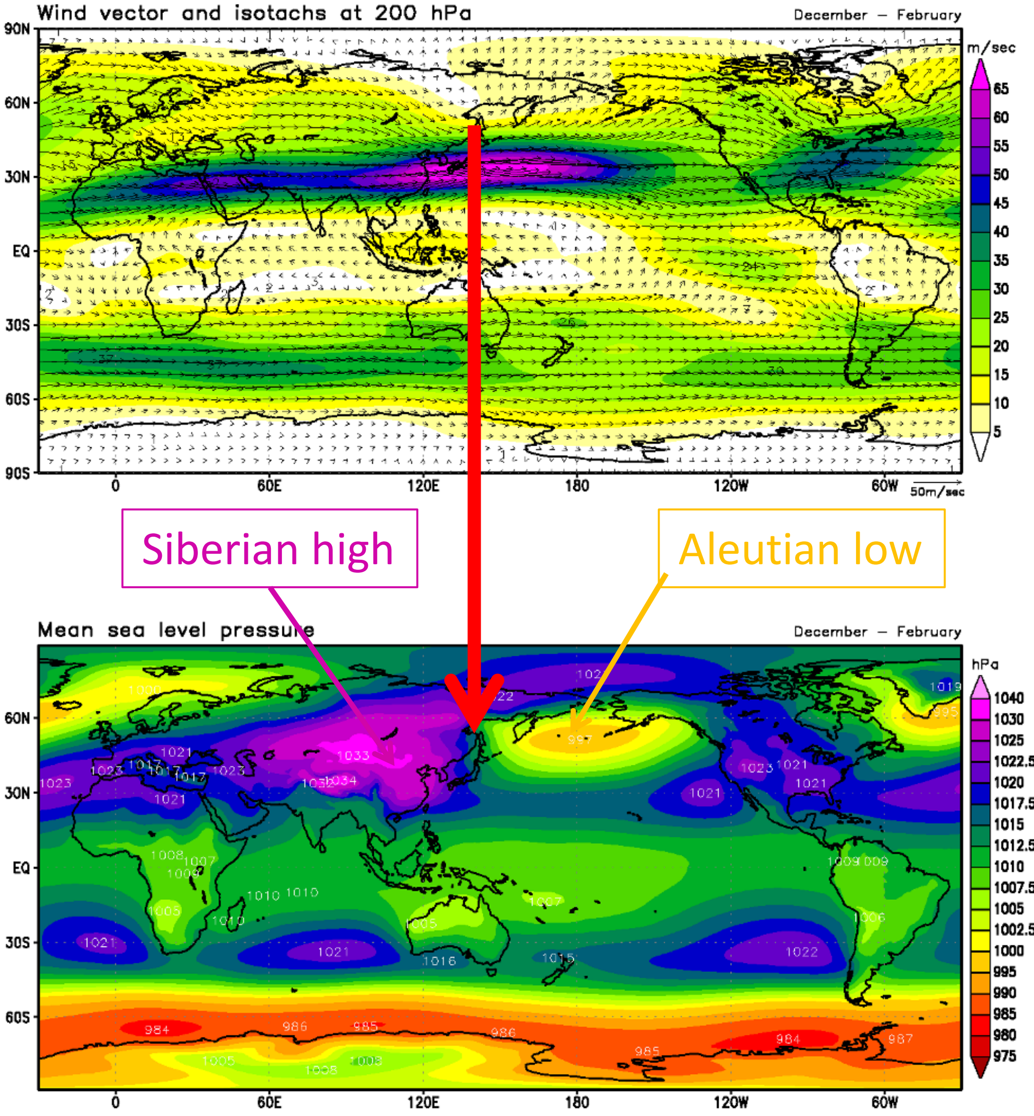




A concrete example

Subtropical jetstream





Siberian high

Aleutian low

To characterize the atmospheric motions we can use also divergence and vorticity of the wind

The vorticity equation

Horizontal momentum equation:

$$\frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + w\frac{\partial u}{\partial z} - fv = -\frac{1}{\rho}\frac{\partial p}{\partial x} \qquad (1)$$

$$\frac{\partial v}{\partial t} + u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + w\frac{\partial v}{\partial z} + fu = -\frac{1}{\rho}\frac{\partial p}{\partial y} \qquad (2)$$

$$\frac{\partial (2)}{\partial x} - \frac{\partial (1)}{\partial y}$$

$$\frac{\partial \zeta}{\partial t} + u\frac{\partial \zeta}{\partial x} + v\frac{\partial \zeta}{\partial y} + w\frac{\partial \zeta}{\partial z} + (\zeta + f)\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) + \left(\frac{\partial w}{\partial x}\frac{\partial v}{\partial z} - \frac{\partial w}{\partial y}\frac{\partial u}{\partial z}\right) + v\frac{df}{dy}$$

$$= \frac{1}{\rho^2}\left(\frac{\partial \rho}{\partial x}\frac{\partial p}{\partial y} - \frac{\partial \rho}{\partial y}\frac{\partial p}{\partial x}\right)$$

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

$$\frac{df}{dt} = v\frac{df}{dy}$$ **: f depends only on y**

$$\frac{d(\zeta + f)}{dt} = -(\zeta + f)\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) - \left(\frac{\partial w}{\partial x}\frac{\partial v}{\partial z} - \frac{\partial w}{\partial y}\frac{\partial u}{\partial z}\right) - \frac{1}{\rho^2}\left(\frac{\partial \rho}{\partial x}\frac{\partial p}{\partial y} - \frac{\partial \rho}{\partial y}\frac{\partial p}{\partial x}\right)$$

ζ+f = vorticity + planetary vorticity = absolute vorticity

Vorticity equation : scale analysis

For synoptic scale motion at midlatitudes:

- U~10m/s, W~1cm/s, L~10^6m, H~10^4m, δp~10 hPa, ρ~1 kg m^{-3}, $\delta\rho/\rho$~10^{-2}.
- $f_0=10^{-4}s^{-1}$

$$\frac{d(\zeta+f)}{dt} = -(\zeta+f)\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right) - \xcancel{\left(\frac{\partial w}{\partial x}\frac{\partial v}{\partial z}-\frac{\partial w}{\partial y}\frac{\partial u}{\partial z}\right)} - \frac{1}{\rho^2}\xcancel{\left(\frac{\partial \rho}{\partial x}\frac{\partial p}{\partial y}-\frac{\partial \rho}{\partial y}\frac{\partial p}{\partial x}\right)}$$

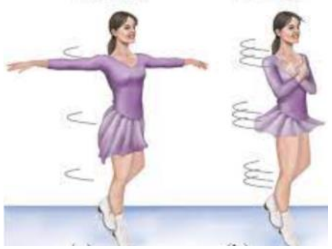
$f_0U/L=10^{-9}\ s^{-2}$ $WU/(HL)=10^{-11}\ s^{-2}$ $\delta p\delta\rho/(\rho^2L^2)=10^{-11}\ s^{-2}$

$$\boxed{\frac{d_h(\zeta+f)}{dt} = -(\zeta+f)\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)}$$

Link between the vorticity and the divergence of the flow

same phenomenon as an ice skating turning on herself

- spreading her arms => decreases her rotating speed
-tightening his arms => increases her rotating speed



Potential voticity

We can further show:

$$PV = (\zeta_\theta + f)\left(-g\frac{\partial\theta}{\partial p}\right) = const$$

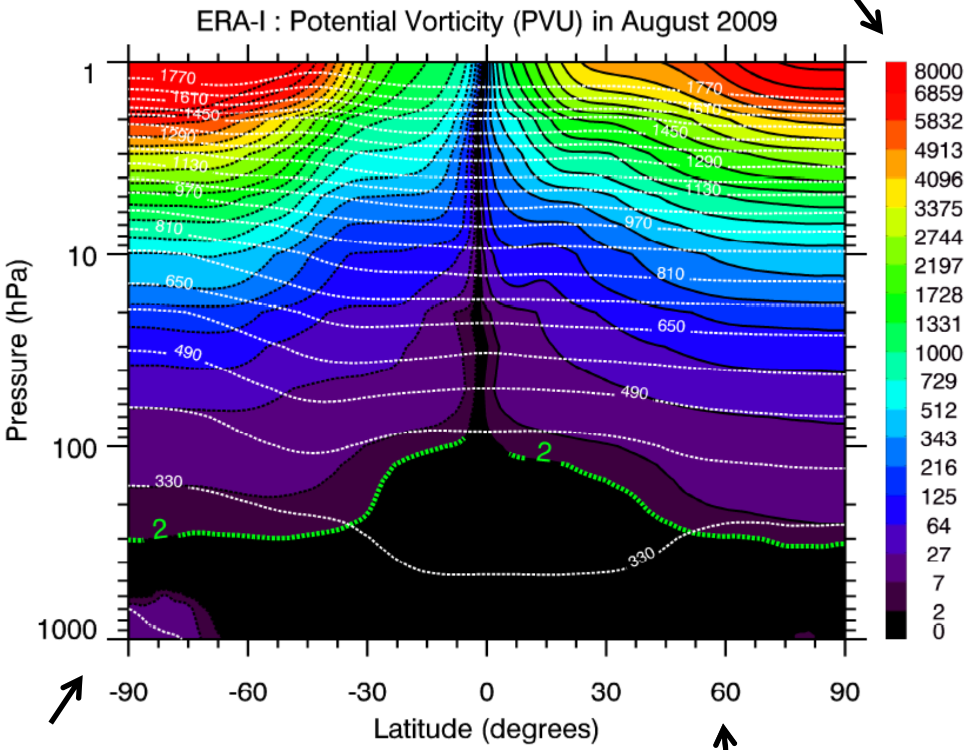
Absolute vorticity

Vertical stability

The potential vorticity is conserved under adiabatic and frictionless motions

- f increases with latitude: meridional gradient
- θ increases with altitude: vertical gradient

Absolute PV values



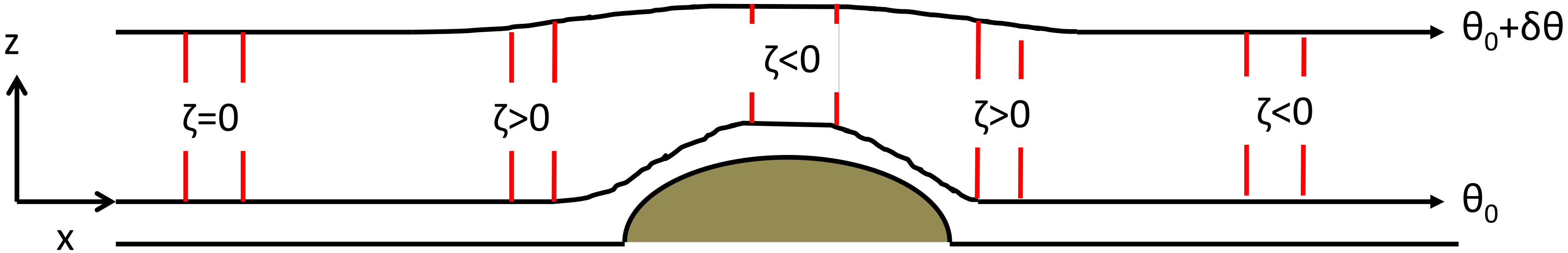
Negative in SH (f<0)

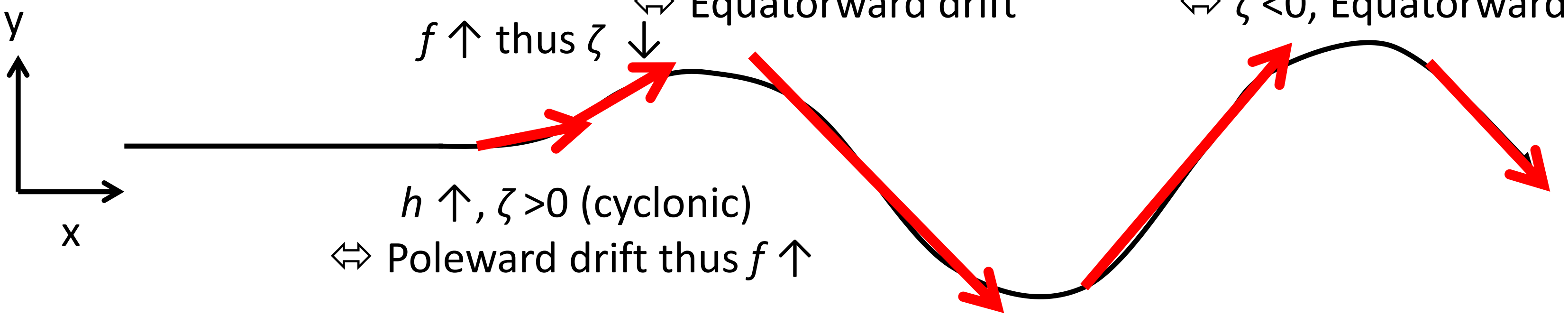
Positive in NH (f>0)

Potential vorticity conservation: planetary waves (1)

Westerly flow

$$PV = (\zeta_\theta + f)\overset{=1/h}{\left(-g\frac{\partial\theta}{\partial p}\right)} = const$$





h ↓, ζ <0 (anticyclonic)
⬄ Equatorward drift

f ↑ thus ζ ↓

f higher than initially
⬄ ζ <0, Equatorward drift

h ↑, ζ >0 (cyclonic)
⬄ Poleward drift thus f ↑

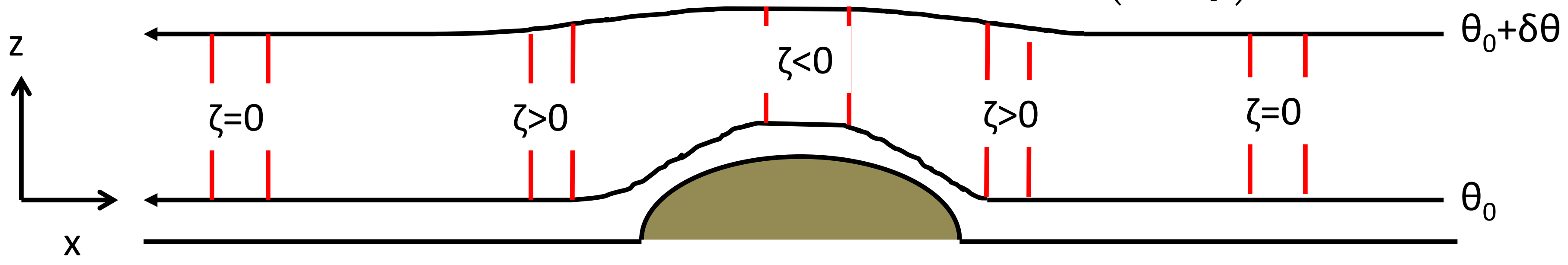
f smaller than initially
⬄ ζ >0, poleward drift

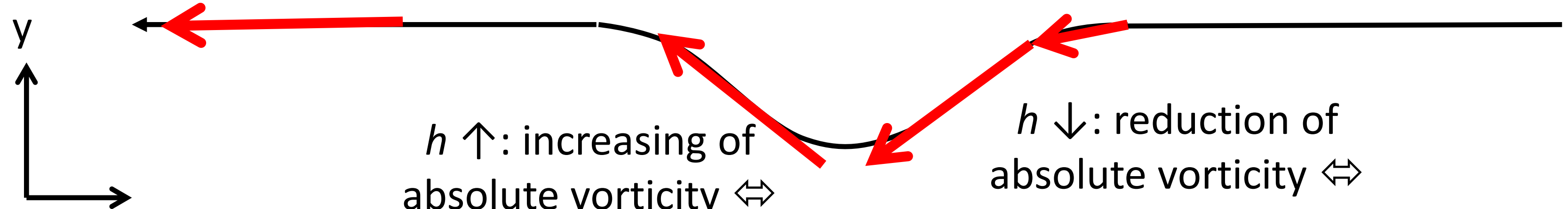
This mechanism generates planetary waves under westerly circulation

Potential vorticity conservation: planetary waves (2)

Easterly flow

$$PV = (\zeta_\theta + f)\overset{=1/h}{\left(-g\frac{\partial\theta}{\partial p}\right)} = const$$





h ↑, *ζ* >0 (cyclonic)
⇔ Equatorward drift *f* ↓

Return to initial state

h ↑: increasing of absolute vorticity ⇔ cyclonic & *f* ↑

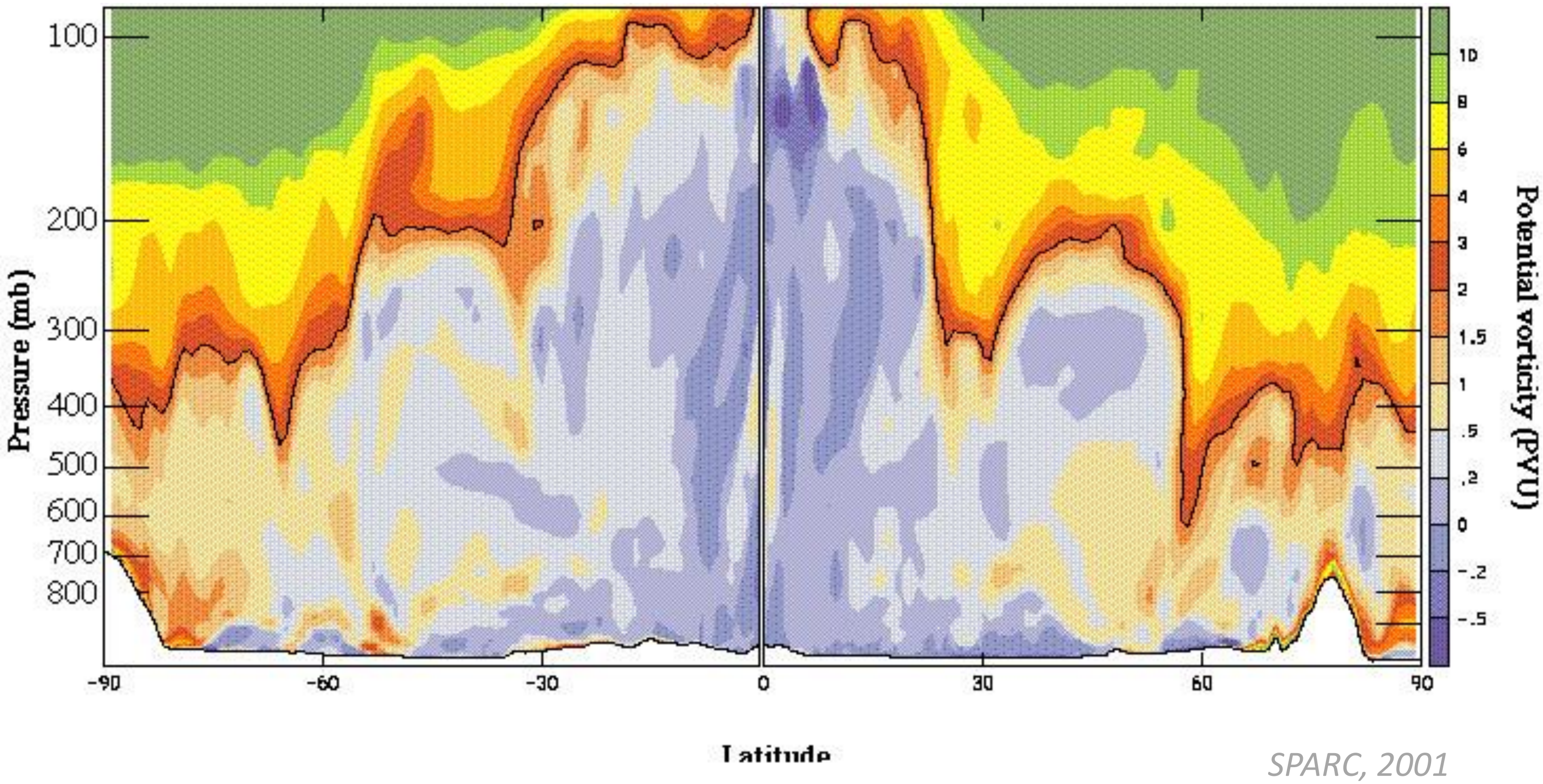
h ↓: reduction of absolute vorticity ⇔ anticyclonic & *f* ↓

No planetary wave generation under easterly flow

Potential vorticity: dynamical tropopause definition

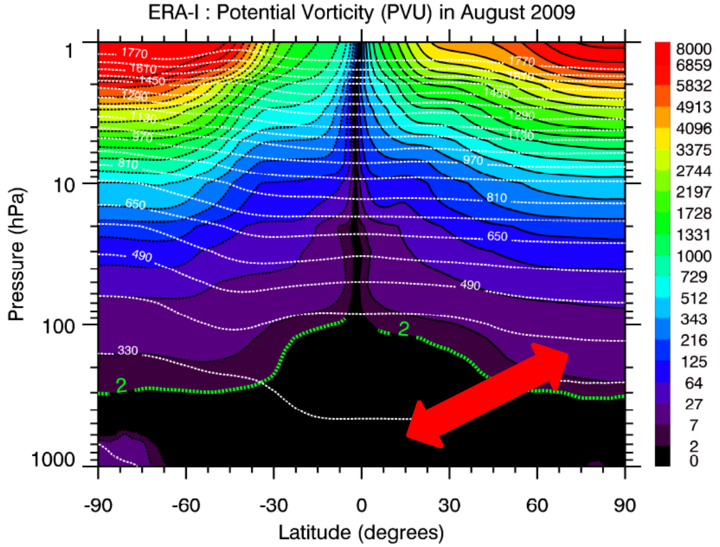
$$PV = (\zeta_\theta + f)\left(-g\frac{\partial\theta}{\partial p}\right) = const$$

Tropopause definition as PV surface of 2 pvu (1 pvu =10^{-6} K m^2 kg^{-1} s^{-1}).). Use of the increase statistic stability at the tropopause.



SPARC, 2001

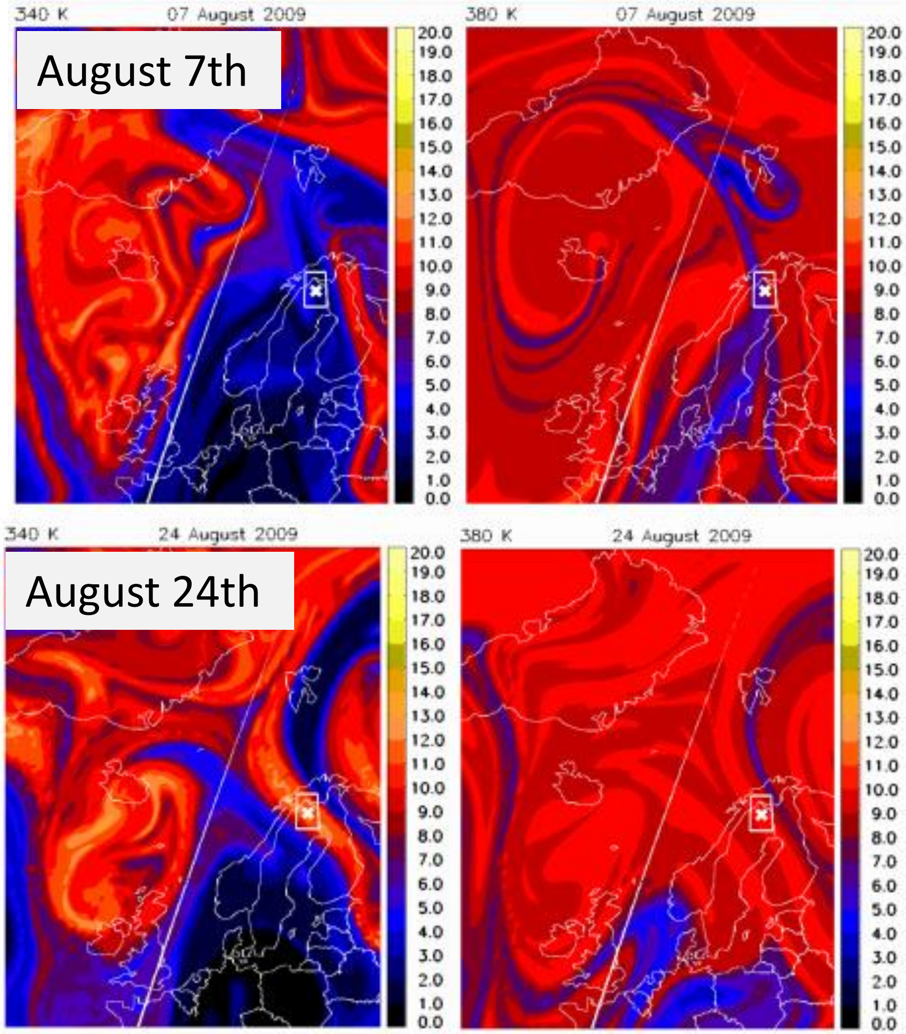
Stratosphere/troposphere exchanges



Isentropic transport

Baloon campaign in polar region

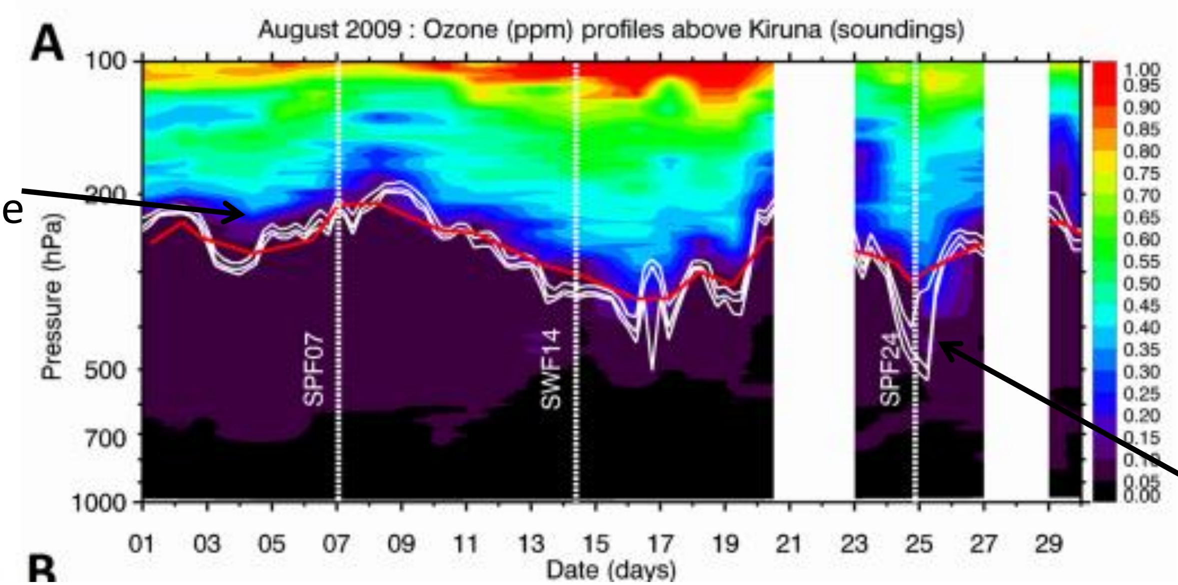
Dynamical conditions (PV)



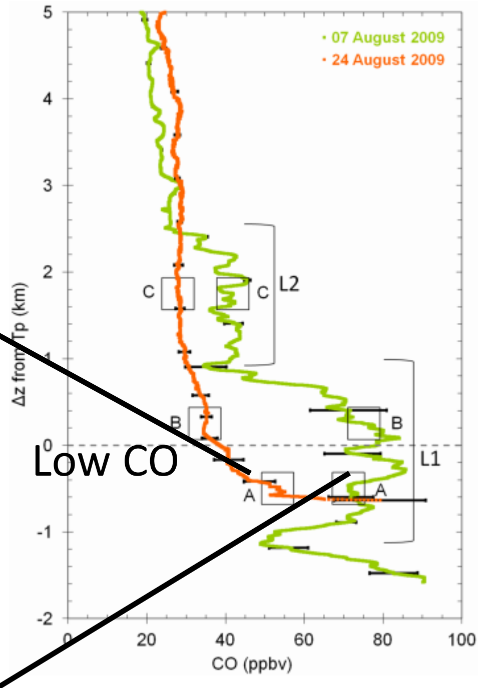
Stratosphere/troposphere exchanges

Ozone (ppm)

Dynamical tropopause



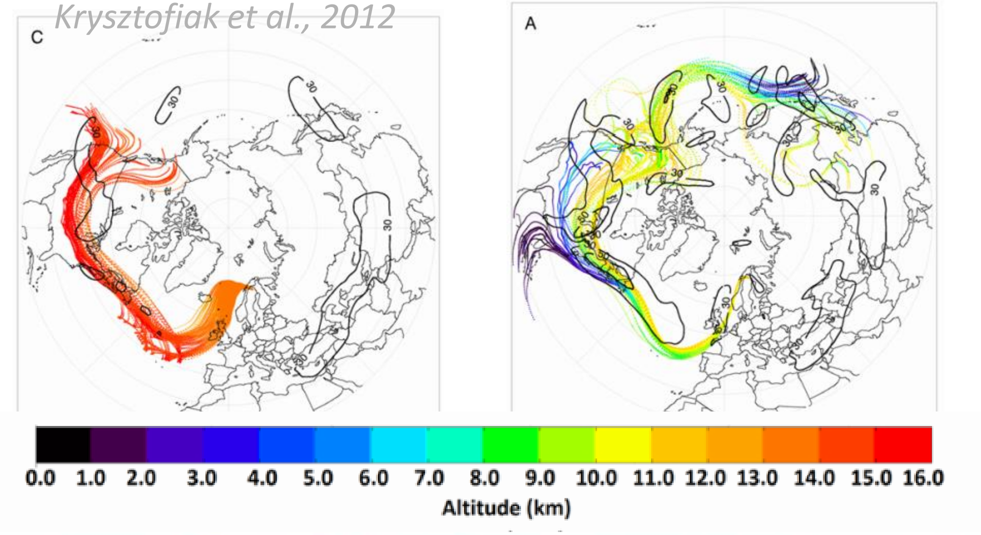
CO profile



Low CO

High CO

Krysztofiak et al., 2012



10-days Backward trajectories computed from FLEXTRA

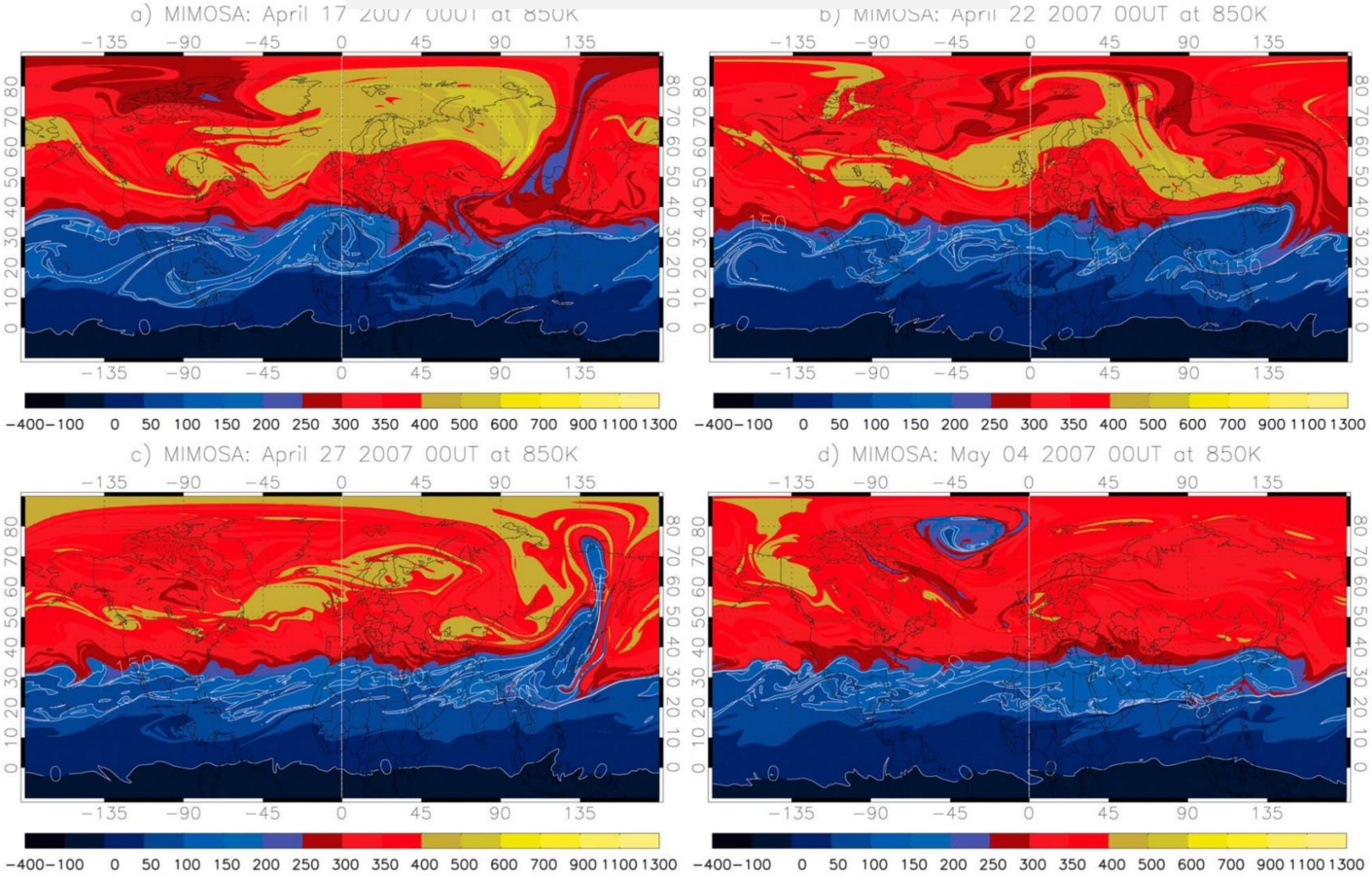
(Kiruna, summer 2009)

Krystofiak et al. (2012)

Stratospheric horizontal transport

~30 km

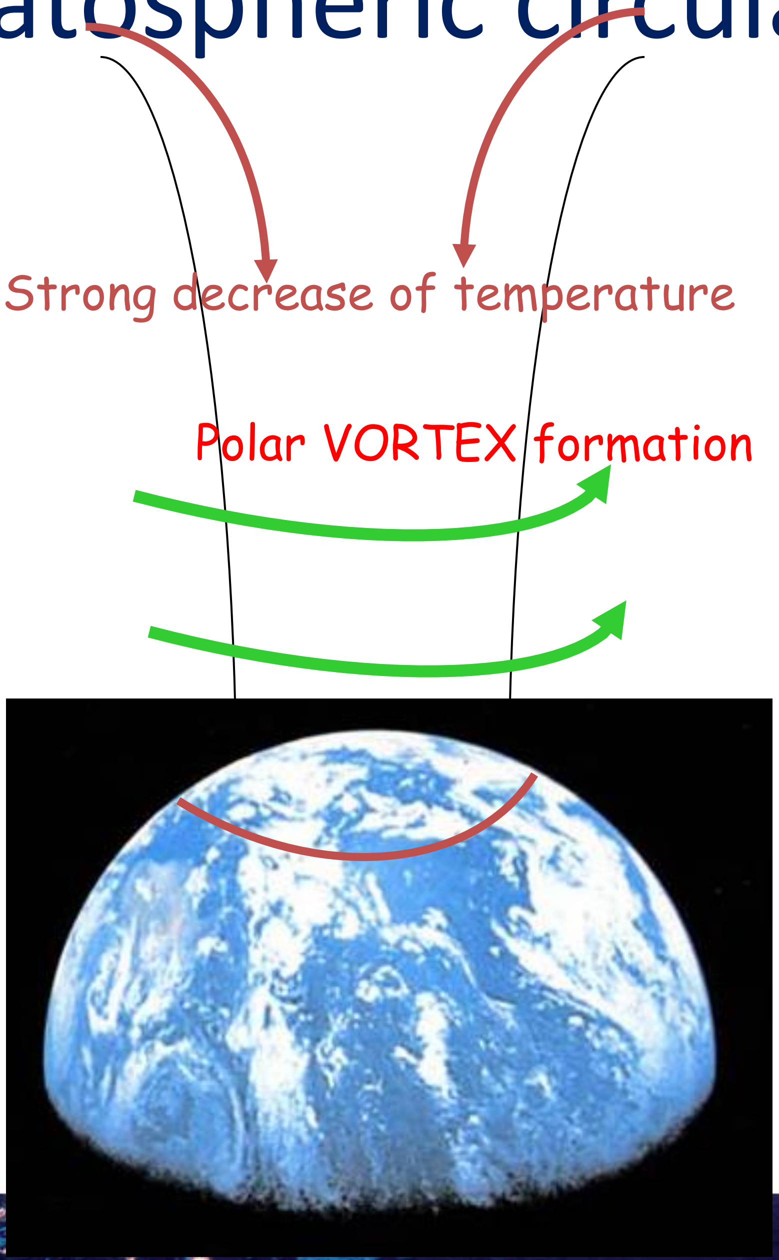
Potential vorticity maps (MIMOSA)



V. Stratosphere

Stratospheric circulation

During polar winter



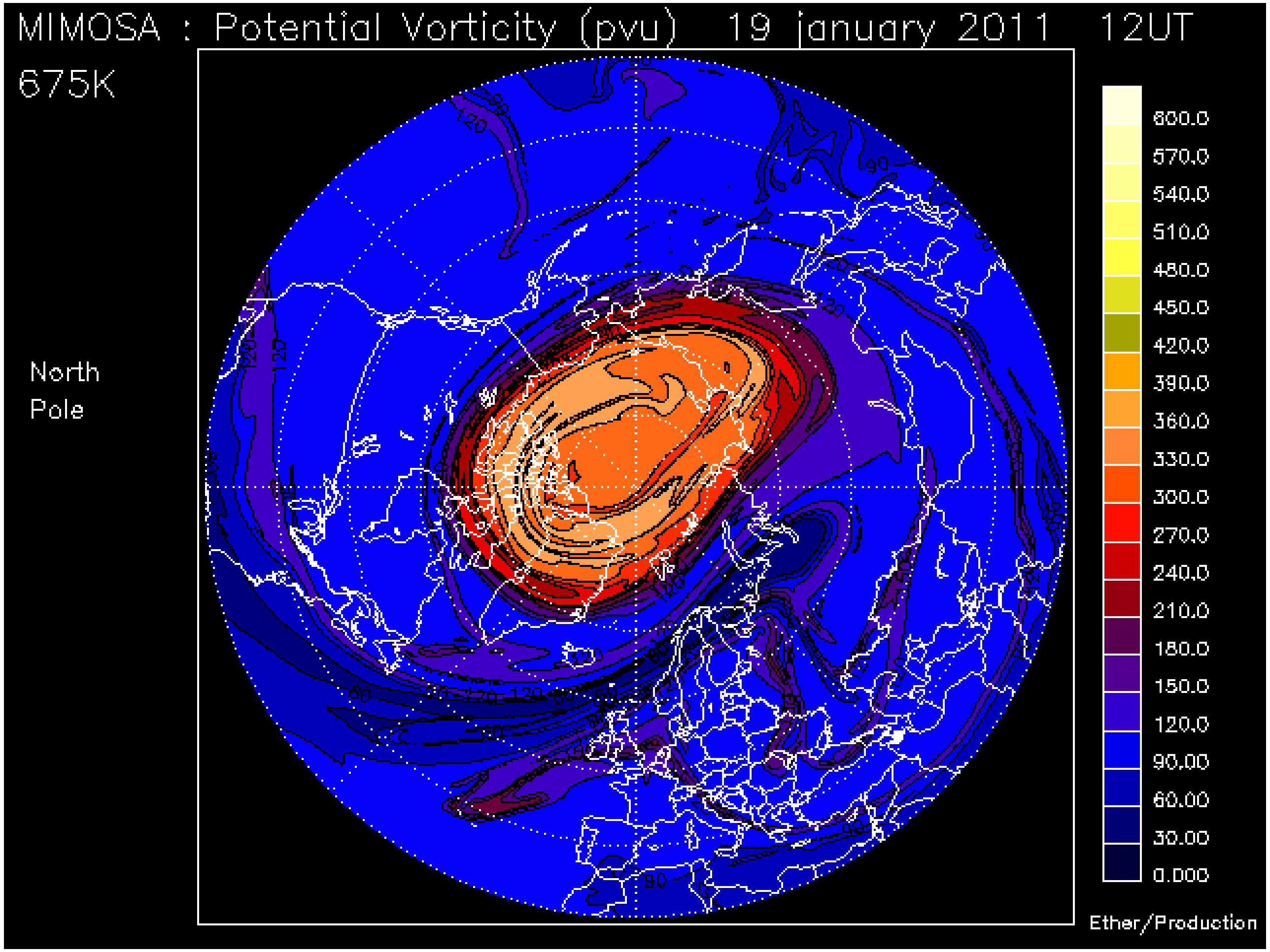
Strong decrease of temperature

Polar VORTEX formation

Strong wind
100 m/s

Polar vortex

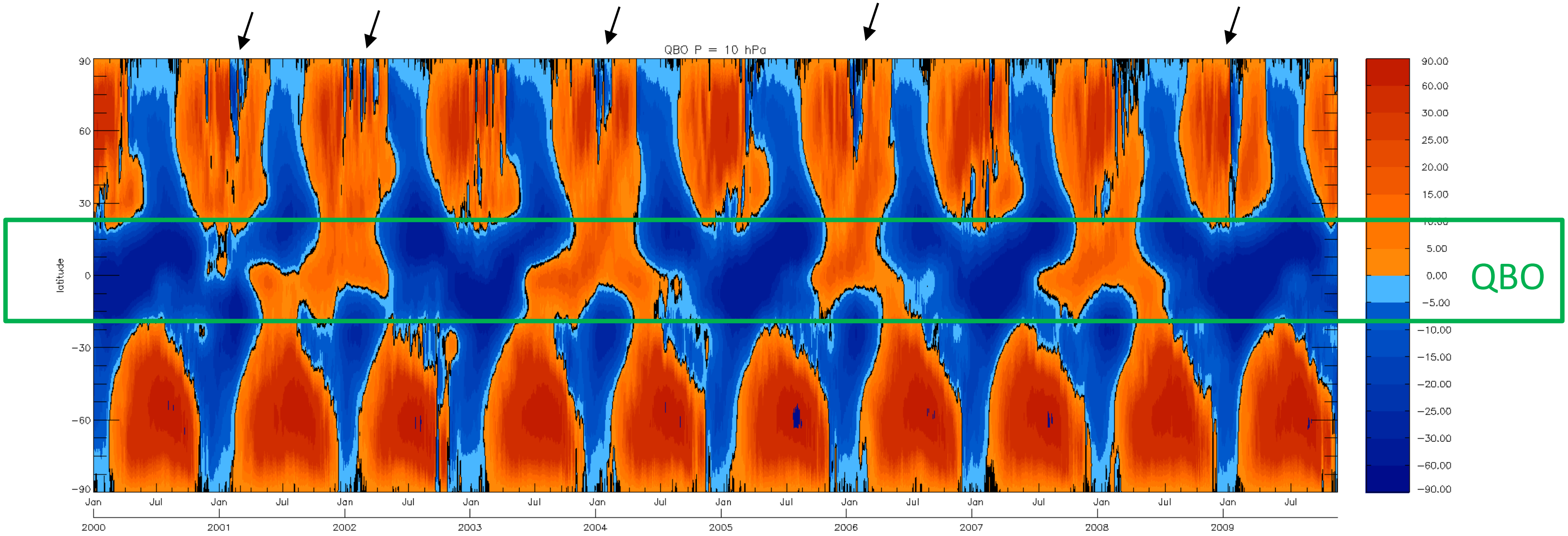
- Potential vorticity map Forecasts on AERIS Database



MIMOSA model hauchecorne et al. (2001)

Zonal circulation cycles at 10 hPa

Sudden Stratospheric Warming –>polar vortex displacement or break up

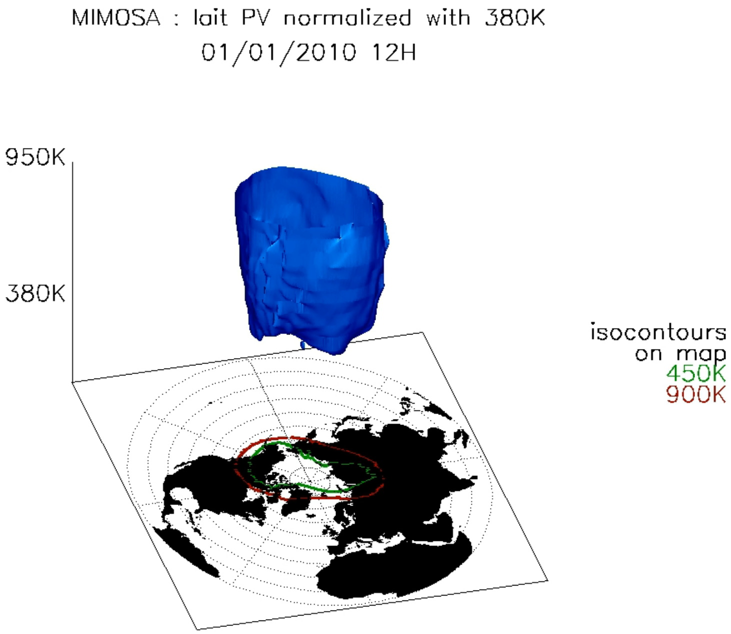


Polar vortex (Estearlies)

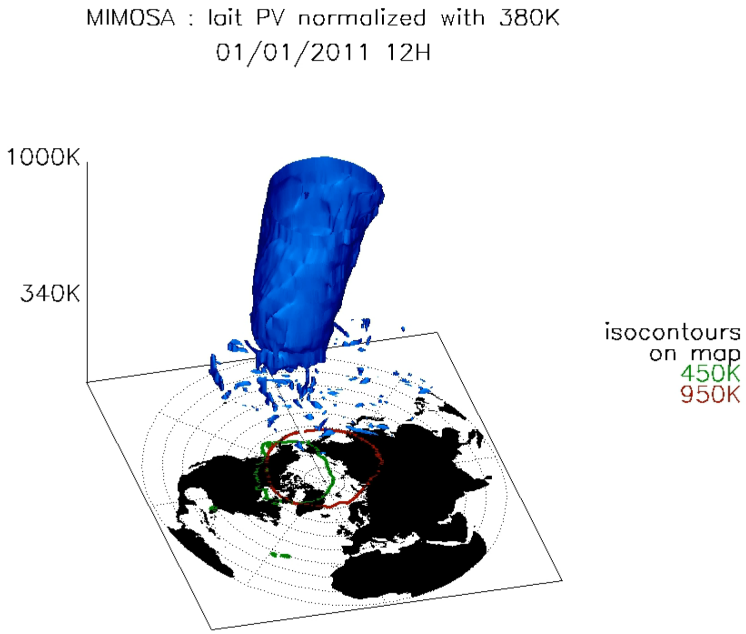
Summer polar Anticyclone (Westerlies)

Equatorial Quasi Biennal Oscillation -QBO

Polar vortex as a function of altitude



Vortex break-up in two lobes
-> Sudden Stratospheric Warming



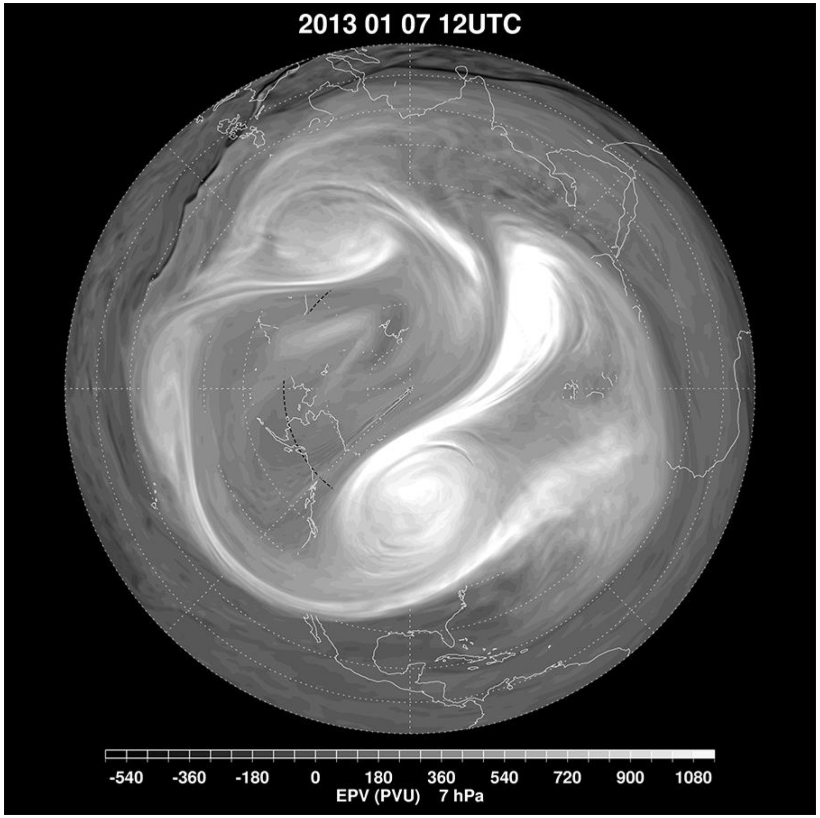
Unusual stable polar vortex
without major SSW

Arctic Sudden Warming of January 2013

- the vortex distorting and breaking up, coincident with the increase in polar temperatures at 10 hPa.

- By January 7, 2013, the shearing of the original polar vortex by the winds led to the existence of three smaller, interconnected vortices (over Canada, northern Eurasia, northeastern Siberia..)

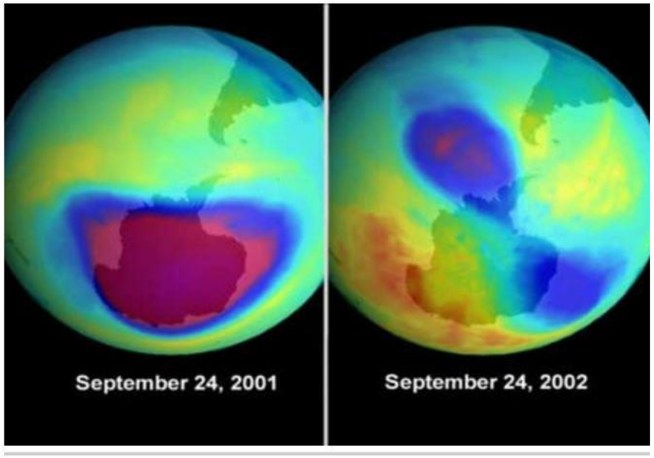
Lawrence Coy, Steven Pawson

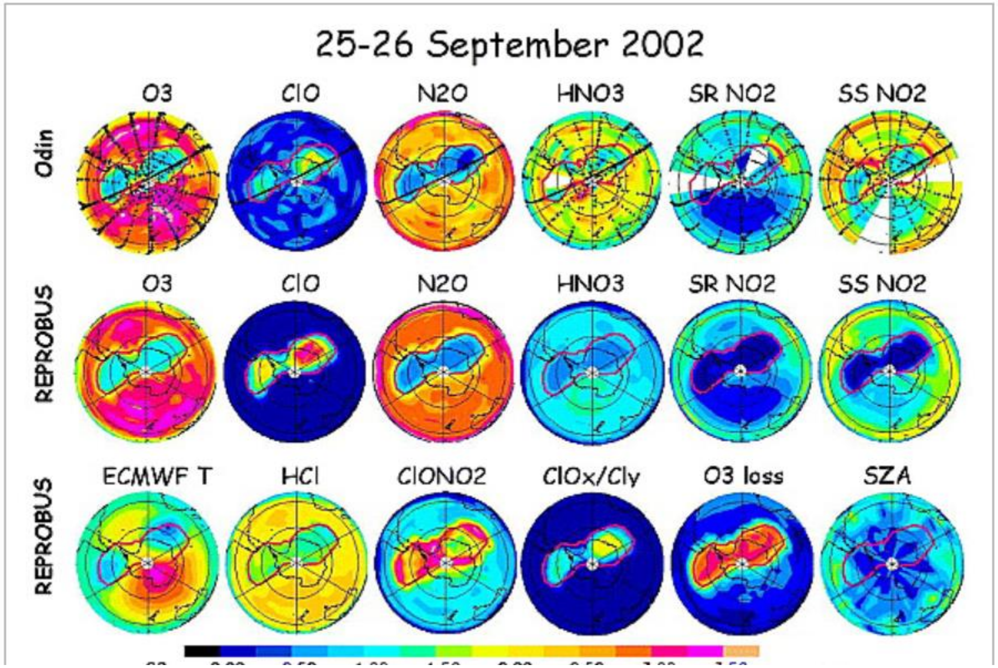


GEOS-5 Analyses

Antarctic polar vortex Split in 2002

In September 2002 the Antarctic polar vortex split in two
under the influence of a sudden warming



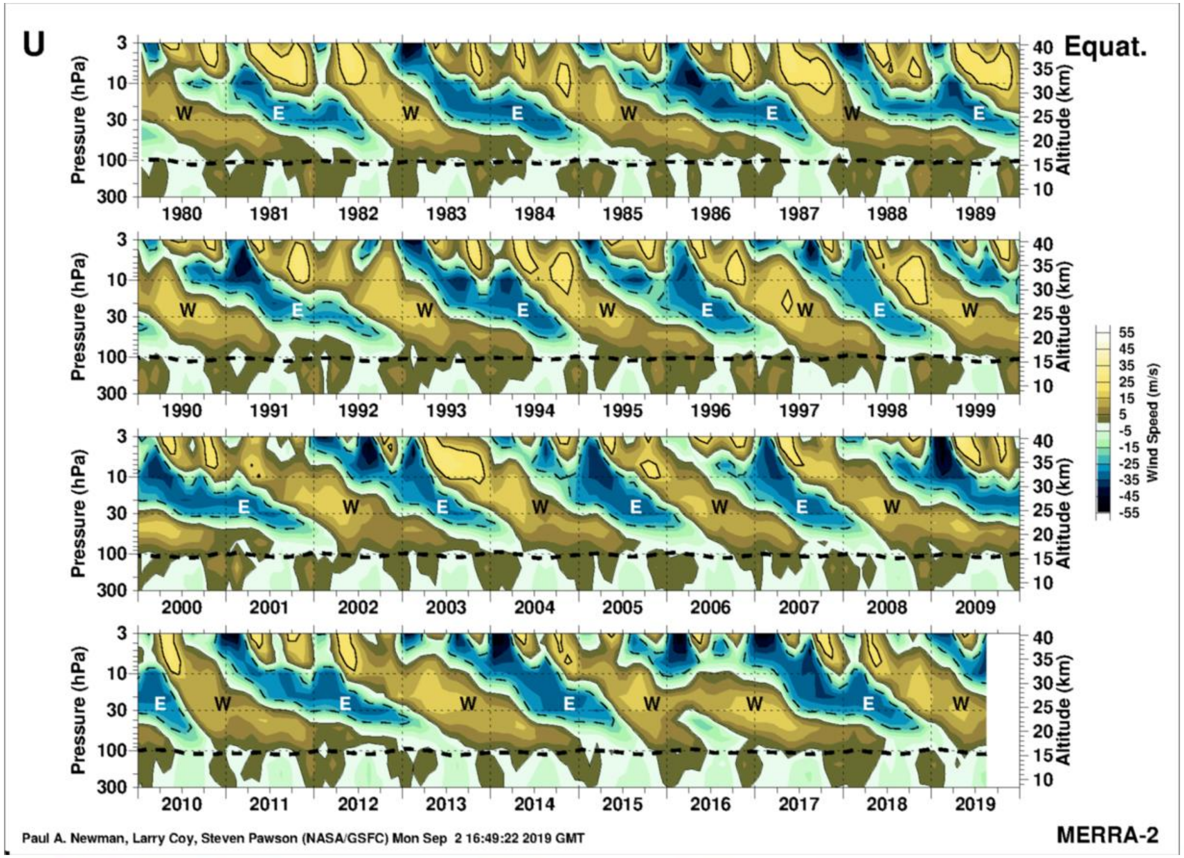


Polar vortex evolution during the 2002 Antarctic major warming as observed
by the Odin satellite
P. Ricaud et al. 2005, JGR, doi.org/10.1029/2004JD005018

Stratosphere and Quasi Biennial Oscillation in equatorial regio

- One of the most repeatable phenomena seen in the atmosphere, the quasi-biennial oscillation (QBO) between prevailing eastward and westward wind jets in the equatorial stratosphere (approximately 16 to 50 kilometers altitude)

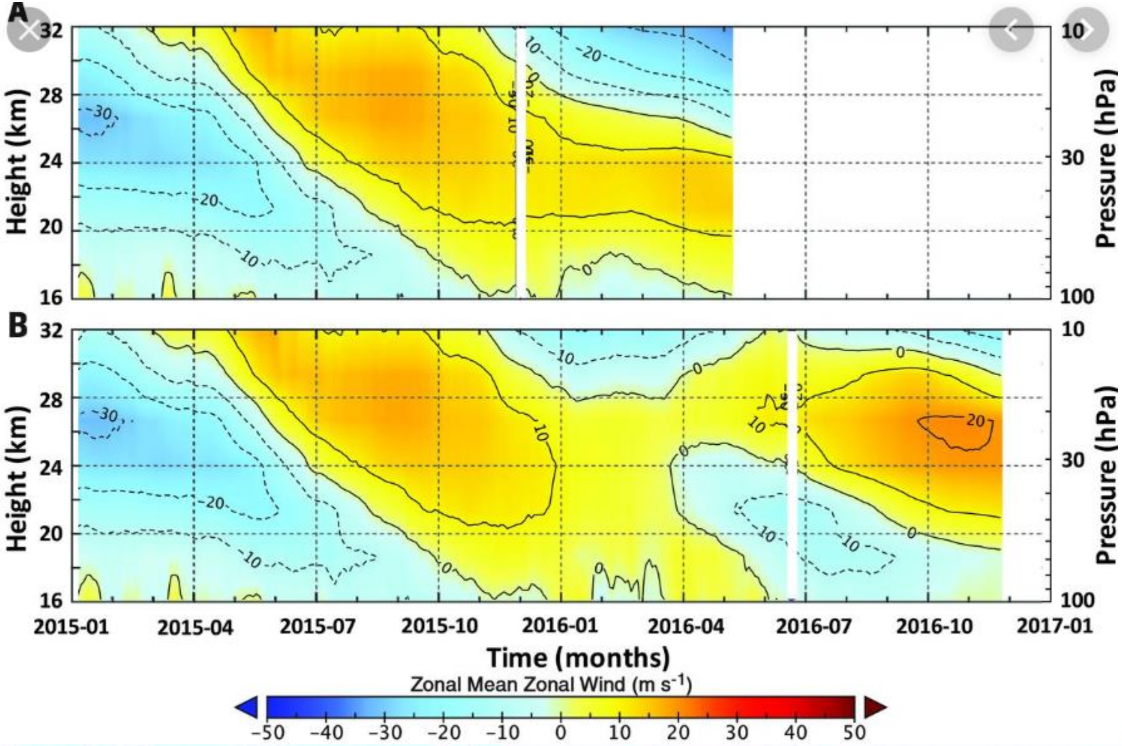
Mean Zonal wind



Paul A. Newman, Larry Coy, Steven Pawson (NASA/GSFC) Mon Sep 2 16:49:22 2019 GMT

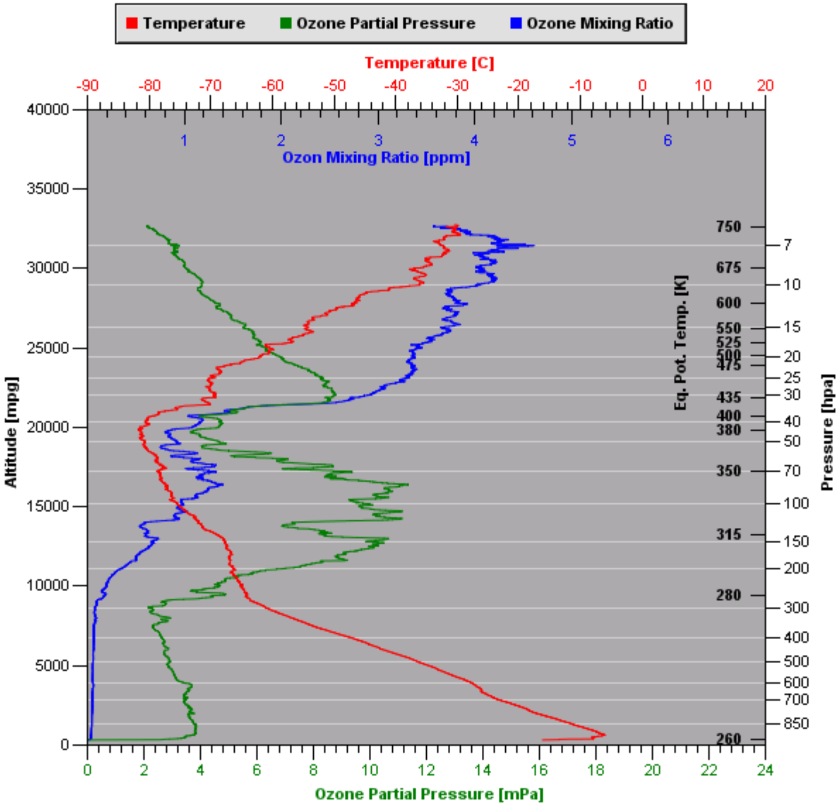
An unexpected disruption of the atmospheric quasi-biennial oscillation 2016

- QBO was unexpectedly disrupted in February 2016. An unprecedented westward jet formed within the eastward phase in the lower stratosphere and cannot be accounted for by the standard QBO paradigm based on vertical momentum transport.
- *S. M. Osprey et al., Science* 23 Sep 2016: Vol. 353, Issue 6306, pp. 1424-1427, DOI: 10.1126/science.aah4156



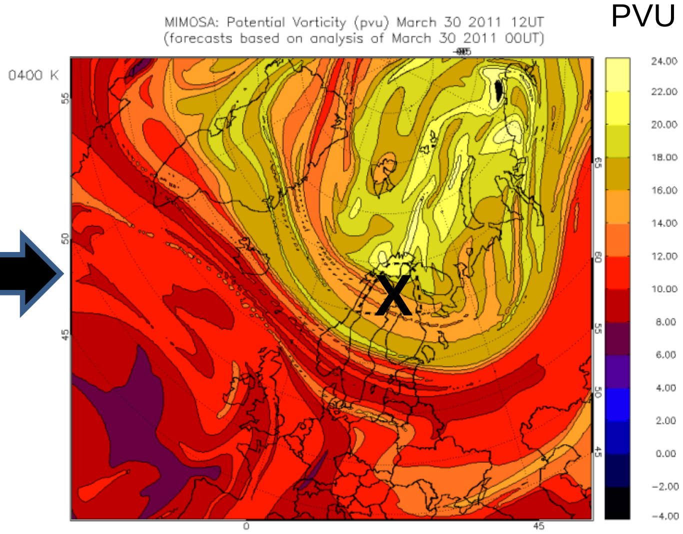
Ozone loss : winter2010/spring 2011 Event

Temperature and Ozone profiles



Potential vorticity map

400K isentropic surface
(~65 hPa, ~19 km)

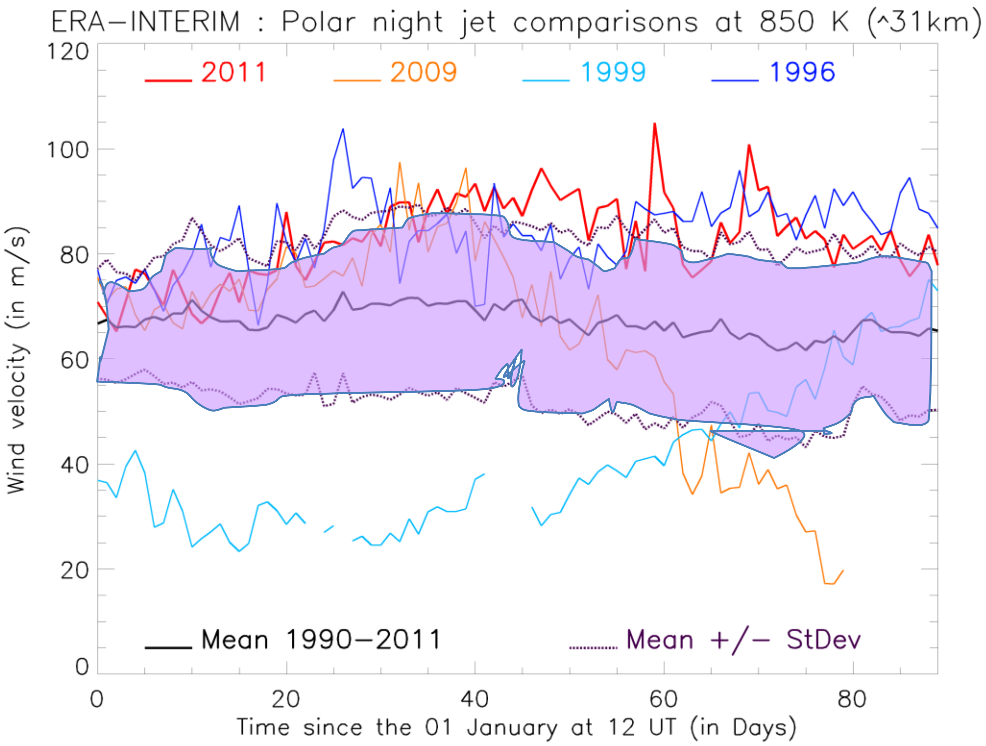


MIMOSA model (Hauchecorne et al. 2002)

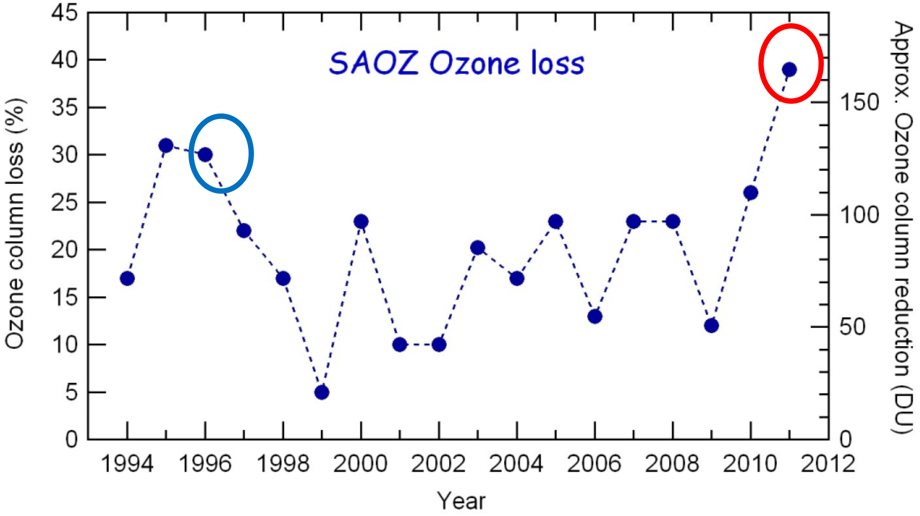
Geophysical conditions on isentropic surface 400K inside the vortex

- Temperature < - 85°C
- Ozone mixing ratio <1 ppmv,~ [17;22] km

2011 Very intense polar vortex in Arctic region



Ozone loss never seen during 20 last years



http://saoz.obs.uvsq.fr/O3Loss.html, *Florence Goutail, LATMOS.*

Exceptional conditions for dynamical and chemical studies

2011 unprecedent Arctic ozone loss

- Unprecedented Arctic ozone loss in 2011

Gloria L. Manney, Michelle L. Santee, Markus Rex, Nathaniel J. Livesey, Michael C. Pitts, Pepijn Veefkind, Eric R. Nash, Ingo Wohltmann, Ralph Lehmann, Lucien Froidevaux, Lamont R. Poole, Mark R. Schoeberl, David P. Haffner, Jonathan Davies, Valery Dorokhov, Hartwig Gernandt, Bryan Johnson, Rigel Kivi, Esko Kyrö, Niels Larsen, Pieternel F. Levelt, Alexander Makshtas, C. Thomas McElroy, Hideaki Nakajima, Maria Concepción Parrondo *et al.*

Nature (2011) doi:10.1038/nature10556

Received 03 May 2011 Accepted 07 September 2011 Published online 02 October 2011

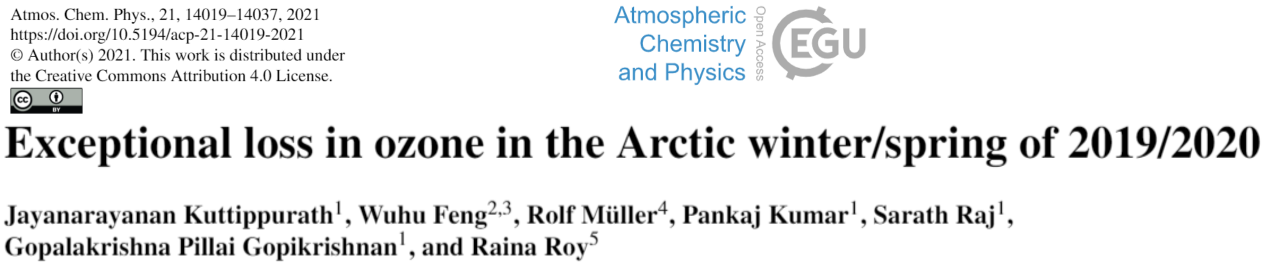
- The Arctic vortex in March 2011: a dynamical perspective

M. M. Hurwitz, P. A. Newman, and C. I. Garfinkel

ACPD (2011)

Received: 5 July 2011 – Accepted: 2 August 2011 – Published: 5 August 2011

What’s new ? winter 2019/spring 2020



Atmos. Chem. Phys., 21, 14019–14037, 2021
https://doi.org/10.5194/acp-21-14019-2021
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Atmospheric Chemistry and Physics
Open Access
EGU

Exceptional loss in ozone in the Arctic winter/spring of 2019/2020

Jayanarayanan Kuttippurath[1], Wuhu Feng[2,3], Rolf Müller[4], Pankaj Kumar[1], Sarath Raj[1],
Gopalakrishna Pillai Gopikrishnan[1], and Raina Roy[5]

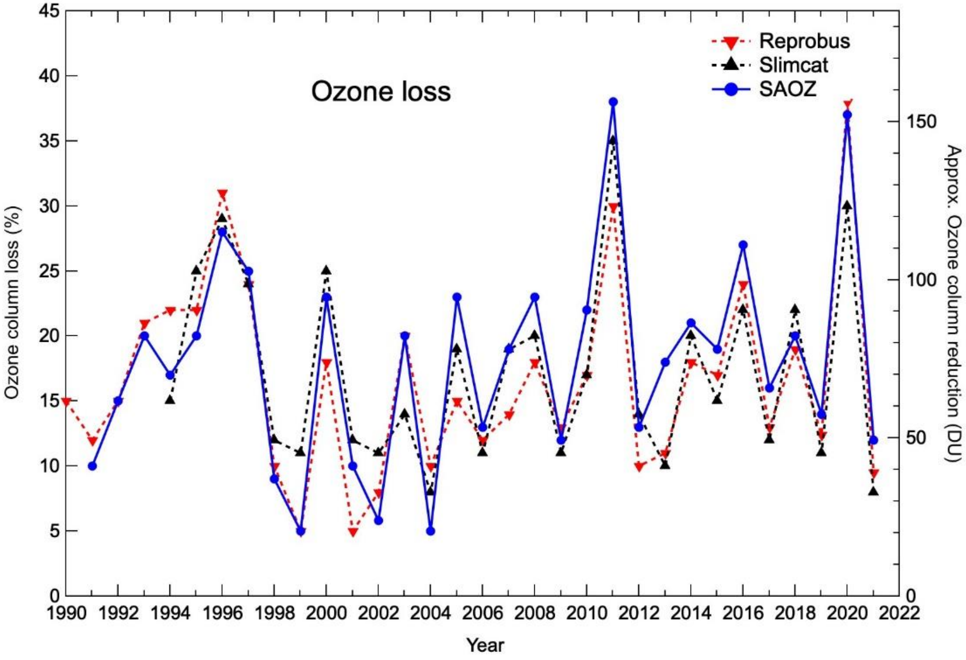
- 2020 ozone loss

similar to 2011 ozone loss :

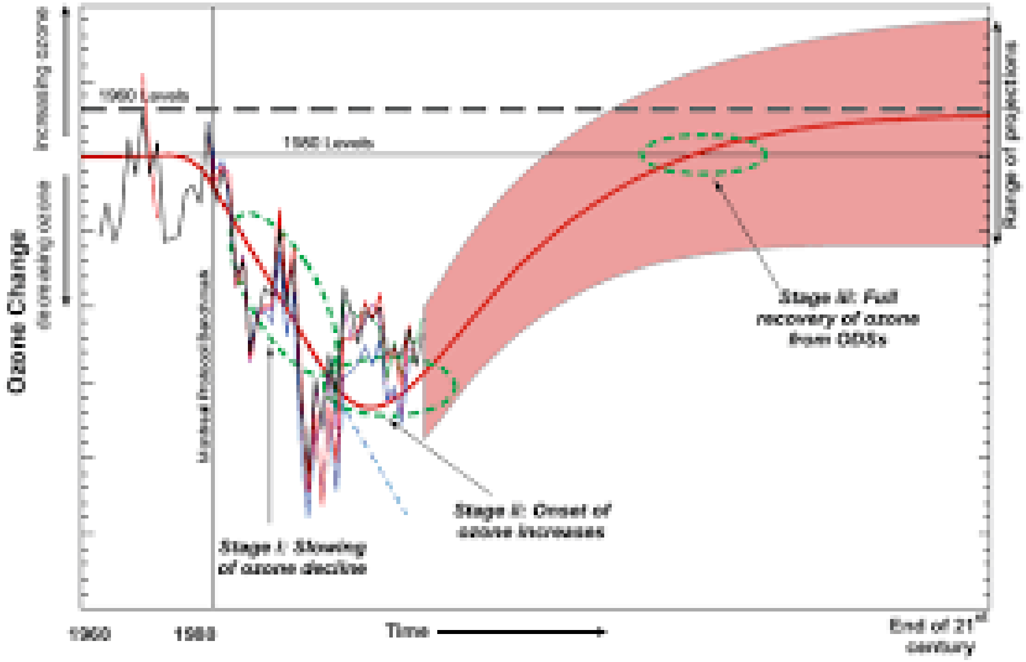
-38% of the O_3 total column

-150 O_3 DU

F. Goutail, LATMOS/CNRS
http://saoz.obs.uvsq.fr/O3Loss.html

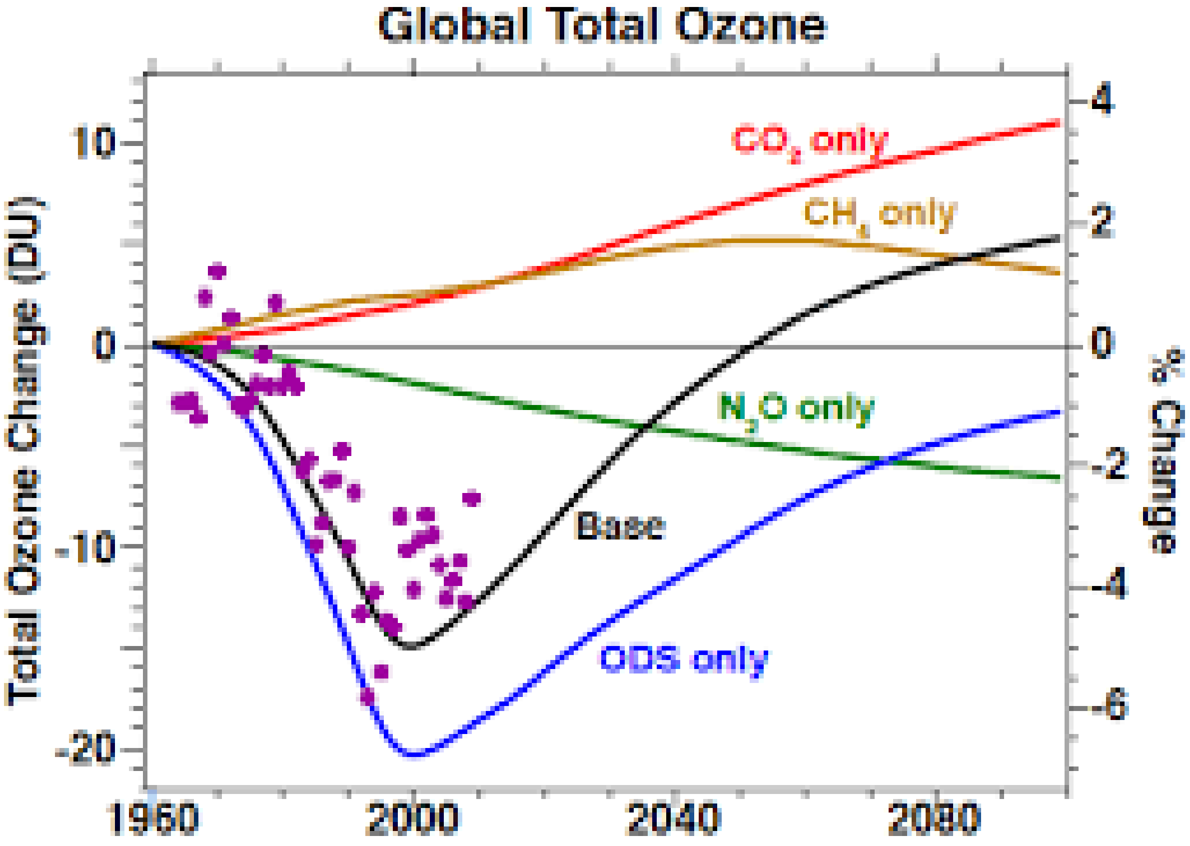


Ozone layer recovery



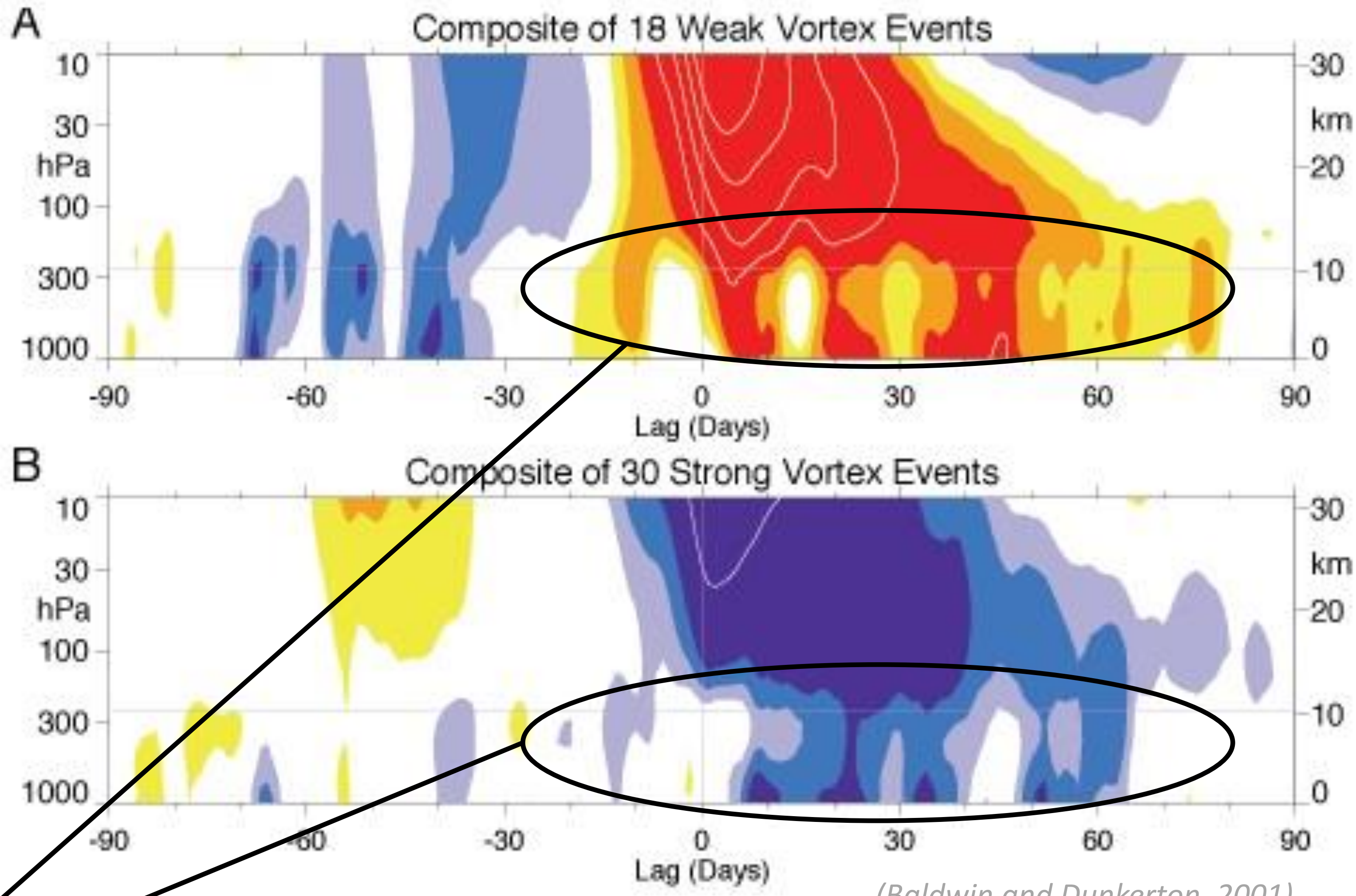
- Recovery could be possible mis 21th century but depends on GES in the troposphere

Ozone layer recovery



- Dépends climate forcing due to impact of GES on

Stratospheric influence on surface climate



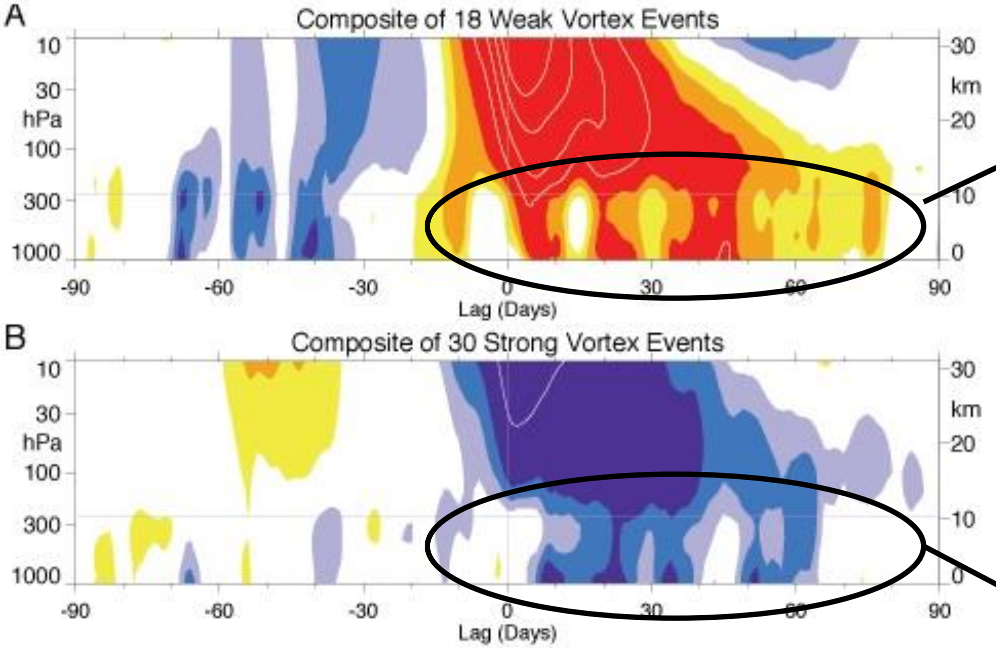
(Baldwin and Dunkerton, 2001)

What does it mean ?

Polar vortex variability affects the surface !

Surface response

Downward propagation of the signal

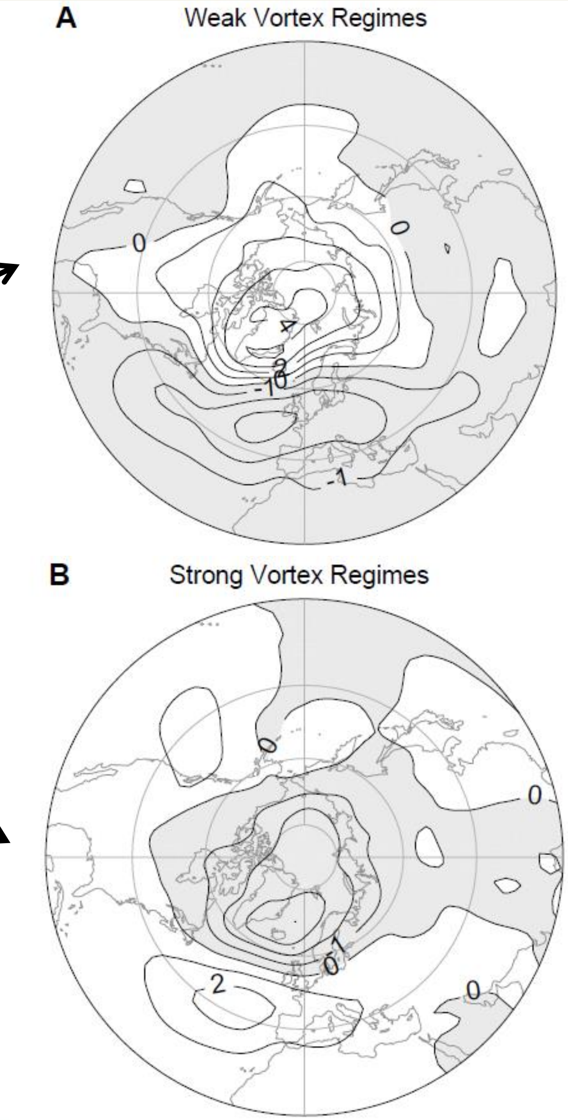


Composite of 18 Weak Vortex Events

Composite of 30 Strong Vortex Events

(Baldwin and Dunkerton, 2001)

Surface response (Sea Level Pressure)



Weak Vortex Regimes

Strong Vortex Regimes

The signal project onto a „North Atlantic Signal“

Influence on troposphere weather ?

Take home messages

- Atmosphere circulation: redistribution of energy due to SW absorption and LW emission asymetry
- Stratosphere radiatively balanced but not the troposphere ⇔ latent heat release from water vapour crucial for troposphere energy budget and circulation.
- Negative vertical temperature gradient in the troposphere leads to high potential unstability and convection, while the stratosphere is very stable with very slow vertical transport.
- Mid-latitudes regional circulation strongly depends on Coriolis acceleration
- Brewer-Dobson large-scale circulation in the stratosphere mainly driven by Rossby waves which propagate from the troposphere and which are responsible for the stratospheric dynamical variability
- Troposphere weather coupled with stratosphere dynamical conditions (but not only !)
- Chemical and dynamical processes are strongly interacting

ESPRI Data Centre (former Ether)

ESPRI is a data centre of the French Atmosphere Infrastructure AERIS

- Aeris data center

Model Forecasts :

3D CTM REPROBUS

Reactive Processes Ruling the Ozone Budget in the Stratosphere) temporal evolution of 55 stratospheric species by 147 réactions chimiques [Lefevre et al., 1998].

MIMOSA :

potential vorticity contour advection model (hauchecorne et al. 2001)

- http://espri.aeris-data.fr

Some literature