The middle atmosphere and the parametrization of non-orographic gravity wave drag

Peter Bechtold and Andrew Orr

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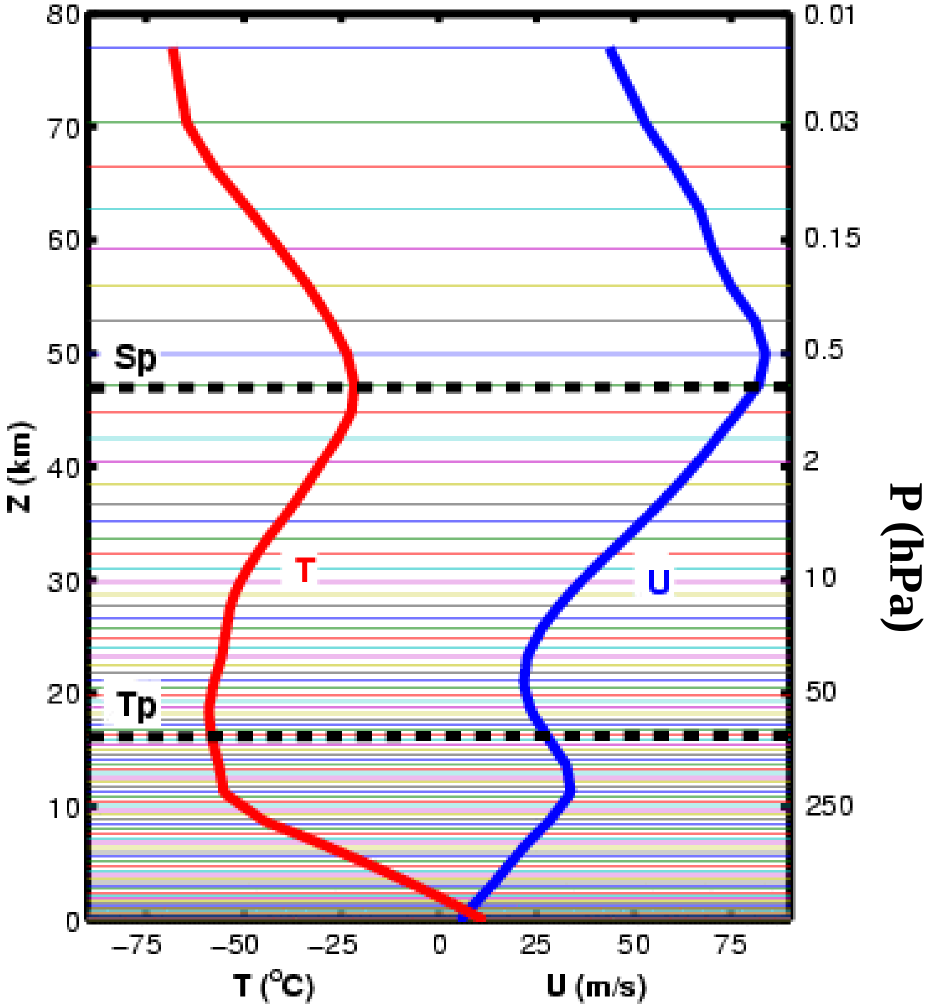
Literature

- Holton, 2004: An introduction to Dynamic Meteorology, *AP*
- Ern, M., P. Preusse, M.J. Alexander & C.D. Warner, 2004: Absolute values of gravity wave momentum flux derived from satellite data. *J. Geophys. Res.*, **109**, D20103, doi:10.1029/2004JD004752
- Ern, M., P. Preusse & C.D. Warner, 2006: Some experimental constraints for spectral parameters used in the Warner and McIntyre gravity wave parameterization scheme. *Atmos. Chem. Phys.*, **6**, 4361-4381.
- Mc Landress, C. & J.F. Scinoccia, 2005: The GCM response to current parameterizations of nonorographic gravity wave drag. *J. Atmos. Sci.*, **62**, 2394-2413.
- Scinocca, J.F, 2003: An accurate spectral nonorographic gravity wave drag parameterization for general circulation models. *J. Atmos. Sci.*, **60**, 667-682.
- Warner, C.D & M.E. McIntyre, 1996: On the propagation and dissipation of gravity wave spectra through a realistic middle atmosphere. *J. Atmos. Sci.*, **53**, 3213-3235.
- Yamashita et al., 2010: Gravity wave variations during the 2009 stratospheric sudden warming as revealed by ECMWF-T999 and observations. GRL, **37**, L22806
- Ern, M and P. Preusse, 2012: Gravity wave momentum flux spectra observed from satellite in the summertime subtropics. GRL, **39**, L15810
- Orr, Bechtold, Scinnoca, Ern, Janiskova, 2010 : *JCL* (see Lecture Note)

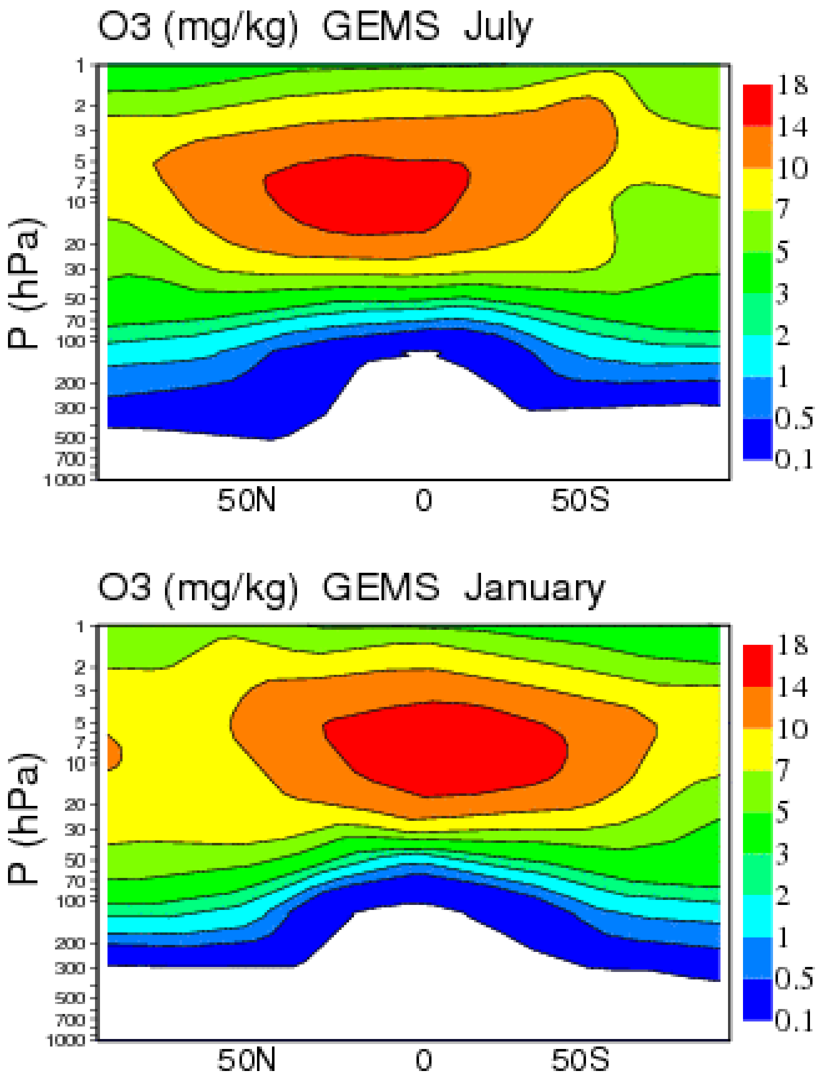
Structure of the Atmosphere: Troposphere & Stratosphere



Typical Temperature and Zonal Wind profiles for July at 40S, together with the distribution of the 91-levels in the IFS. **Tp** denotes the **Tropopause**, **Sp** the **Stratopause**, the model top also corresponds to the **Mesopause**

- Temperature decrease in the Troposphere is due to adiabatic decompression
- Midlatitude upper-tropospheric Jet form due to strong temperature gradient between Pole and Equator.
- The temperature in the Stratosphere increases due to the absorption of solar radiation by ozone

O3 zonal mean concentrations from GEMS



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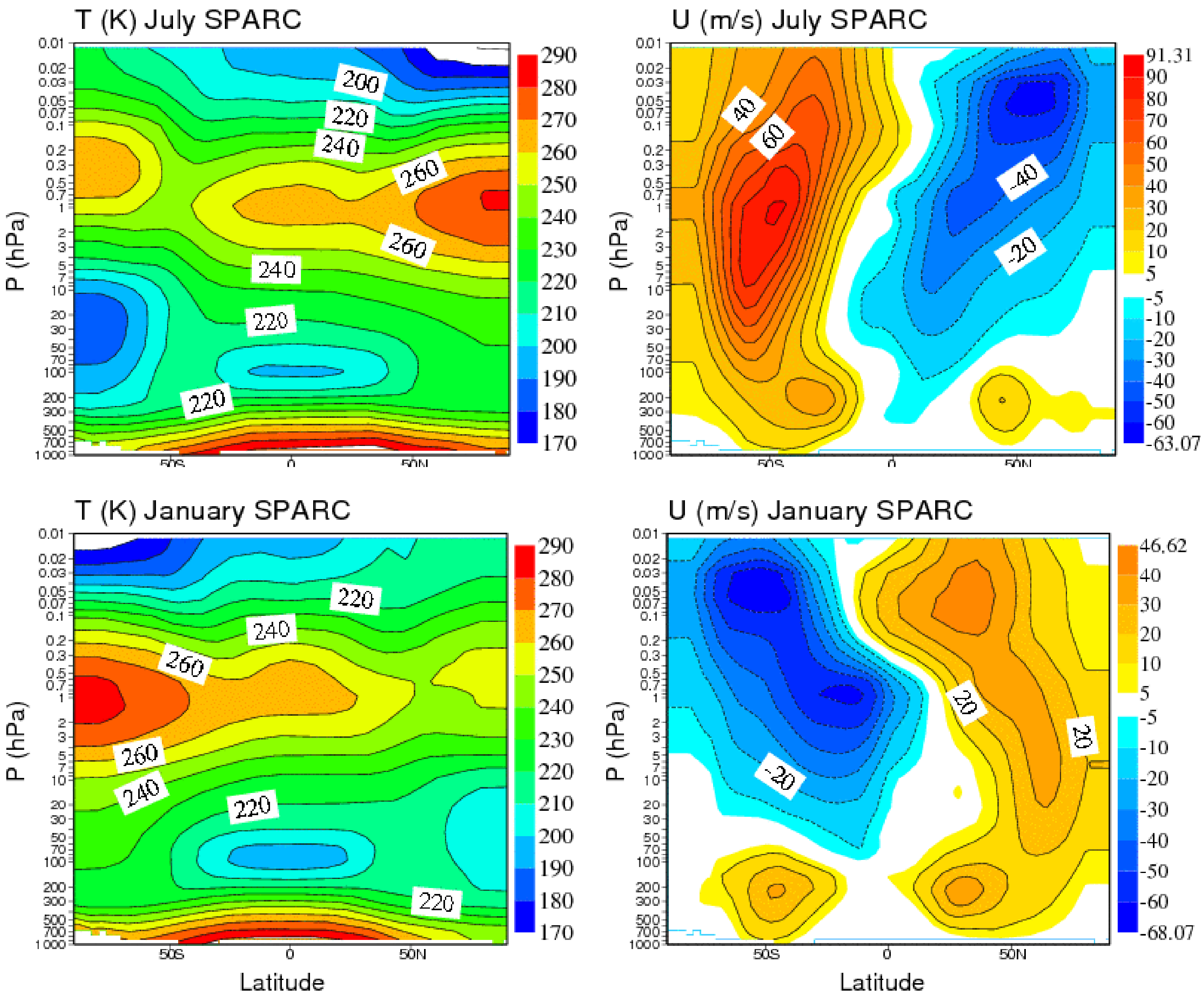
July & January climatology

SPARC



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Structure of the Stratosphere & Mesosphere

- As heating due to ozone starts to decrease with height, so does the temperature
- Radiative heating/cooling of summer/winter hemispheres causes air to rise/sink at the summer/winter poles, inducing a summer to winter pole meridional circulation
- Coriolis acceleration of meridional circulation produces easterly/westerly jets in the summer/winter hemispheres
- The large Coriolis acceleration implies the existence of some eddy forcing to balance the momentum budget. This is provided by the breaking/dissipation of vertically propagating planetary waves, and small-scale non-orographic gravity waves. GWs transport energy and momentum vertically

Radiosonde observations of gravity waves

Pronounced waviness in profiles due to gravity waves
Vertical wavelength ~12km

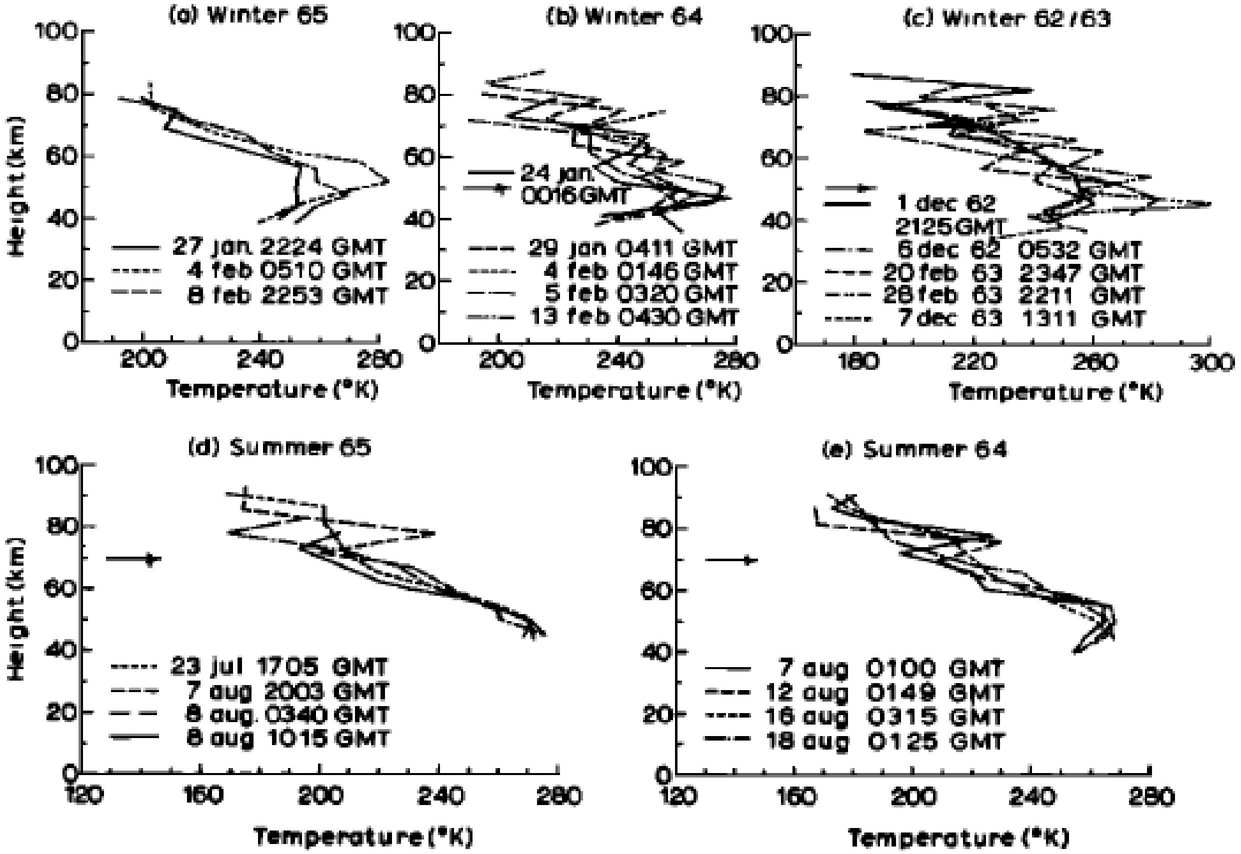
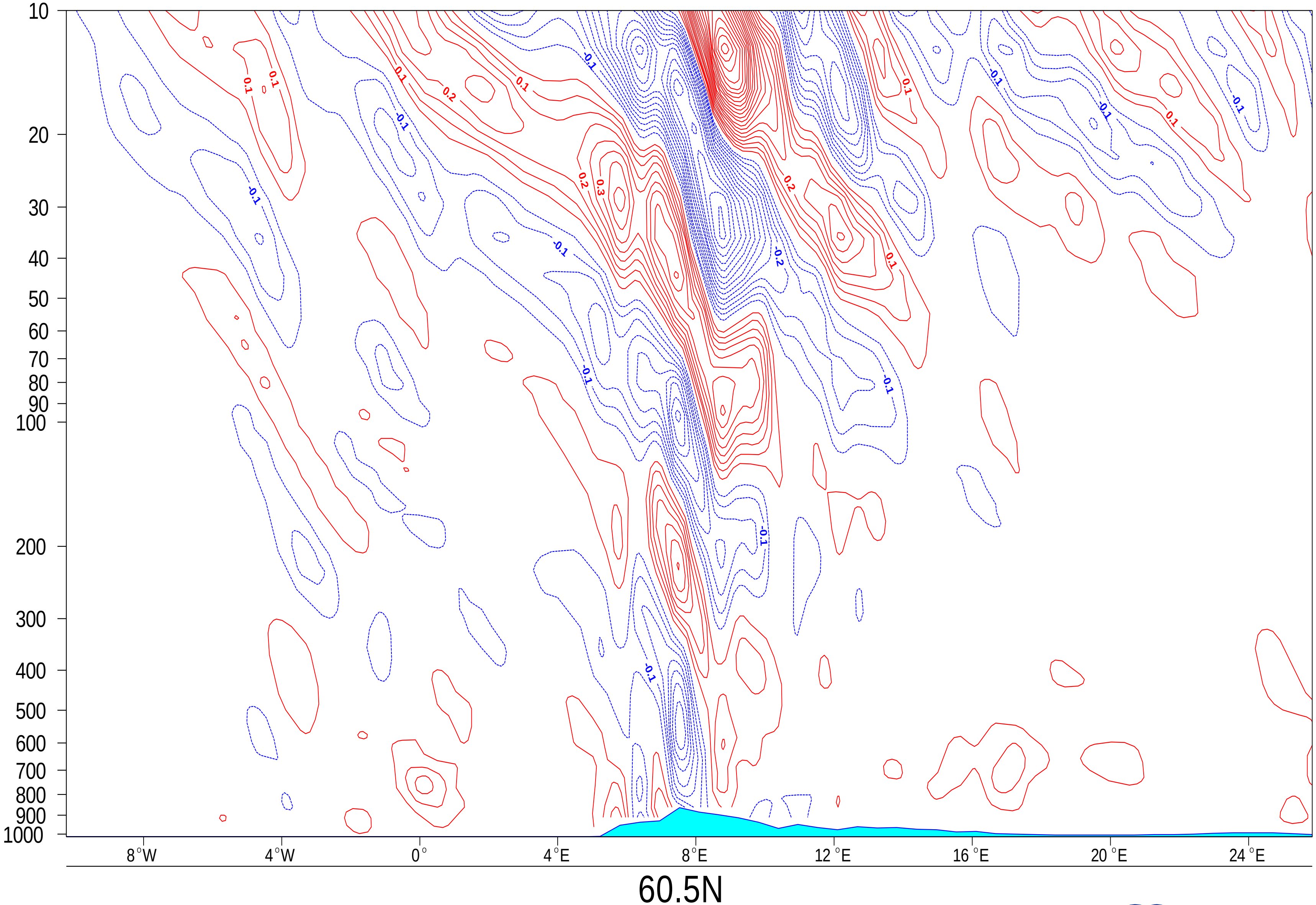


Fig. 1. Winter and summer temperature sounding at Wallops Island (38°N).

From Lindzen (1981)

Hydrostatic-IFS 1279 Vertical velocity w



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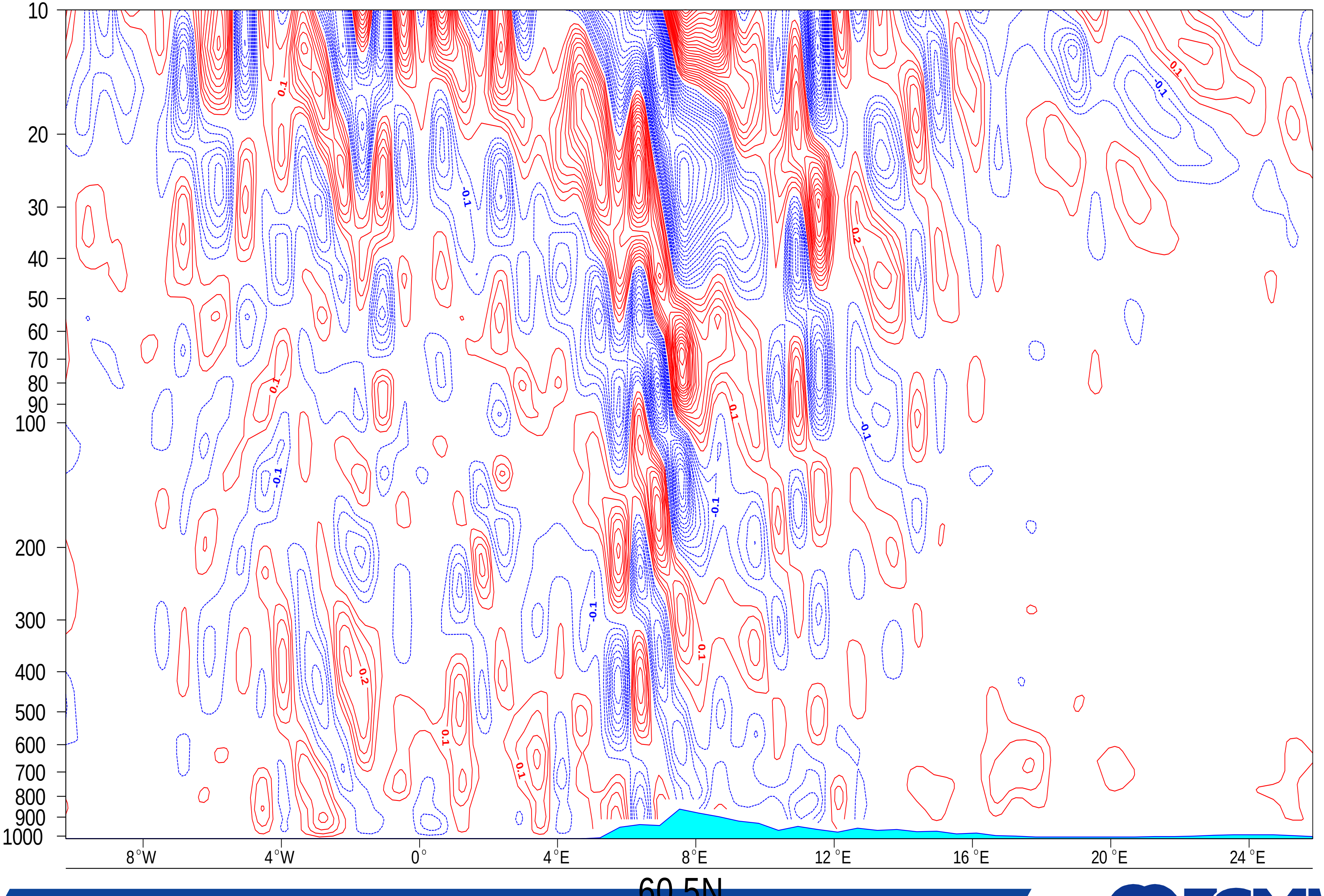
NonHydrostatic-IFS 3999 Vertical velocity w



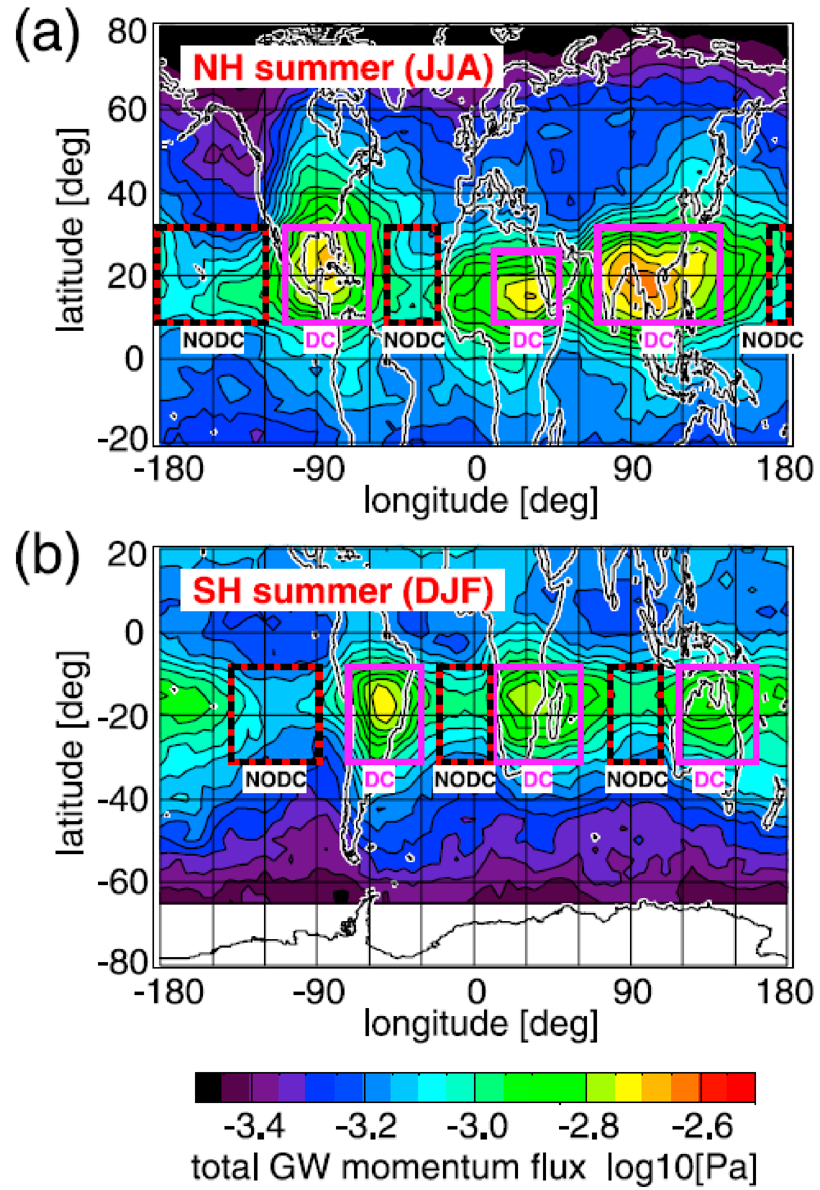
Gravity waves: from Observation

High resolution Limb sounder

Vertical wavelengths: 1-10s km
Horizontal wavelengths: 10s-1000s km

Sources: convection, fronts, jet-stream activity

Unresolved or under-resolved by the IFS



Ern and Preusse, GRL, 2012

Latitudinal and seasonal dependence

Wave representation

$$\psi(t,\vec{x},z) = \hat{\psi}(z)\cos(\vec{k}\,\vec{x} + mz - \omega t);$$

$$\psi(t,\vec{x},z) = \hat{\psi}(z)e^{i(\vec{k}\,\vec{x}+mz-\omega t)} = \hat{\psi}(z)e^{i(\vec{k}\,(\vec{x}-\vec{c}t)+mz)}$$

$$\omega = \quad frequency$$

$$\left|\vec{k}\right| = \quad horizontal\ wavenumber$$

$$m = \frac{2\pi}{\lambda_z} \quad vertical\ wavenumber; \lambda_z\ vertical\ wavelength$$

$$c = \frac{\omega}{k} = horizontal\ \ phase\ speed$$

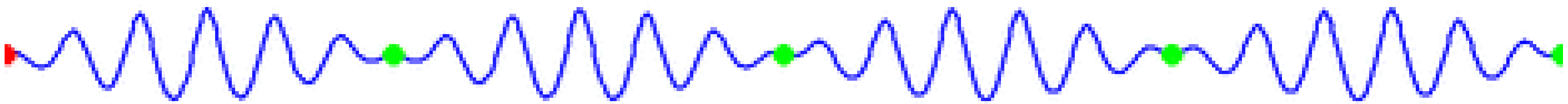
$$c_g = \frac{\partial\omega}{\partial k} = group\ velocity$$

(Hamilton, 1839, Rayleigh 1877)

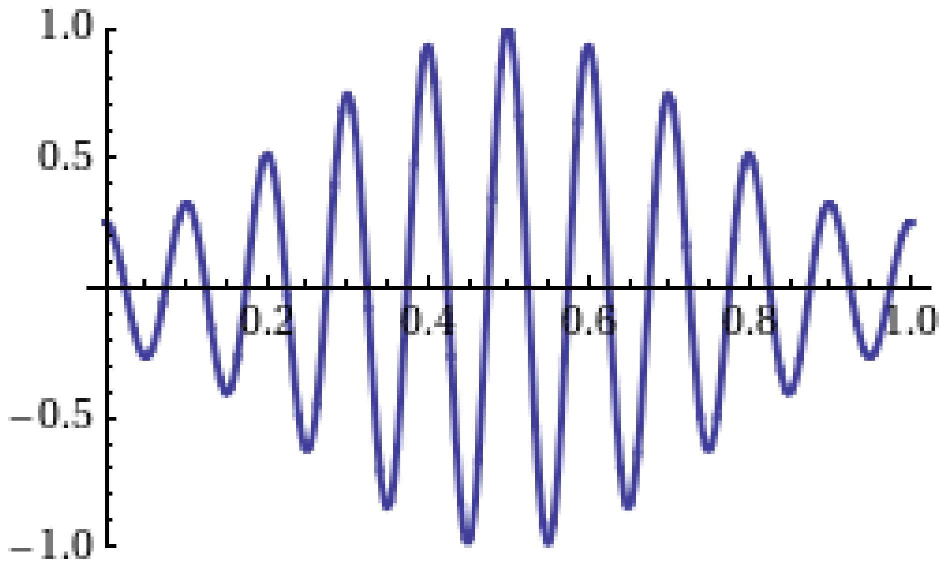
$$\tilde{\omega} = \omega - kU; \quad \tilde{c} = c - U = \text{intrinsic}\ frequency$$

$$and\ \ phase\, speed$$

Examples of waves as function of x,t (source WIKIPEDIA)



Example of group of gravity waves: The red dot moves with the phase speed, the green dots with the group speed



Example where phase and group velocity go in different directions (e.g. Rossby waves in absence of westerly background wind)

What is a “ non-orographic” gravity wave?

Orographic gravity waves are supposed to be stationary (ω=0 => zero horizontal phase speed)

Non-orograpgic gravity waves are non-stationary, and therefore have non-zero phase speed. The parametrization problem is therefore 5-dimensional!

$$\Psi(j, z, k, \omega, \phi)$$

Depending on gridpoint *j*, height *z*, wavenumber *k*, frequency **ω**, and direction **Φ**

How to proceed for a simple parametrization

- Define a launch spectrum

- Define the relation between ω and κ. This is called the dispersion relation, and depends on the equation system (hydrostatic or non-hydrostatic shallow water) used to derive the waves (see Appendix in convection Lecture Note).

- Define in which physical space one wants to propagate the wave spectrum: either ω-κ coordinate frame or $\tilde{\omega}$ -m resp. C -**m** coordinate frame.
  For practical reasons one wants to have conservative propagation from one level to the other as long as there is no dissipation.

- Define dissipation procedure, i.e. critical level filtering + some adhoc nonlinear dissipation mechanism to account for wave braking as amplitude increases with height due to decreasing density.

Dispersion relation for gravity waves

$$m^2 = \frac{k^2 N^2}{\tilde{\omega}^2} \frac{(1-\tilde{\omega}^2 / N^2)}{(1- f^2 / \tilde{\omega}^2)}$$

$$\tilde{\omega}^2 / N^2 \Rightarrow 0 \quad (\textit{hydrostatic wave dynamics})$$

$$f^2 / \tilde{\omega}^2 \Rightarrow 0 \quad (\textit{no rotation})$$

Wavelike solutions exist for $f^2 < \tilde{\omega}^2 < N^2$

For simplicity we only consider hydrostatic, non-rotational waves which also allows to ignore the effect of back reflection of waves

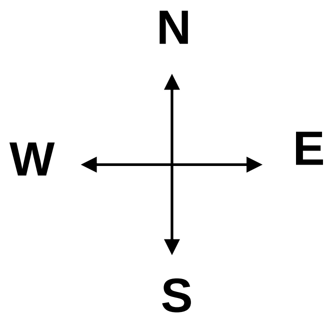
$$m^2 = \frac{k^2 N^2}{\tilde{\omega}^2} = \frac{N^2}{\tilde{c}^2}$$

Wavelike solutions exist for $\tilde{\omega}^2 > 0$ and critical level filtering occurs when the intrinsic phase speed approaches zero

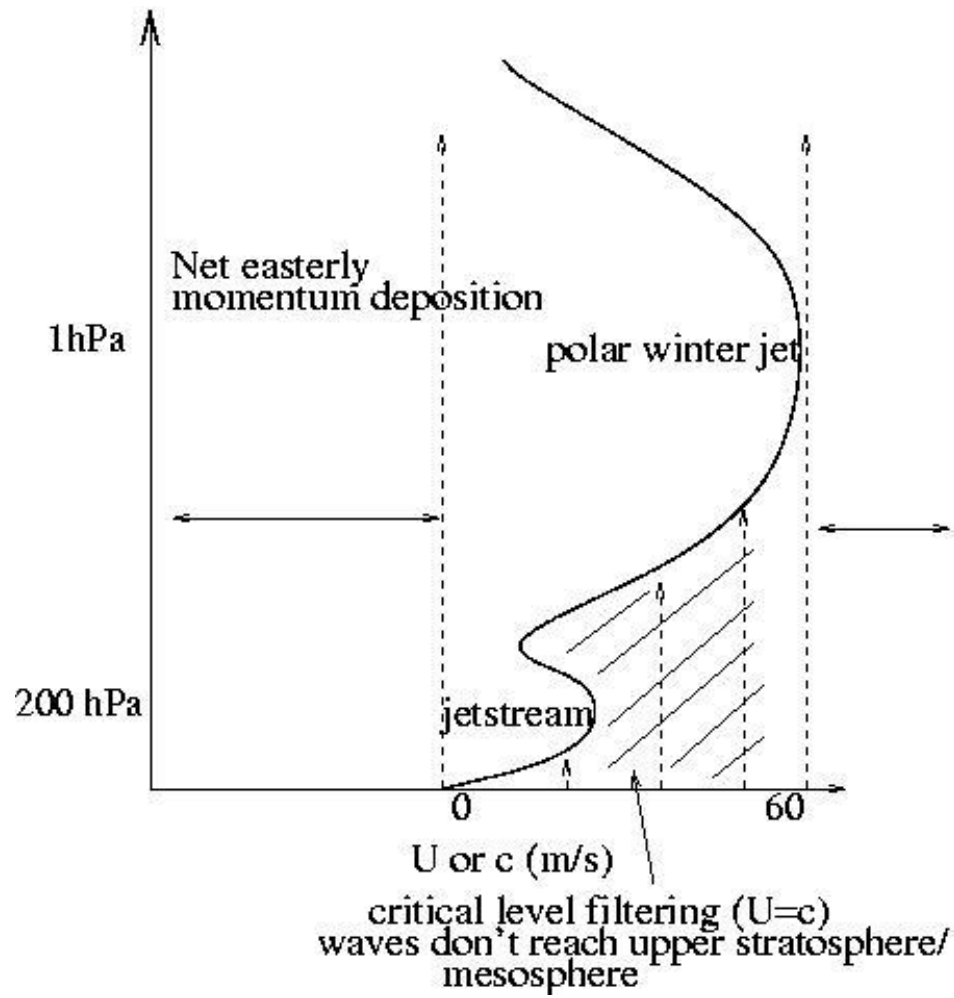
Physically based gravity wave scheme

Rely on realistic winds to filter the upward propagating (unrealistic) gravity wave source.

Gravity wave source: launch globally constant isotropic spectrum of waves at each grid point as function of, for example, c. Assume constant input momentum flux



Net input momentum flux = 0



Simplified hydrostatic non-rotational version of Warner and McIntyre (1996) scheme (Scinocca, 2003) - WMS

- **In any azimuth, φ, the launch spectrum is specified by the total wave energy per unit mass,** $\hat{E}_0$

- **This is chosen to be the standard form of Fritts and VanZandt (1983), which in** $m-\tilde{\omega}$ **space is**

$$\hat{E}_0(m,\tilde{\omega},\varphi) = B\left(\frac{m}{m_*}\right)^s \frac{N_0^2 \tilde{\omega}^{-p}}{1-\left(\frac{m}{m_*}\right)^{s+t}}$$

- ***It is assumed to be separable in terms of*** m ***and*** $\tilde{\omega}$ ***.***

- **The 'tail spectrum' is largely independent of time, season, and location (VanZandt, 1982)**

Observations: Fritts and VanZandt (1983) and VanZandt (1982)

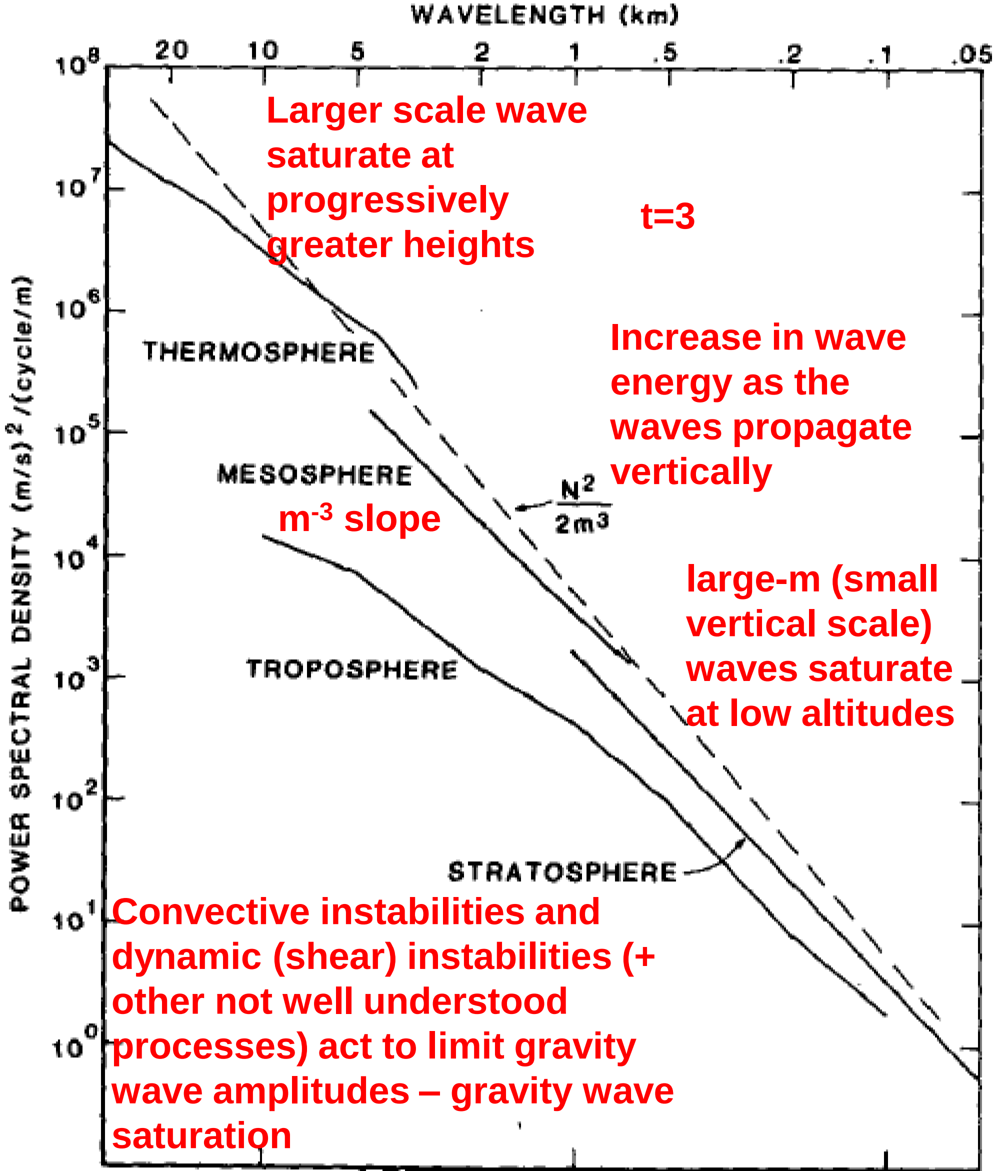


FIG. 1. Spectra of horizontal velocity versus vertical wavenumber as a function of altitude.

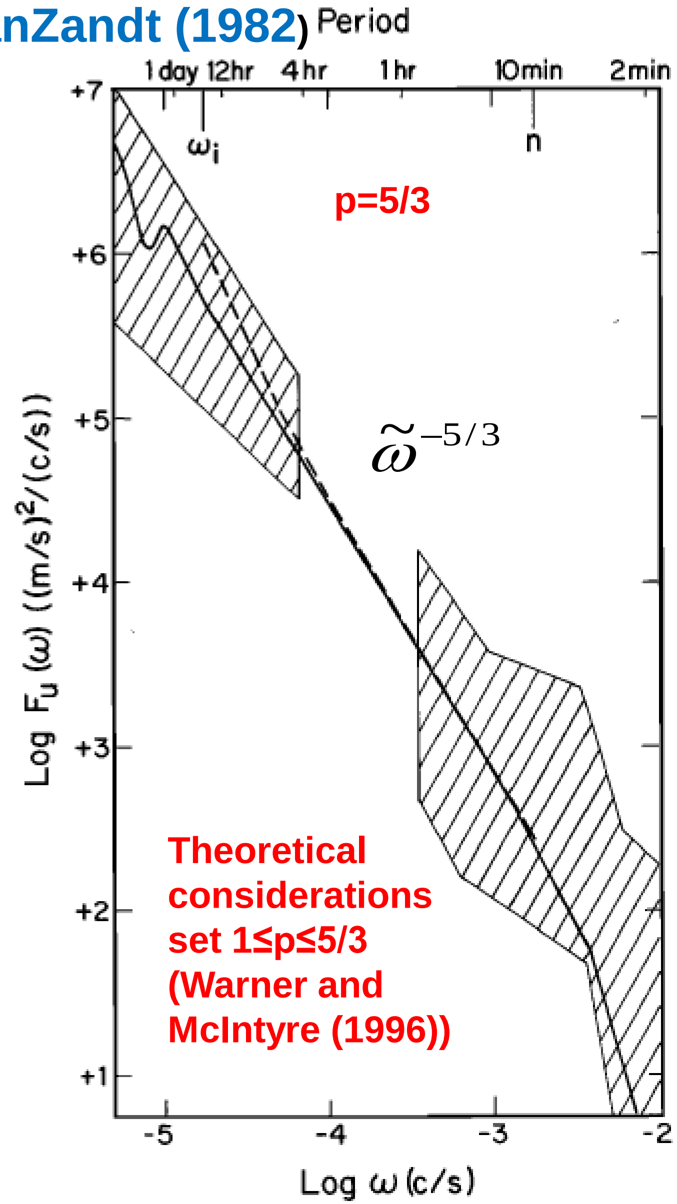
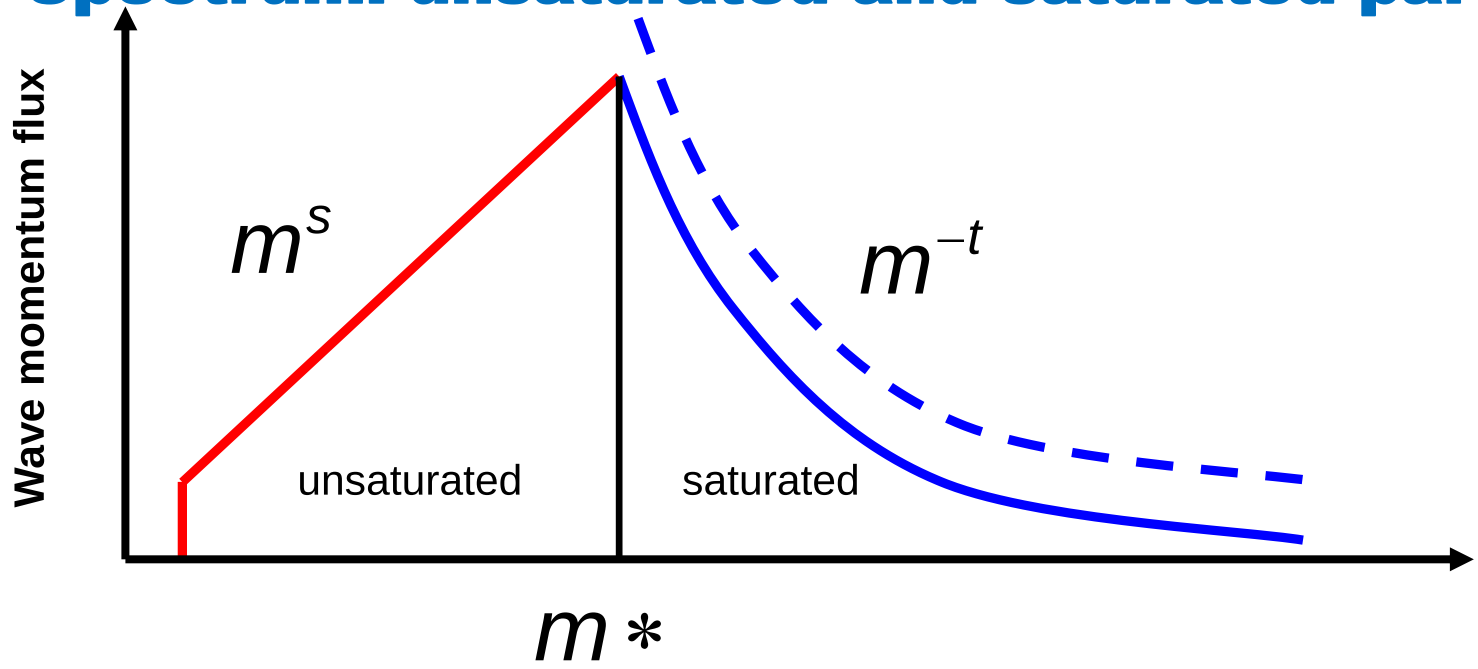


Fig. 1. Power spectral density $F_u(\omega)$ of the horizontal wind u versus frequency. The central solid curve is the mean observed spectrum estimated by Vinnichenko (1970). The upper and lower limits of the hatched areas are the "envelopes formed by the maxima and minima" of the observed spectra that went into estimating the mean. The model $F_u(\omega)$ spectrum is shown by the dashed curve.

Spectrum: unsaturated and saturated part



- The characteristic vertical wavenumber m*, separating the saturated and unsaturated slopes is 2π/ 2km, with 2km the characteristic vertical wavelength.

- There is one free parameter in the scheme that allows to shift the saturation curve (dashed blue curve) to the right, with the result that non-linear dissipation is occuring at greater heights. As we will se, and as documented in the literature, this has important consequences for the simulation of the QBO

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- Specify in terms of momentum flux spectral density, $\rho_0 F_0$ using group velocity rule

$$\rho_0 F_0(m,\tilde{\omega},\varphi) = \rho_0 c_{gz} \frac{k}{\tilde{\omega}} \hat{E}_0(m,\tilde{\omega},\varphi) \qquad c_{gz} = \partial\tilde{\omega}/\partial m = \tilde{\omega}/m$$

- m and $\tilde{\omega}$ are not invariant to vertical changes in ***U(z)*** and ***N(z)*** (i.e. dependent variables).

- Chose $c-\varphi$ space to describe the vertical propagation of the wave field

Galilean Transform

$$\widehat{U} = U - U_0 \text{ and } \widehat{c} = c - U_0$$

$$\rho\bar{F}^*(\widehat{c},\phi) = \rho \frac{\widehat{c}-\widehat{U}}{N}\left(\frac{\widehat{c}-\widehat{U}}{\widehat{c}}\right)^{2-p} \frac{1}{1+\left(\frac{m_*(\widehat{c}-\widehat{U})}{N}\right)^{s+3}}$$

Final result of scaled Eliassen-Palm flux density (ρ m^2s^{-2}/dc) obtained through scaling with launch momentum flux, the most important parameter of scheme

$$\rho\bar{F}(\widehat{c},\phi) = \rho\bar{F}^*(\widehat{c},\phi) * A; \quad A = (\rho_0 F_{launch}) / \int \rho F^*(\widehat{c},\phi_1) d\widehat{c}$$

How to do critical level absorption

At the launch level we have $\widehat{U}_0 = 0$

Which sets an absolute lower bound for critical level absorption of $\widehat{c} = 0$.

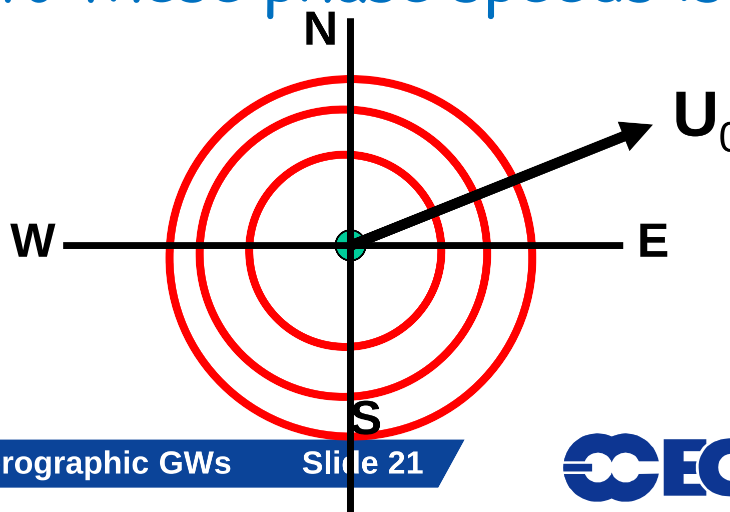
If on the next vertical level $z_1(>z_0)$ U increases such that

$\widehat{U}_1 > \widehat{U}_0$ then waves with phase speeds in the range

$\widehat{U}_0 < \widehat{c} < \widehat{U}_1$ encounter critical level absorption and the

momentum flux corresponding to these phase speeds is

removed from $\rho\overline{F}(\widehat{c},\phi)$



The saturated spectrum

Convective instabilities and dynamic (shear) instabilities (+ other not well understood processes) act to limit gravity wave amplitudes – gravity wave saturation

Results in the "universality" of the GW spectrum m^{-3} (Smith et al. 1987)

WMS scheme deals with non-linear dissipation in an empirical fashion by limiting the growth of the GW spectrum so as not to exceed saturated spectrum ~m-3.

1) **Achieved by specifying a saturation upper bound on the value of the wave energy density at each level with the observed m^{-3} dependence at large-m**

$$\hat{E}^{sat}(m,\tilde{\omega},\varphi) = C^{*}\left(\frac{m}{m_{*}}\right)^{-3} N^{2}\tilde{\omega}^{-p}$$

2) **Which can be expressed as**

$$\rho\overline{F}^{sat}(\hat{c},\phi) = \rho C^{*}\,\frac{\hat{c}-\hat{U}}{N}\left(\frac{\hat{c}-\hat{U}}{\hat{c}}\right)^{2-p}$$

3) **Unlike the unsaturated spectrum, $\rho\overline{F}$, the saturated spectrum is not conserved, $\rho\overline{F}^{sat}$, and so decreases in amplitude with height as a result of diminishing density. This limits $\rho\overline{F}$ to a saturation condition**

$$\rho\overline{F}(\hat{c},\varphi) \le \rho\overline{F}^{sat}(\hat{c},\varphi)$$

Parameter specification

1) t=3 (fixed)
2) p=1 (or 3/2 ; observed/theoretical $1 \leq p \leq 5/3$)
3) s=1 (or 0,1; s=1 most common, ie positive slope required)
4) m* (= $2\pi / 2km$, see Ern et al (2006))
5) C* (=1; raising this raises the height momentum is deposited)
6) φ =4 (number of azimuths, although can have 8, 16, …)

7) z_0 (launch level; Ern et al. (2006) suggests either 450 hPa or 600 hPa)
8) $\rho_0 F_0$ input momentum flux into each azimuth is set to 3.75×10^{-3} (Pa)

Ern, Preusse, and Warner, 'Some experimental constraints for spectral parameters used in the Warner and McIntyre gravity wave parameterization scheme', Atmos. Chem. Phys., 6, 4361-4381, 2006

Discretize $\hat{c}$ using $n_{\hat{c}}$ phasespeeds

$$\hat{c}_{\min} = 0.25 \text{ m/s}; \; \hat{c}_{\max} = 100 \text{ m/s}; \; n_{\hat{c}} = 20 - 50$$

Co-ordinate stretch applied: higher resolution at large-c (i.e. small-m)

PROCEDURE

1) Check for critical level absorption, i.e. if U increases such that $\hat{U}(z_1) > \hat{U}(z_0)$ then waves with phase speeds in the range $\hat{U}(z_0) \le \hat{c} \le \hat{U}(z_1)$

2) Phase speeds which survive critical level absorption propagate conservatively to next level

3) Possible nonlinear dissipation is modelled by limiting the momentum flux

$$\rho\overline{F}(\hat{c},\varphi) <= \rho\overline{F}^{sat}(\hat{c},\varphi)$$

1) Repeat procedure for subsequent layers and all azimuths

2) Results in momentum flux profiles used to derive the net eastward, $\rho\overline{F}_E$, and northward, momentum flux $\rho\overline{F}_N$

3) The wind tendency (i.e. gravity wave drag) in each of these directions is given by the vertical divergence of the momentum flux

$$\frac{\partial U,V}{\partial t} = g\frac{\partial \rho\overline{F}_E, \rho\overline{F}_N}{\partial p}$$

Evaluation

ensemble of T159 (125 km) 1-year climate runs and compare mean circulation and temperature structure against SPARC dataset

- IFS before 2009: Uses so called Rayleigh friction, a friction proportional to the zonal mean wind speed, to avoid unrealistically high wind speeds (polar night jet) in middle atmosphere.

- Non-orographic GWD introduced in Cy35r3 (September 2009)

January

NH

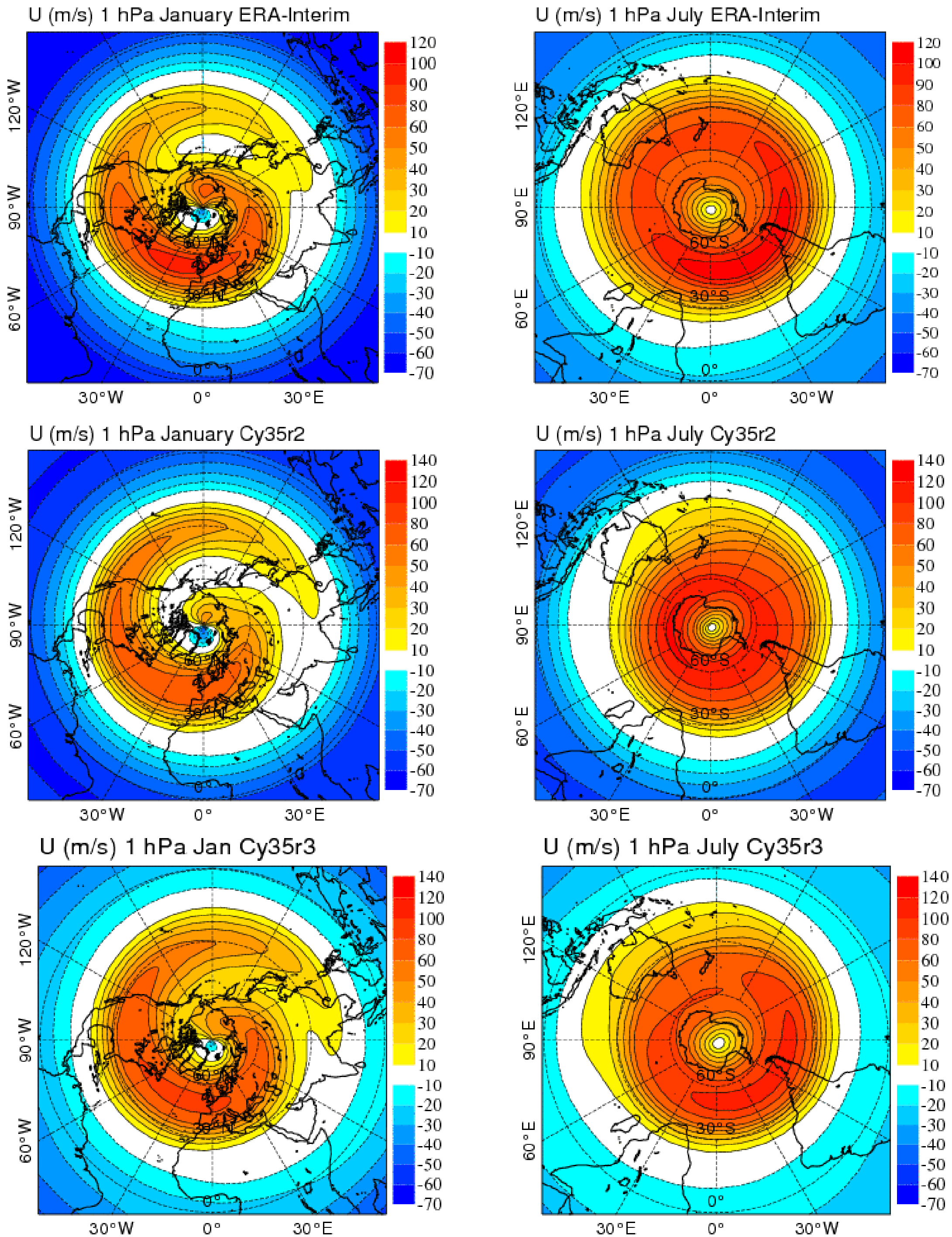
ERA

Cy35r2

Operational since March 2009

Cy35r3

Operational in summer 2009 with GWD + GH(



July

SH

Polar winter vortex

SH wintertime vortex is quasi-symmetric, but not NH polar vortex, due to braking quasi-stationary Rossby waves emanating in the troposphere

The Eliassen Palm flux vectors (D.G Andrews 1987)

describing the action of resolved waves on the mean flow

$$EPVector = \left\langle -R\cos\phi\,\overline{u^{*}v^{*}},\ f\,R\cos\phi\,\overline{v^{*}\theta^{*}}\left(\partial\theta/\partial p\right)^{-1}\right\rangle$$

$$f = Coriolis; \quad \phi = latitude; \quad \theta = pot.temperature$$

$$*\,denote\ anomalies\ \ from\ zonal\ mean$$

$$\left(Peixoto\,\&\,Oort\ \ 1992,\ p.388\right)$$

- EP Flux vectors give the net wave propagation for stationary Rossby waves (derived from quasi-geostroph. eq.)
- Stationary Rossby waves are particularly prominent in the NH during winter. They propagate from the troposphere upward into the stratosphere

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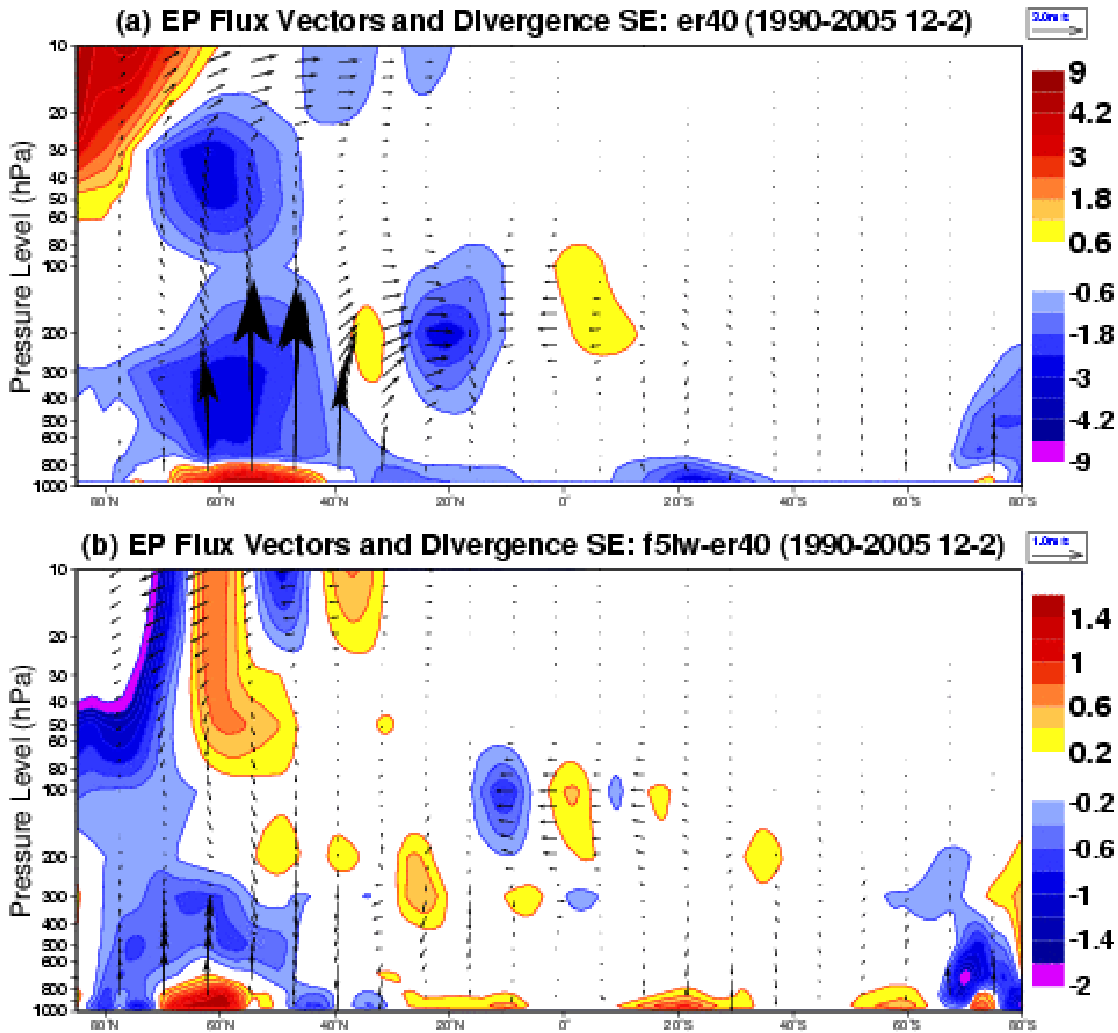
EP-Fluxes in Winter

ERA40

Cy35r3-ERA40



Stationary Rossby waves are particularly prominent in the NH during winter. They propagate from the troposphere upward into the stratosphere

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EP-Fluxes in Summer

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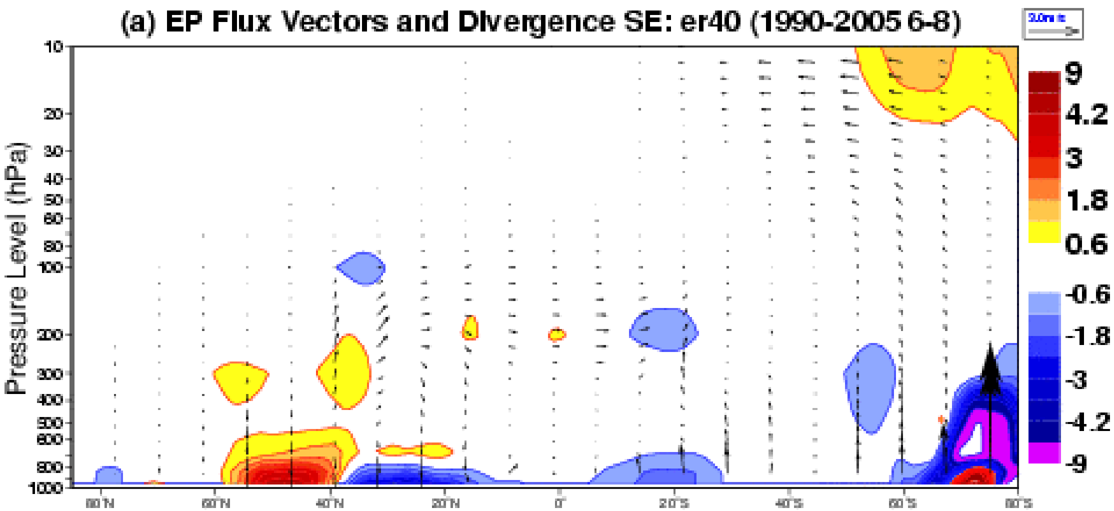
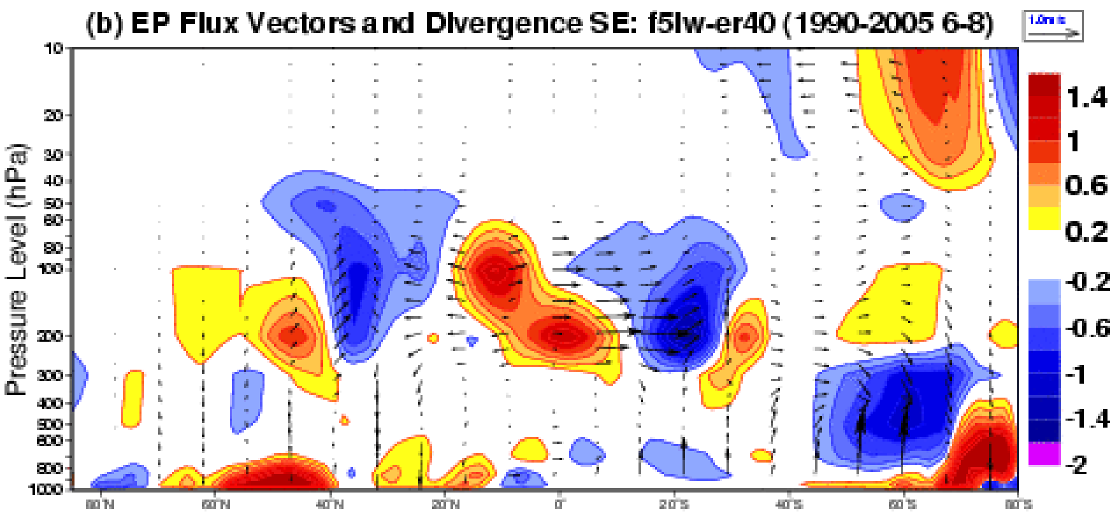


Cy35r3-
ERA40



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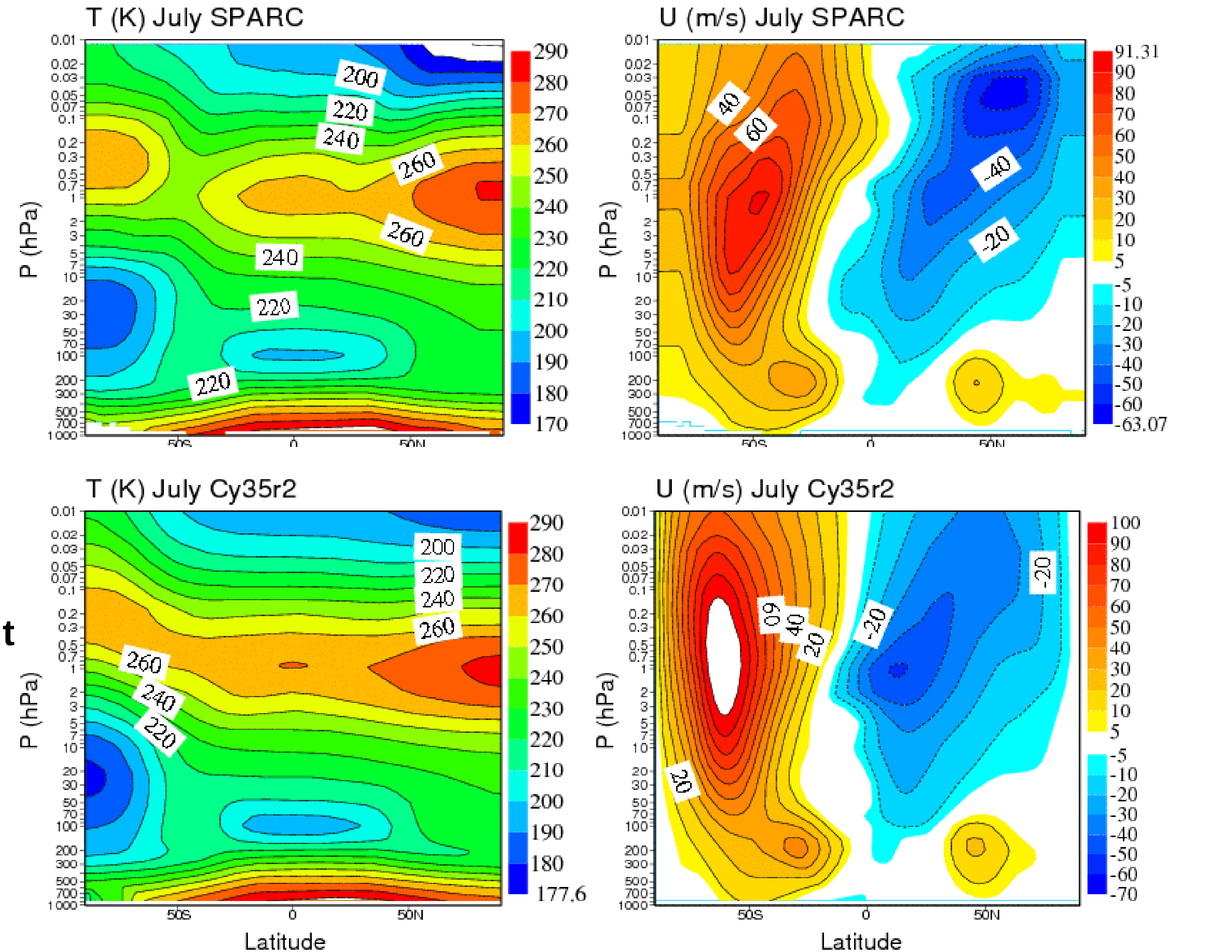
July climatology

SPARC

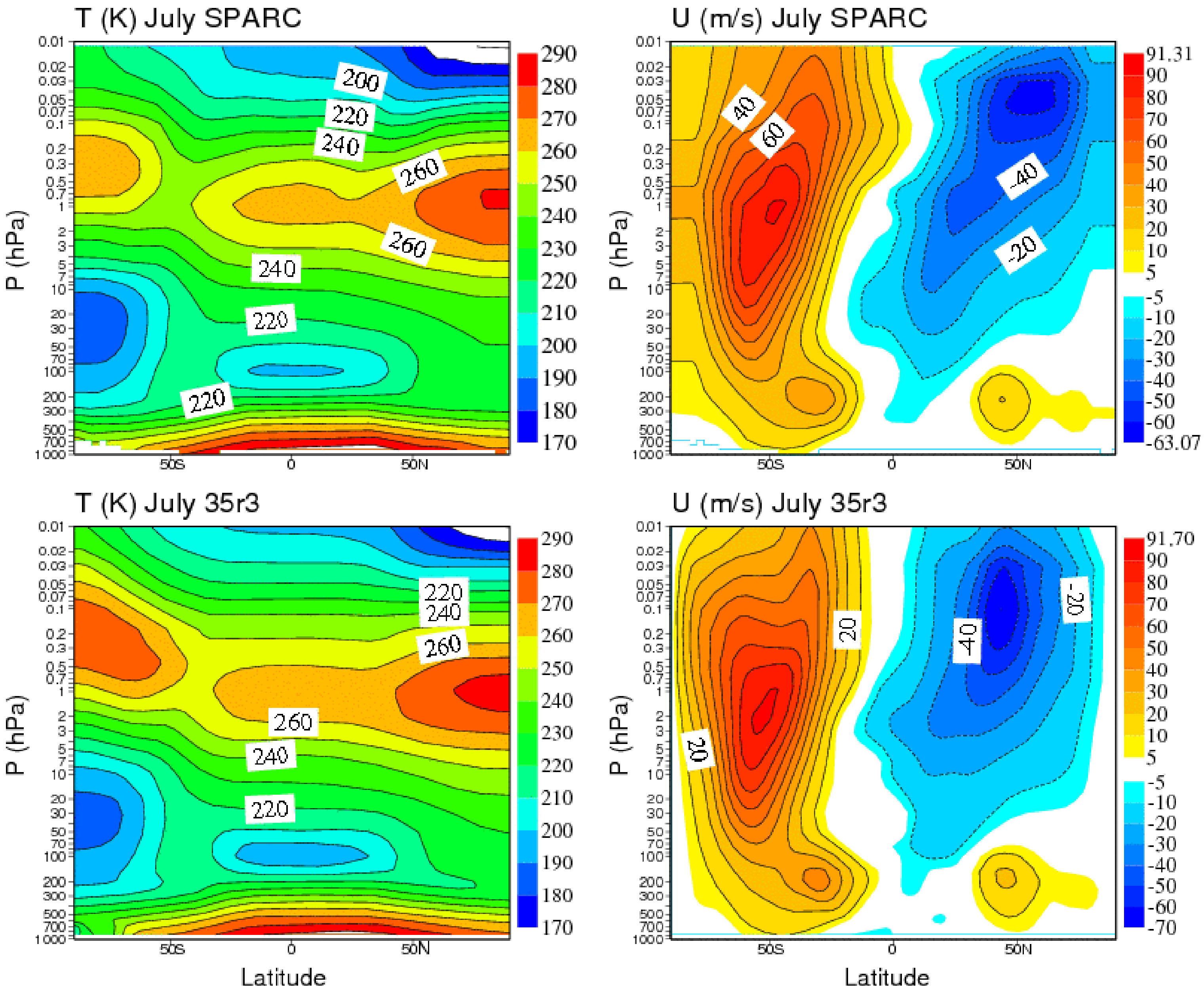
no GWD param, but Rayleigh friction



July climatology

SPARC

35r3 with GWD scheme



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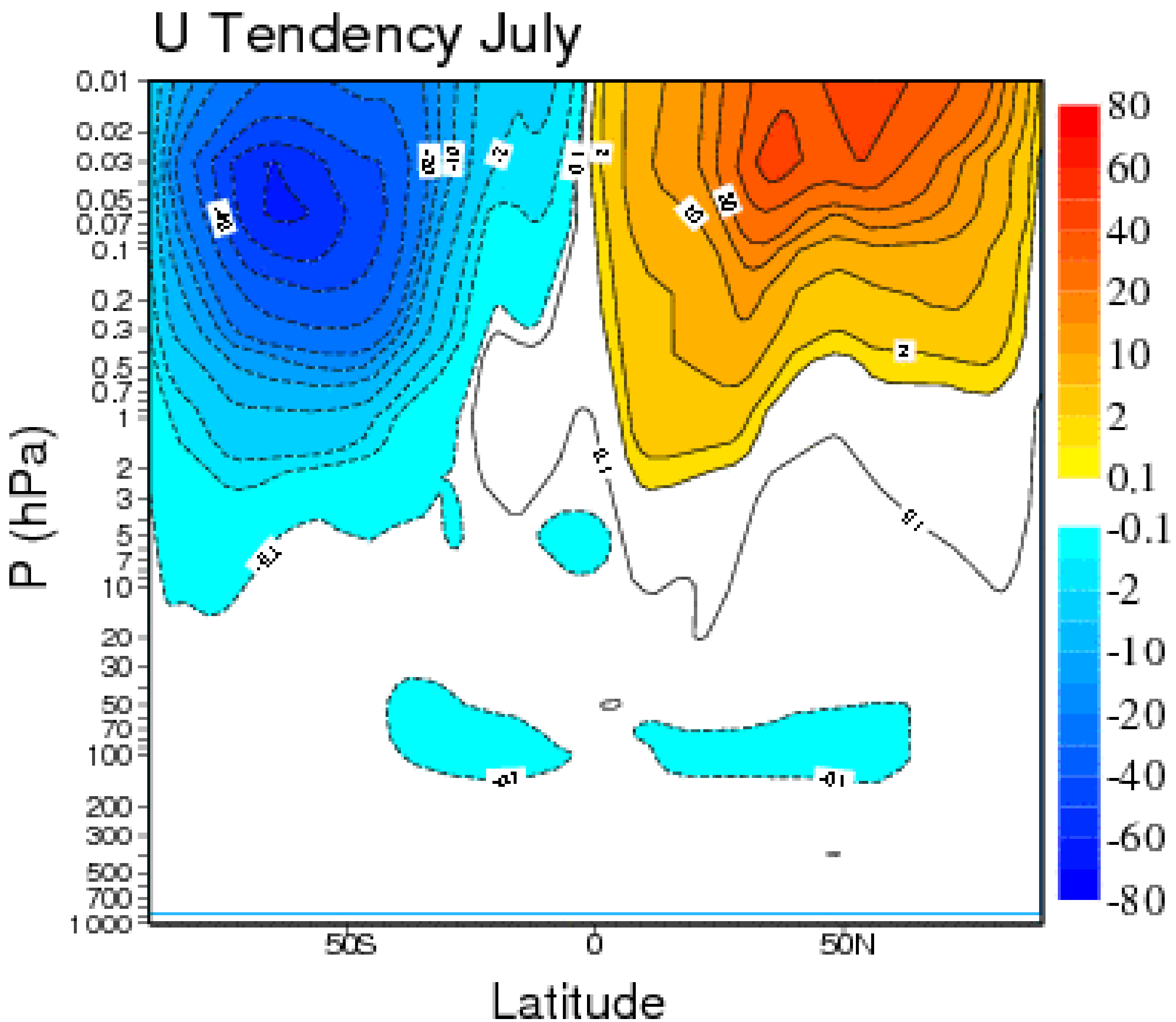
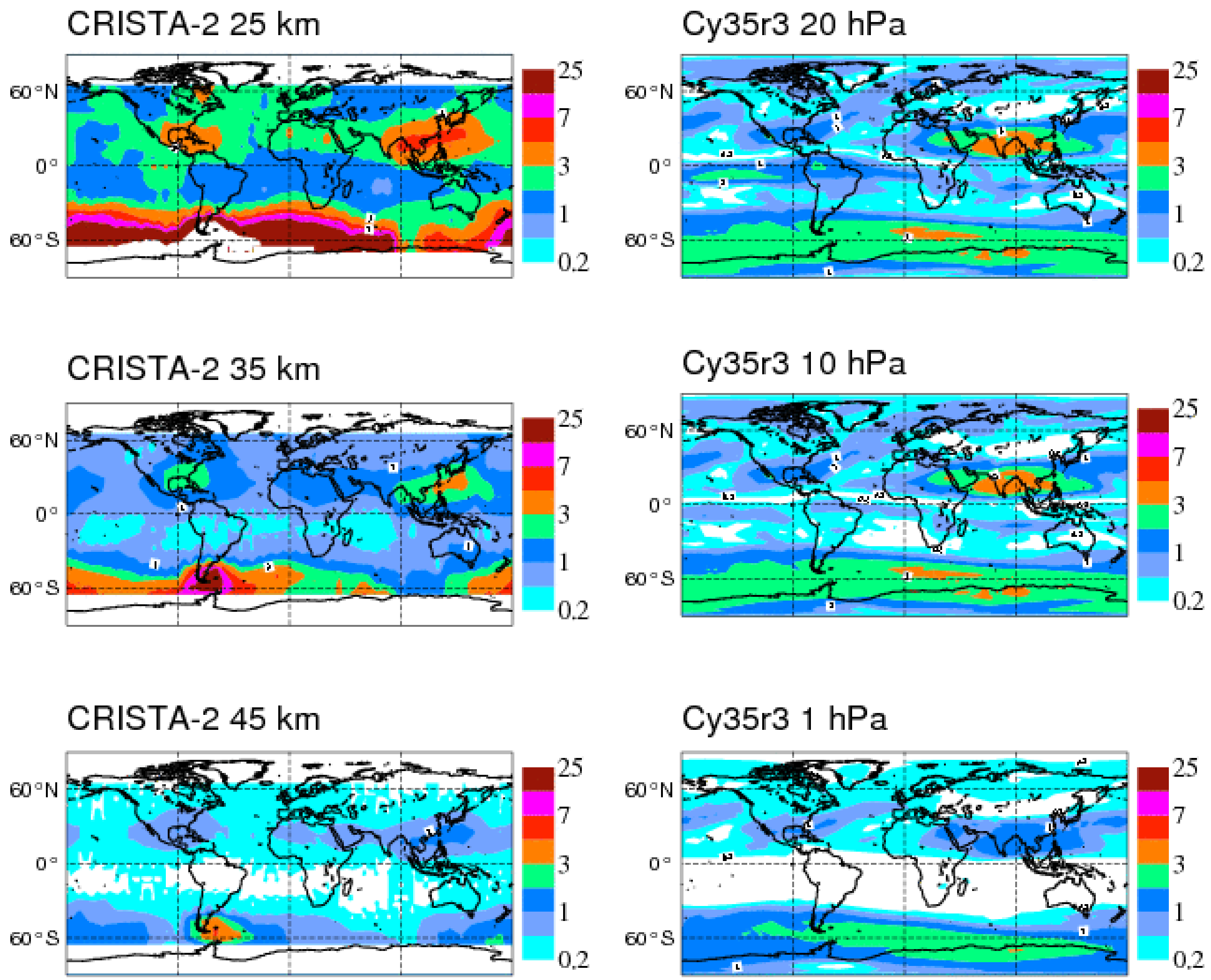
U Tendencies (m/s/day) July from non-oro GWD

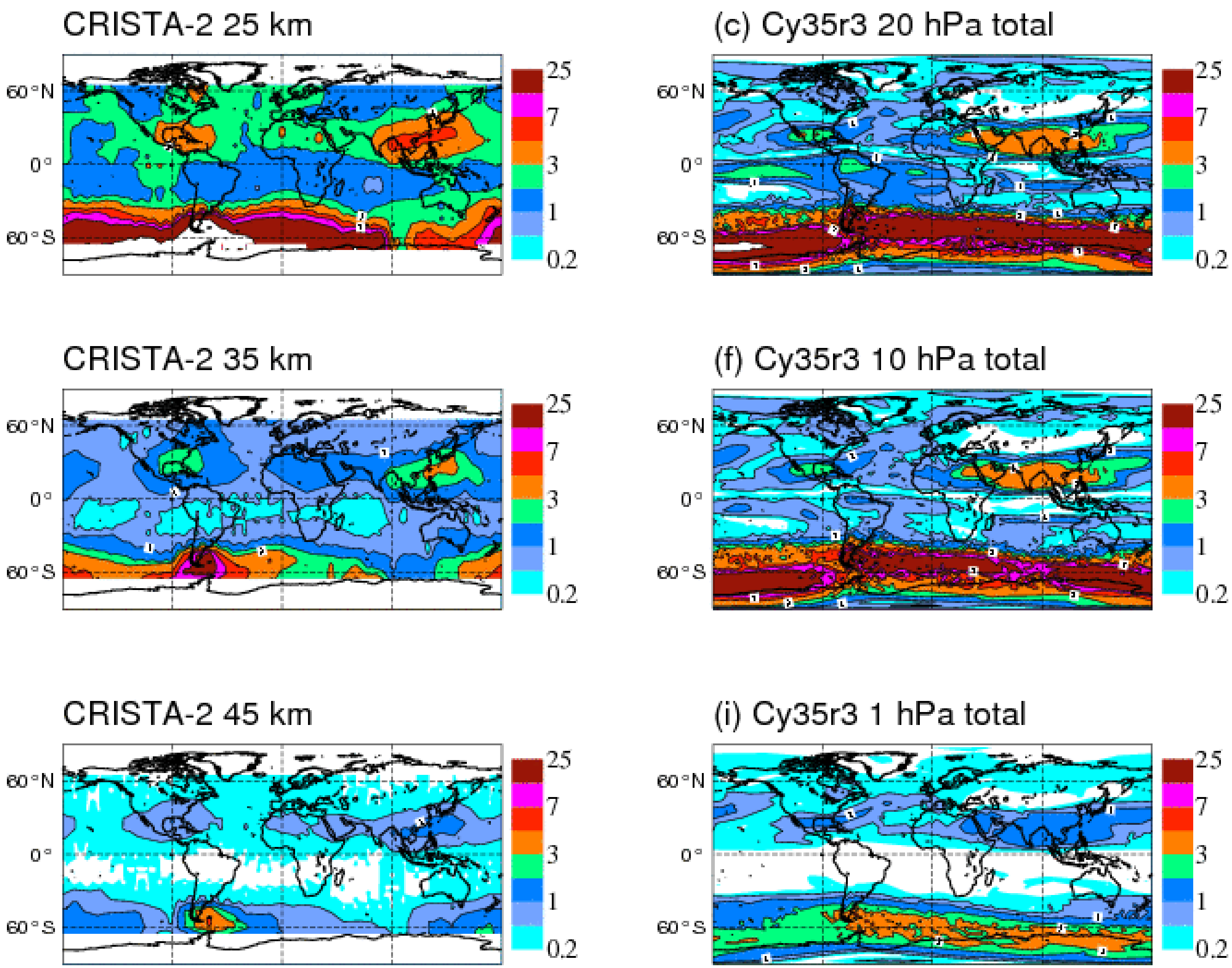


Momentum fluxes from CRISTA campaign compared to parametr non-orogr GW momentum fluxes



Comparison of observed and parametrized GW momentum flux for 8-14 August 1997 horizontal distributions of absolute values of momentum flux (mPa) Observed values are for CRISTA-2 (Ern et al. 2006). Observations measure temperature fluctuations with infrared spectrometer, momentum fluxes are derived via conversion formula.

Total = resolved +parametrized (orograph+non-orograph.) wave momentum flux



Conclusions from comparison against SPARC & ERA-Interim reanalysis

- Polar vortex during SH winter quasi symmetric, but asymmetric NH winter polar vortex, due to vertically propagating quasi-stationnary Rossby waves (linked to mountain ranges)
- Without GWD parameterization SH polar vortex too strong, westerly polar night Jet is wrongly tilted with height, large T errors in mesosphere. Jet maximum in summer hemisphere easterly jet at wrong height (at stratopause instead of mesopause)
- Results qualitatively similar for January, invert NH and SH

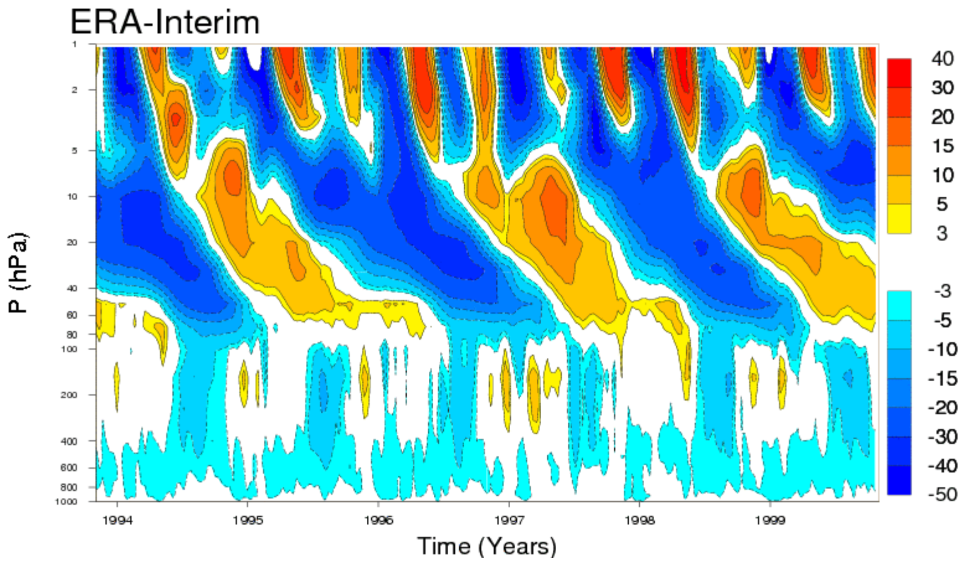
The QBO

Prominent oscillations in the tropical middle atmosphere are

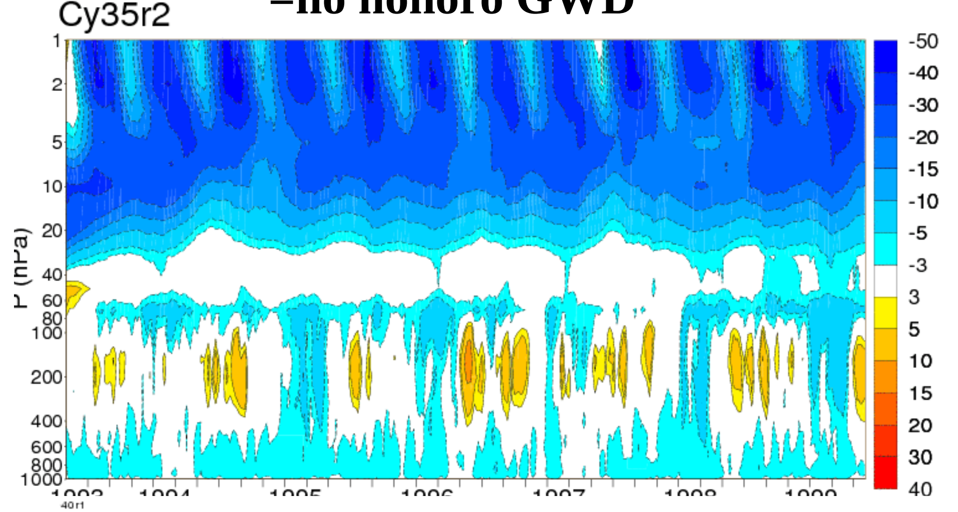
- A quasi bi-annual oscillation in the stratosphere, and a
- Semi-annual oscillation in the upper stratosphere and mesosphere

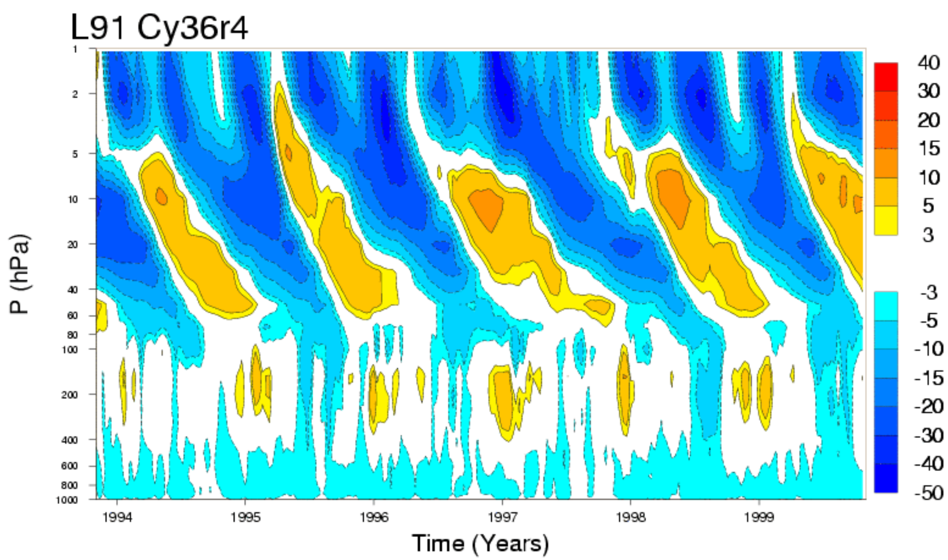
These oscillations are wave induced. Whereas the waves are moving upward, these oscillations propagate downward. Why ? Waves deposit momentum at critical level, wind changes, and so does the critical level, etc

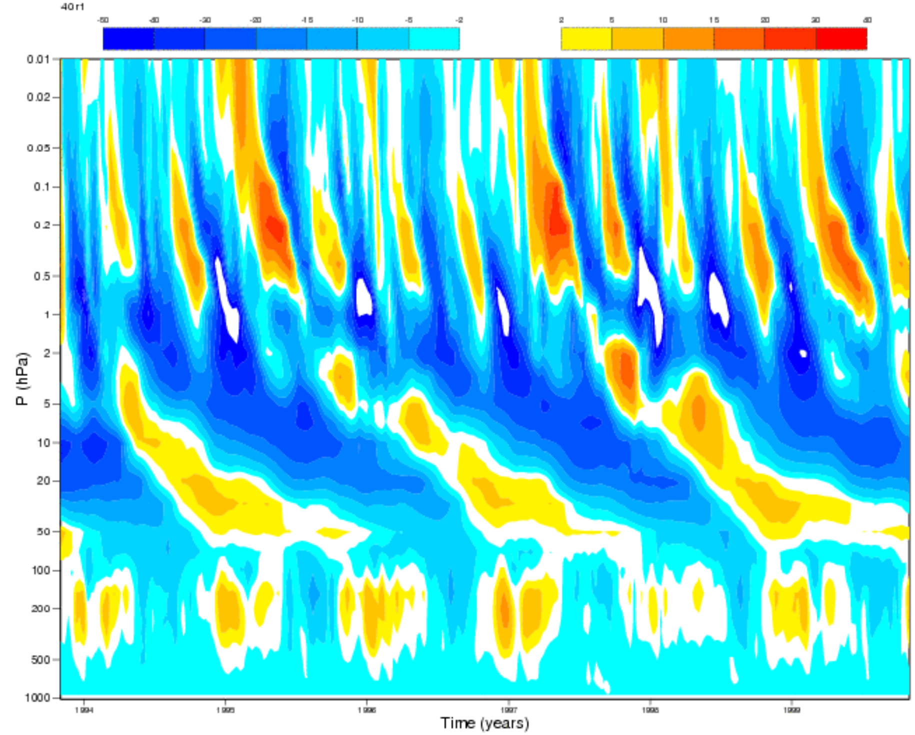
QBO : Hovmöller U from free 6y integrations



=no nonoro GWD

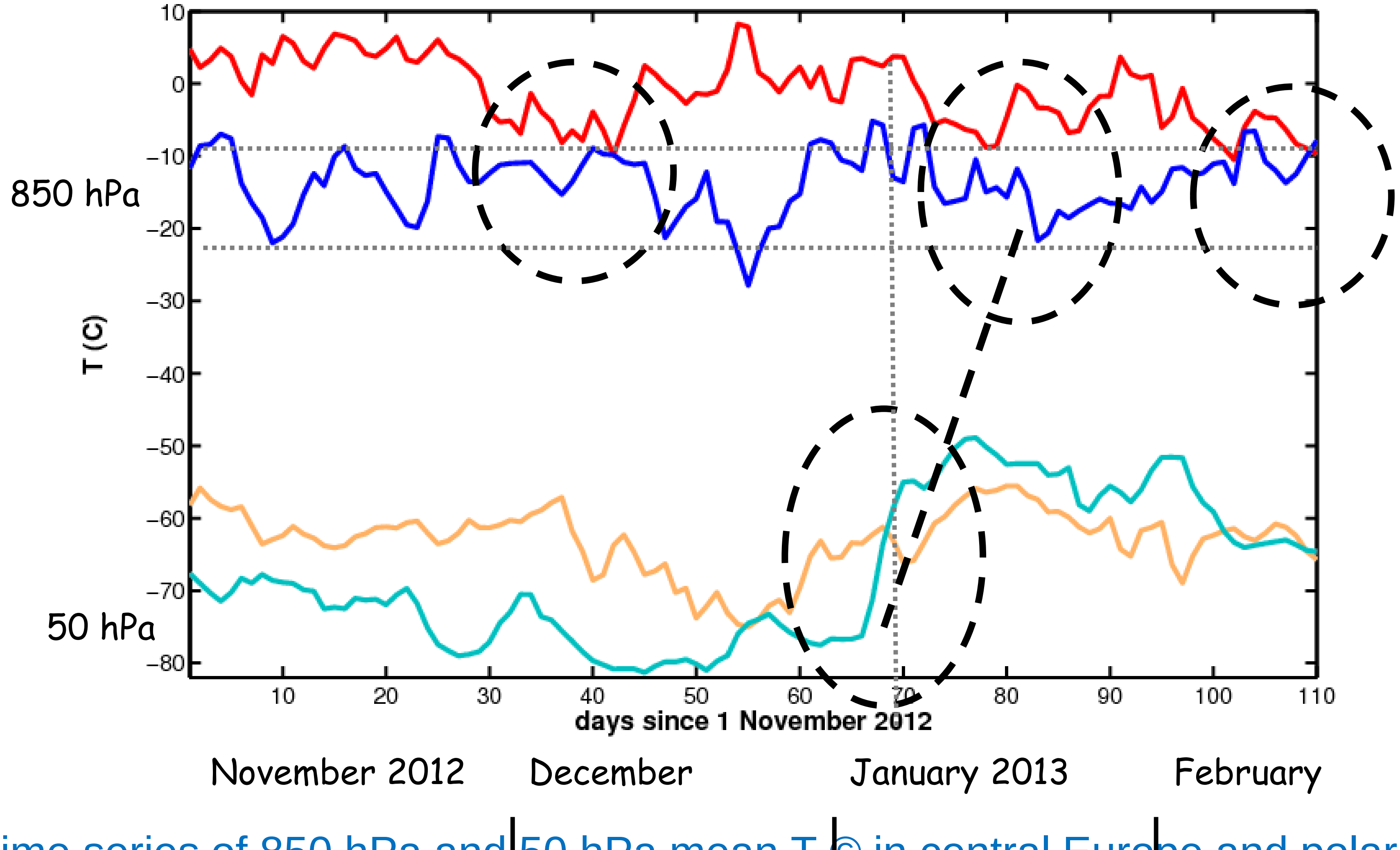






L137 Cy40r1 (2014)

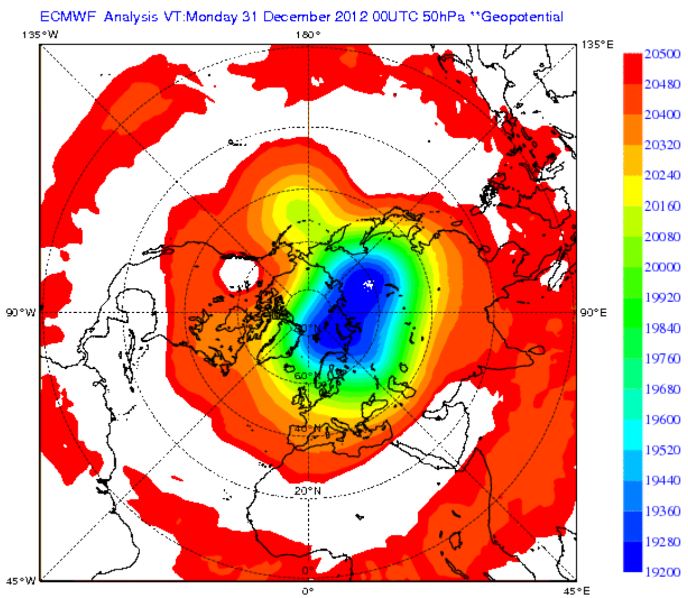
Sudden Stratospheric Warming=SSW



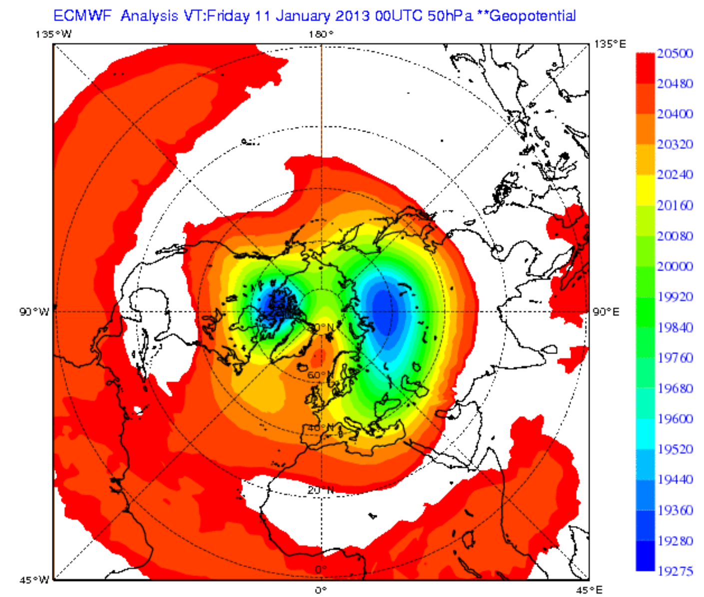
Time series of 850 hPa and 50 hPa mean T © in central Europe and polar Europe - cold spells & Sudden Stratospheric Warming in January

SSW and the polar stratospheric Vortex

31.12.2012 Z 50 hPa



11.1.2013 Z 50 hPa



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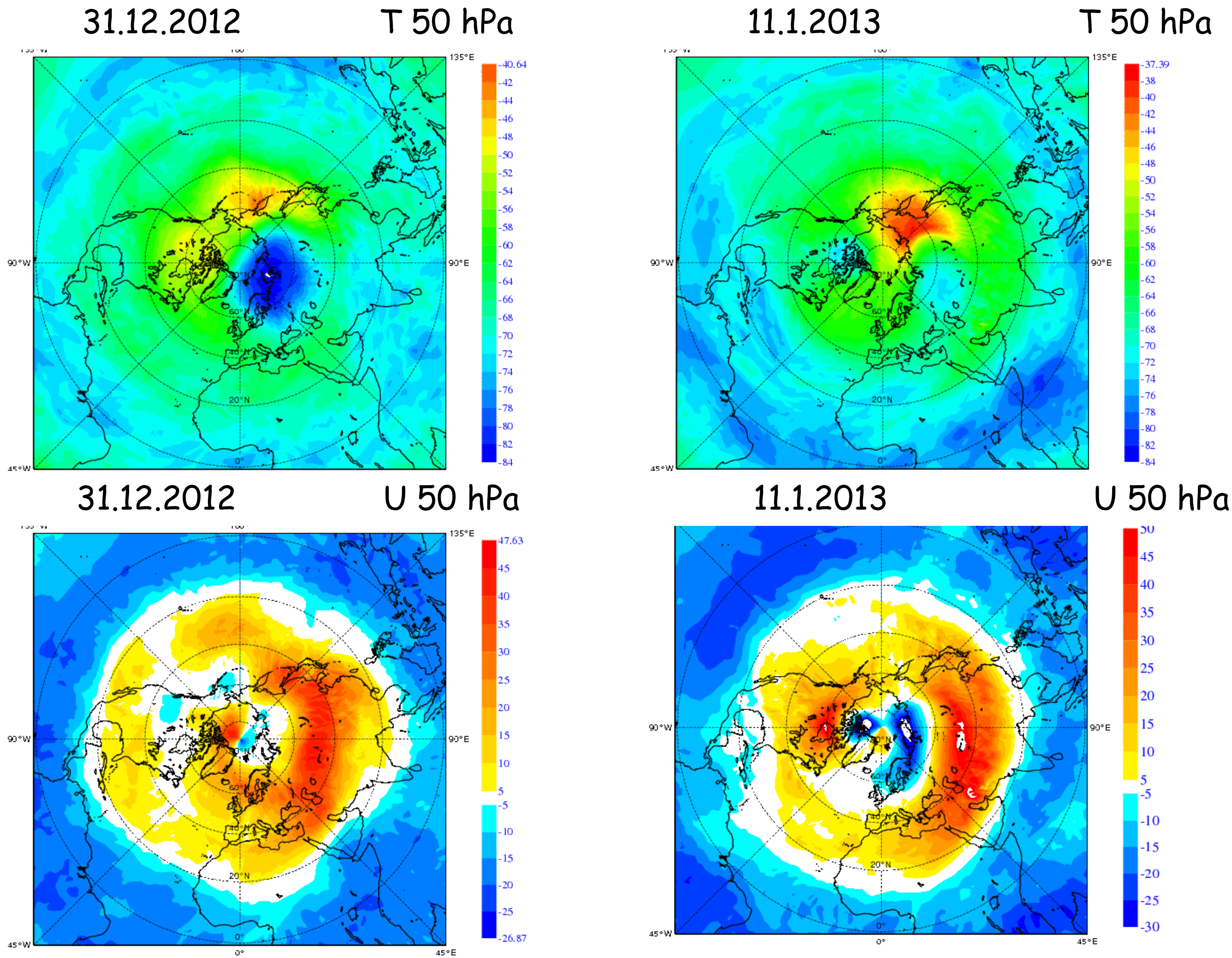
SSW with wind reversal



The stratospheric warming precedes the low-level, 850hPa cooling by 5-10 days