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On Influence of a variation of heating sources on structure baroclinic turbulence and thermal stratification of the extratropical troposphere in simplified GCM.

V.Krupchatnikov(1),I.Borovko(2),Yu.Martynova(1)

(1 - The Siberian hydrometeorological institute of Federal Hydrometereology and Environmental Monitoring Service, Novosibirsk,
2 - Institute of computational mathematics and mathematical geophysics SB RAS)

Contents

An Intermediate Complexity General Circulation Model (ICGCM) with prescribed heating

Dynamical Core

$$\frac{\partial \zeta}{\partial t} = \frac{1}{1-\mu^2}\frac{\partial}{\partial \lambda}F_v - \frac{\partial}{\partial \mu}F_u - \frac{\xi}{\tau_f} - k(-1)^n \nabla^{2n}\xi$$

$$\frac{\partial D}{\partial t} = \frac{1}{1-\mu^2}\frac{\partial}{\partial \lambda}F_u + \frac{\partial}{\partial \mu}F_v - \nabla^2\left(\frac{U^2+V^2}{2(1-\mu^2)} + \Phi + T_R \ln p_s\right) - \frac{D}{\tau_f}$$
$$- k(-1)^n \nabla^{2n} D$$

$$\frac{\partial T'}{\partial t} = -\frac{1}{1-\mu^2}\frac{\partial}{\partial \lambda}(uT') - \frac{\partial}{\partial \mu}(vT') + D\cdot T' - \dot{\sigma}\frac{\partial T}{\partial \sigma} + \kappa\frac{T\omega}{p}$$
$$+ \frac{T_R - T}{\tau_R} - k(-1)^n \nabla^{2n} T'$$

Heat forcing

$$\frac{\partial \ln p_s}{\partial t} = -\frac{U}{1-\mu^2}\frac{\partial \ln p_s}{\partial \lambda} - V\frac{\partial \ln p_s}{\partial \mu} - D - \frac{\partial \dot{\sigma}}{\partial \sigma}$$

Dynamical Core (cont.)

$$\frac{\partial \Phi}{\partial \ln \sigma} = -T$$

$$\frac{\omega}{p} = \vec{V} \cdot \nabla \ln p_s - \frac{1}{\sigma}\int_0^\sigma \left(D + \vec{V} \cdot \nabla \ln p_s\right) d\sigma$$

$$U = u\sqrt{1-\mu^2}$$

$$F_u = V\zeta - \dot{\sigma}\frac{\partial U}{\partial \sigma} - T'\frac{\partial \ln p_s}{\partial \lambda}$$

$$V = v\sqrt{1-\mu^2}$$

$$F_v = -U\zeta - \dot{\sigma}\frac{\partial V}{\partial \sigma} - T'\left(1-\mu^2\right)\frac{\partial \ln p_s}{\partial \mu}$$

$$T_R(\sigma,\varphi) = T_r(\sigma) + h(\sigma)$$

Sensitivity to heating in a ICGCM

Troposphere is weakly stratified (to vertical displacements)

- Solar heating of the Earth's surface leads to a radiative equilibrium state that is dynamically unstable, either convectively (as in the tropics) or baroclinically (as in the extratropics).

- The heat transfer due to large-scale **turbulent baroclinic motion**, both vertical and meridional, extend to region of finite depth that we may consider to be the troposphere

Stratosphere: The radiative equilibrium state T_{rad} is dynamically stable and departures from this state occur only through external forcing by waves propagating up from the troposphere.

Atmospheric waves transfer angular momentum and energy (but not heat) from the surface of the Earth and the troposphere into the region above.

In the stratosphere, the negative wave drag from planetary-scale Rossby waves drives an equator-to-pole mass circulation

Mass conservation then demands upwelling in the tropics and downwelling in the extratropics. This vertical motion leads to adiabatic heating or cooling which is balanced, respectively, by radiative cooling or heating.

Simulation scenario

By means of system of the atmosphere dynamics equation with zonally symmetric forcing sensitivity of circulation of extratropical troposphere to thermal indignations of a polar stratosphere is investigated.

A thermal source is set in the form of Newton with the set equilibrium profile of temperature which only depends on latitudes and pressure

$$T_R\ (\sigma,\varphi) = T_r(\sigma) + h(\sigma) \qquad T_r\ (\sigma) = (\Gamma_{\max} - \Gamma) H \ln\left(\frac{\sigma}{\sigma_T}\right)$$ **- Radiative equilibrium temperature**

$$h(\sigma,\varphi) = \begin{cases} \sin\frac{\pi}{2}\left(\frac{\sigma-\sigma_T}{1-\sigma_T}\right)\left(\Delta T_{сю}\frac{\mu}{2} - \Delta T_{эп}\left(\mu^2 - \frac{1}{3}\right)\right) & \sigma > \sigma_T \\ \omega(\varphi)\Gamma H \ln\left(\frac{\sigma}{\sigma_T}\right) & \sigma < \sigma_T \end{cases}$$

In the stratosphere

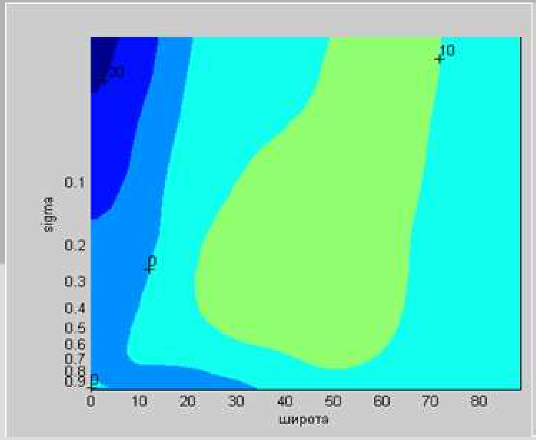
$$T_r(\sigma) = T_{tr} + (\Gamma_{\max} - \Gamma) H \ln\left(\frac{\sigma}{\sigma_T}\right)$$

where T_{tr} =210K.

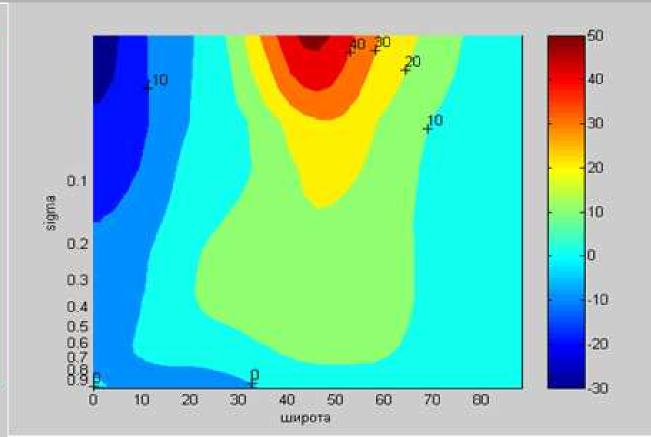
In this equation, the parameter Γ defines a temperature gradient.

There great values of a gradient of temperature radiating balance and more intensive Newton cooling into stratosphere correspond to greater values Γ.

Two experiments were conducted, in the first one, Γ was equal to 0 (weak polar vortex and in the second, Γ was equal to 4 (strong polar vortex).



a)



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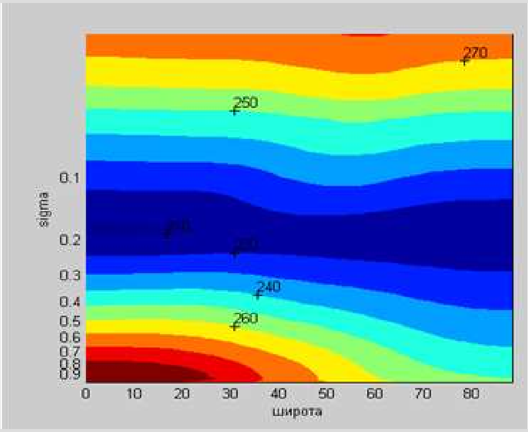
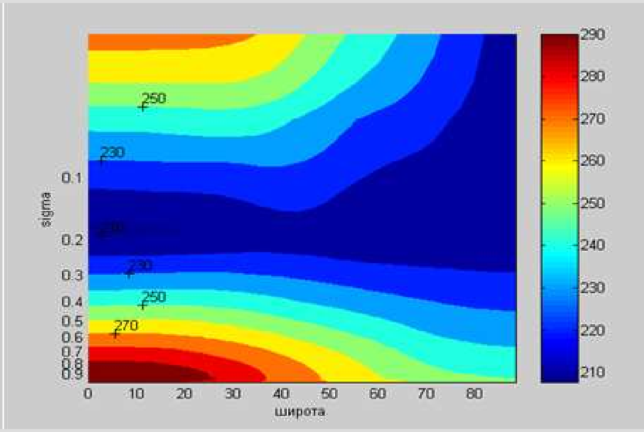
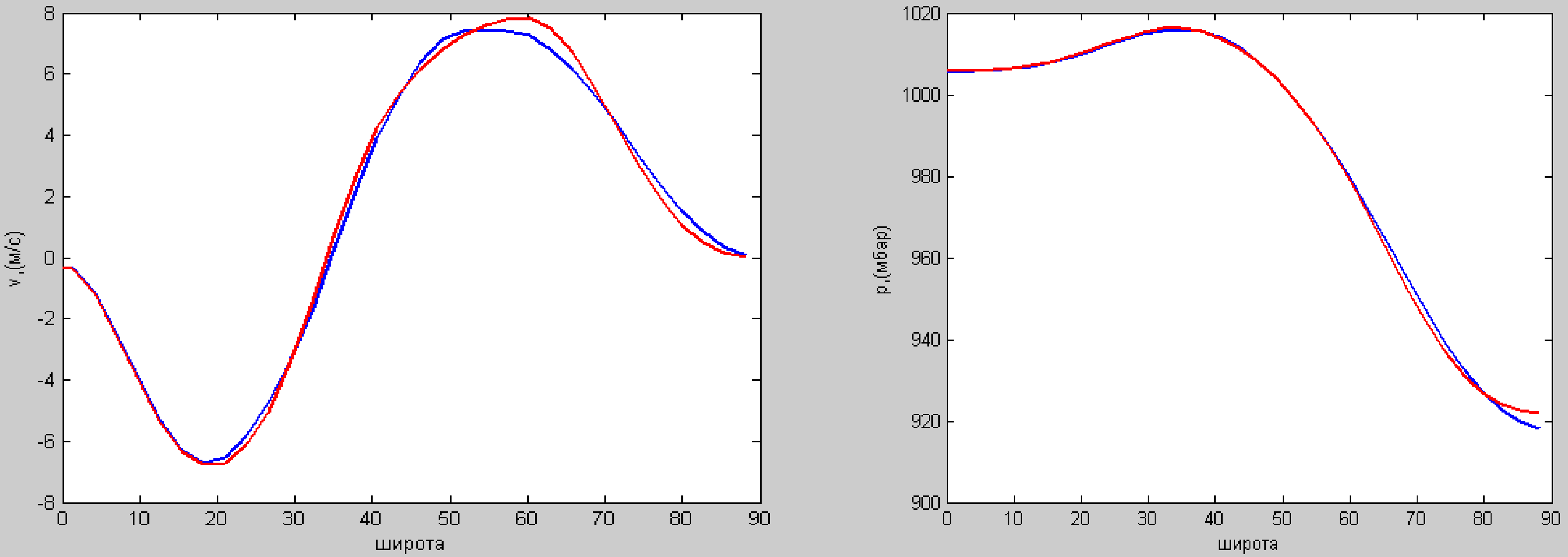


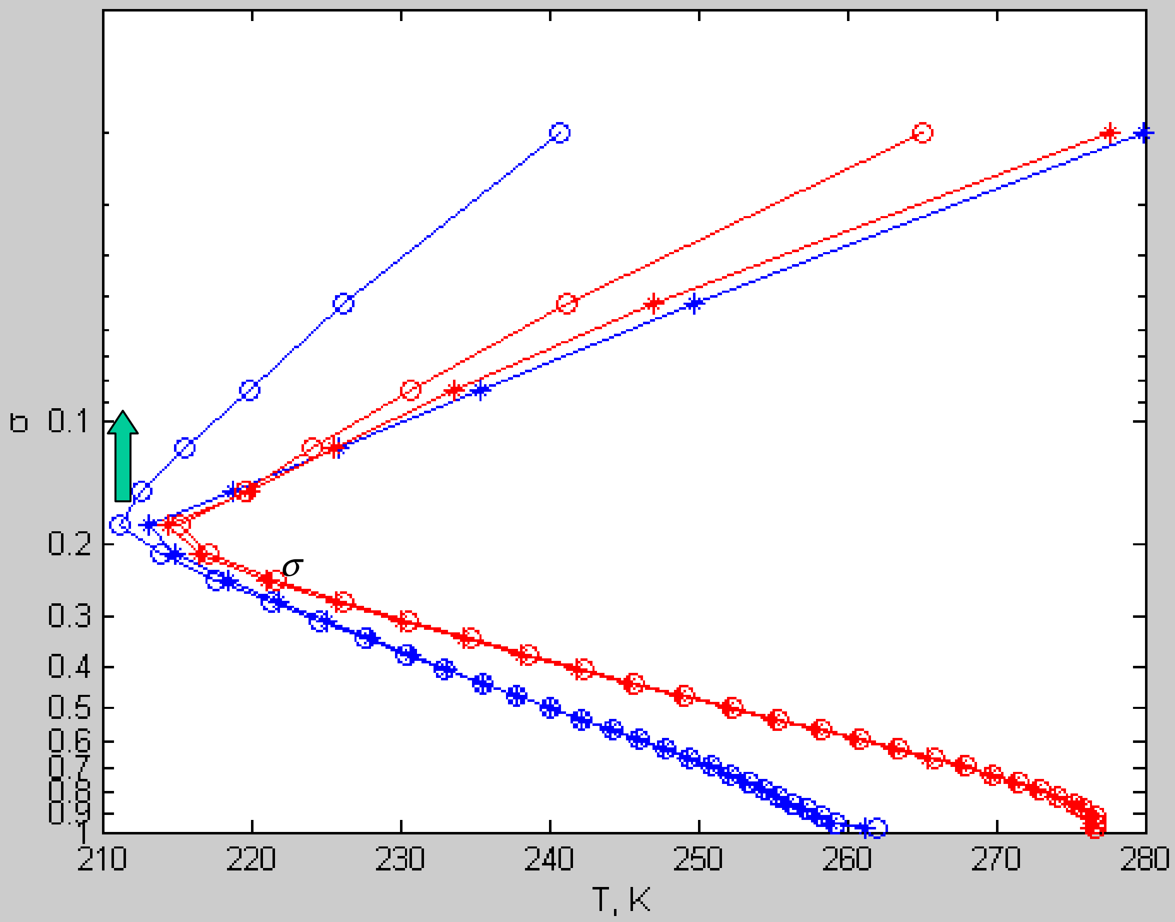


c) d)

Meridional cross sections for zonal wind velocities and mean zonal temperature. a) and b)- zonal velocities for cases of weak and strong vortex; c) and d) – mean zonal temperature



Zonal mean surface wind (м/sec), and pressure (hpa) in NH. Blue – Γ = 0, red – Γ = 4 (strong vortex)



Temperature profiles

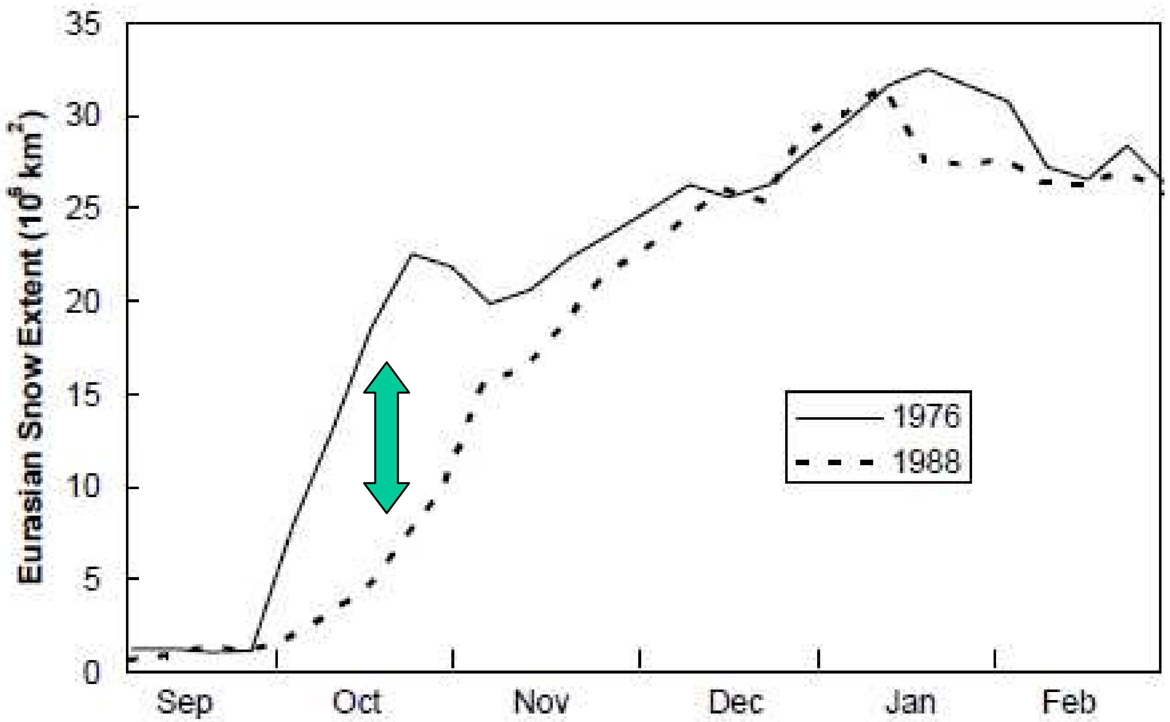
Blue line – averaged around 60^0 latitude belt, red line – averaged around 40^0 latitude belt; «stars» - $\Gamma = 0$; «circles»- $\Gamma = 4$

Can the Stratosphere Control the Extratropical Circulation Response to Surface Forcing?

1. Is snow active or passive in driving seasonal variability of the winter tropospheric circulation?

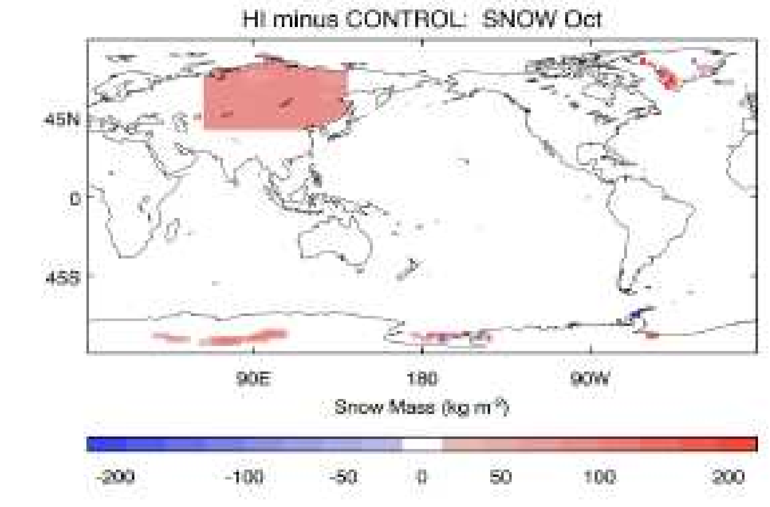
2. Can fall-season snow drive upward propagating wave activity (WAF) from the surface into the stratosphere?

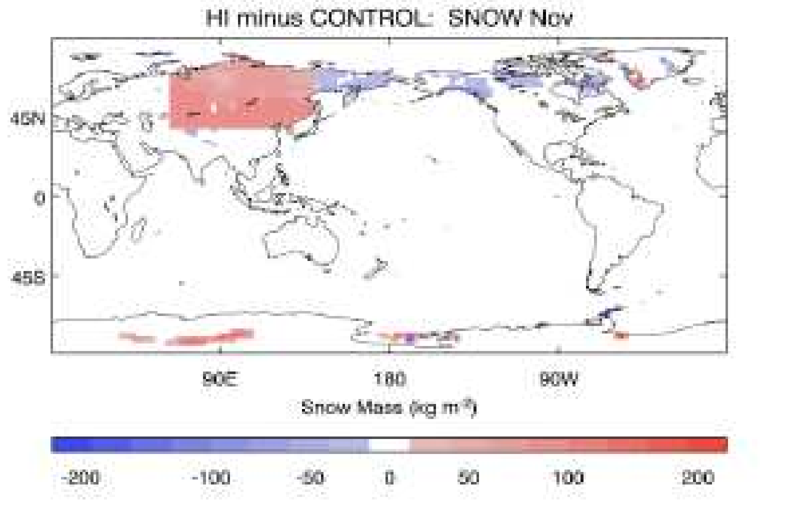
Surface albedo response?!

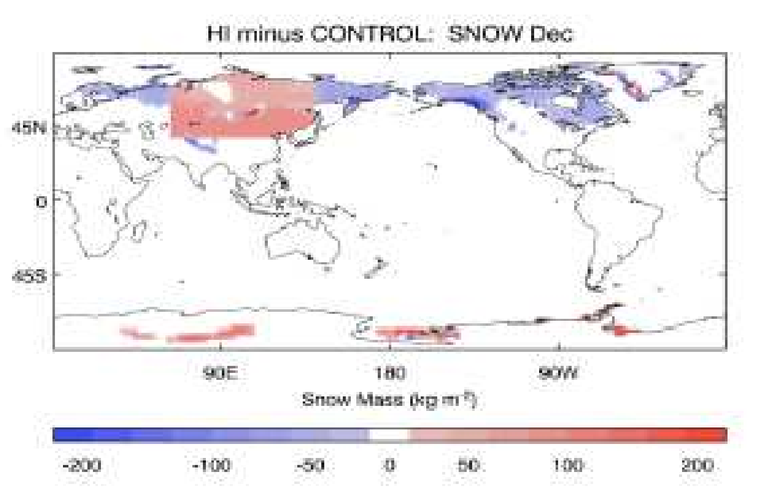


Weekly timeseries of NOAA satellite-observed snow cover extent over Eurasia, for the period September 1976 – February 1977 (solid line) and September 1988 – February 1989 (dashed line).

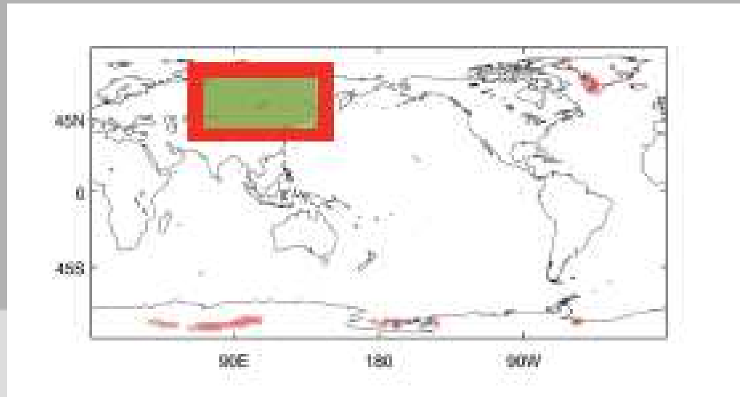
RESPONSE = HI minus CONTROL (ensemble means).

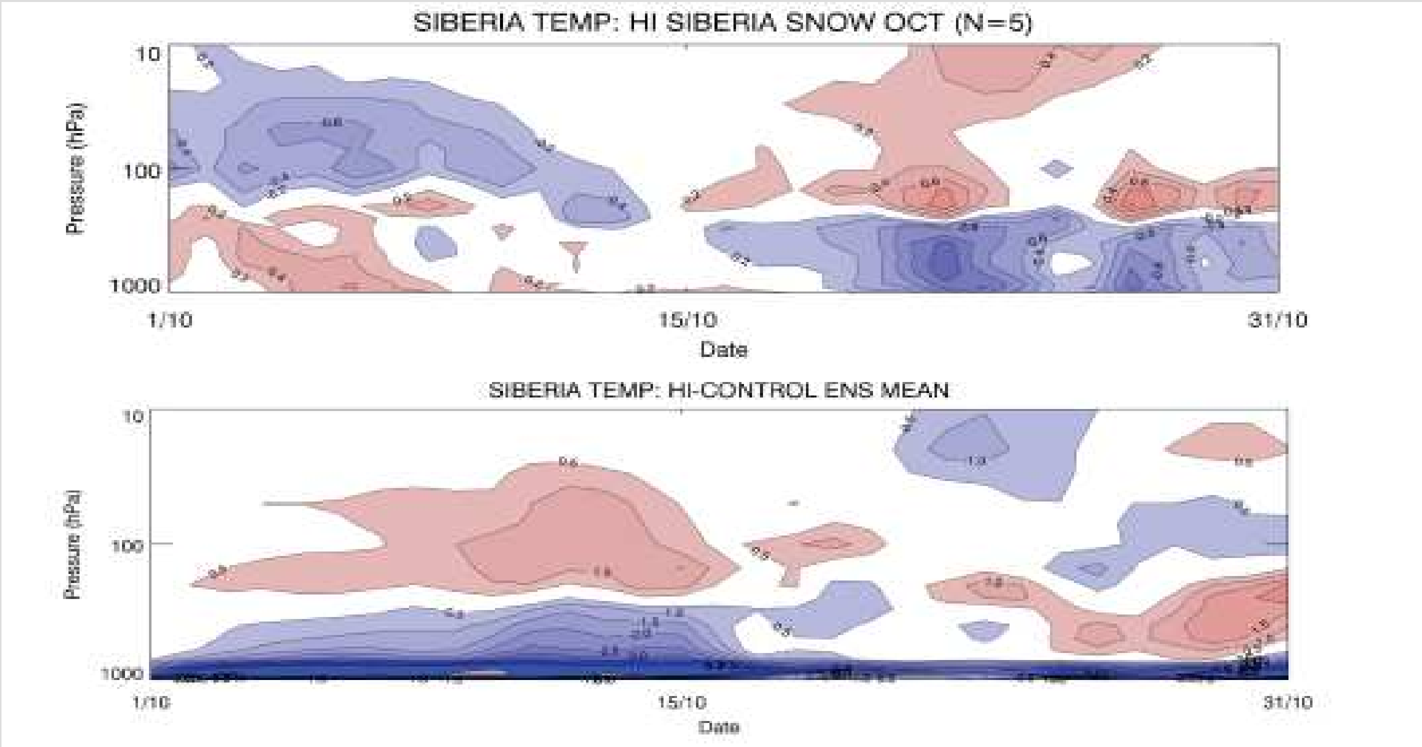


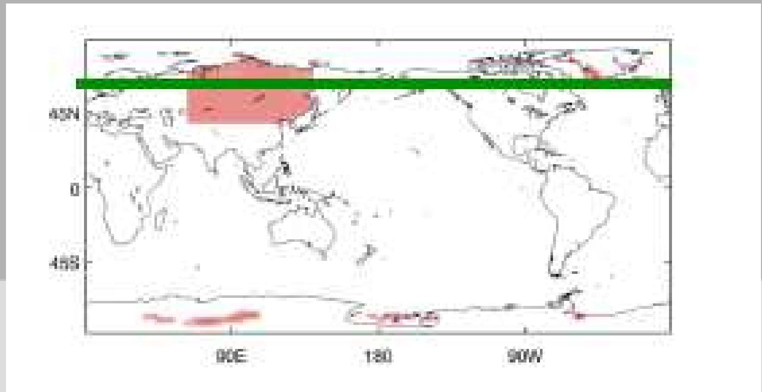




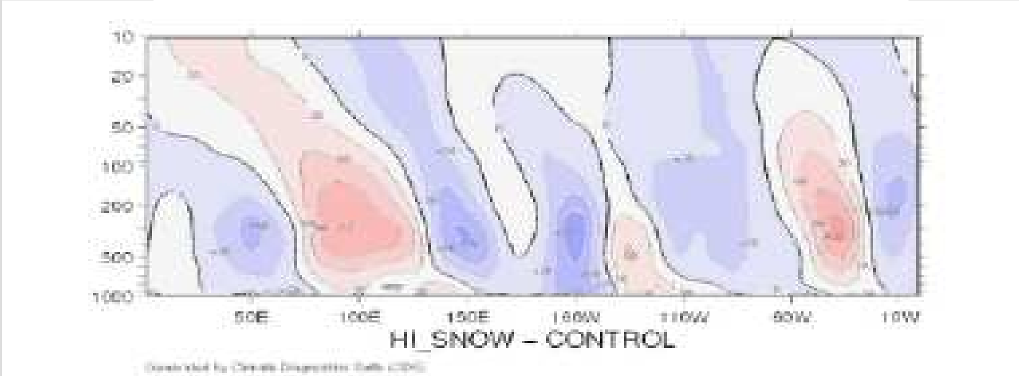
(C. Fletcher, P. J. Kushner, 2006)


45N
0
45S
90E
180
90W

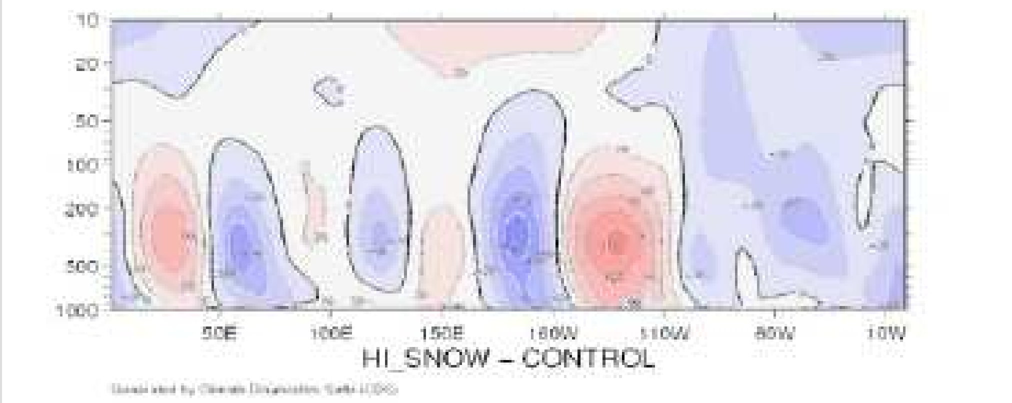

SIBERIA TEMP: HI SIBERIA SNOW OCT (N=5)
Pressure (hPa)
10
100
1000
1/10
15/10
31/10
Date
SIBERIA TEMP: HI-CONTROL ENS MEAN
Pressure (hPa)
10
100
1000
1/10
15/10
31/10
Date



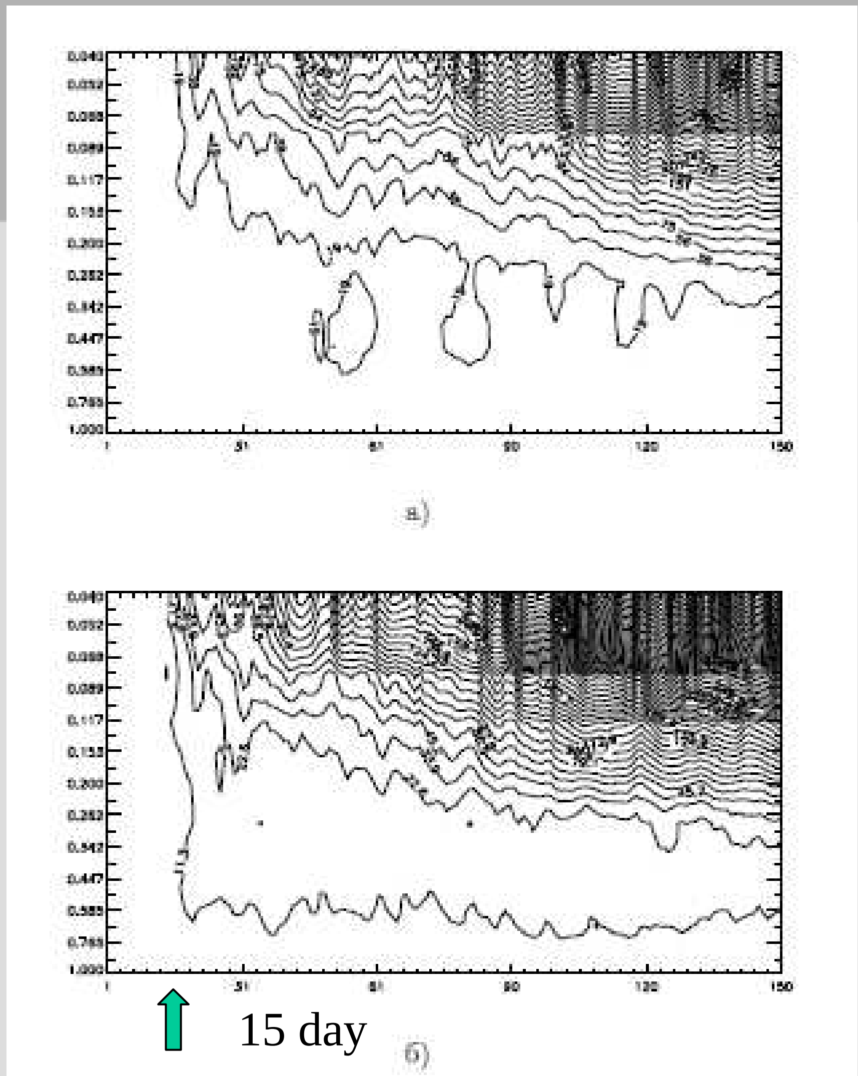
Cross-section: Z 60N



Day 15



Day 27



Anomaly of geopotential height avareged over 60 – 85 lat. belt. Snow forcing begins Oct 1st, strat-trop interaction is associated with WAF: (a) for “max” – ensemble (positive anomaly snow mass); (b) for “min” - ensemble (negative anomaly snow mass) (Y. Martynova, V.Krupchatnikov, 2010)

Conclusions

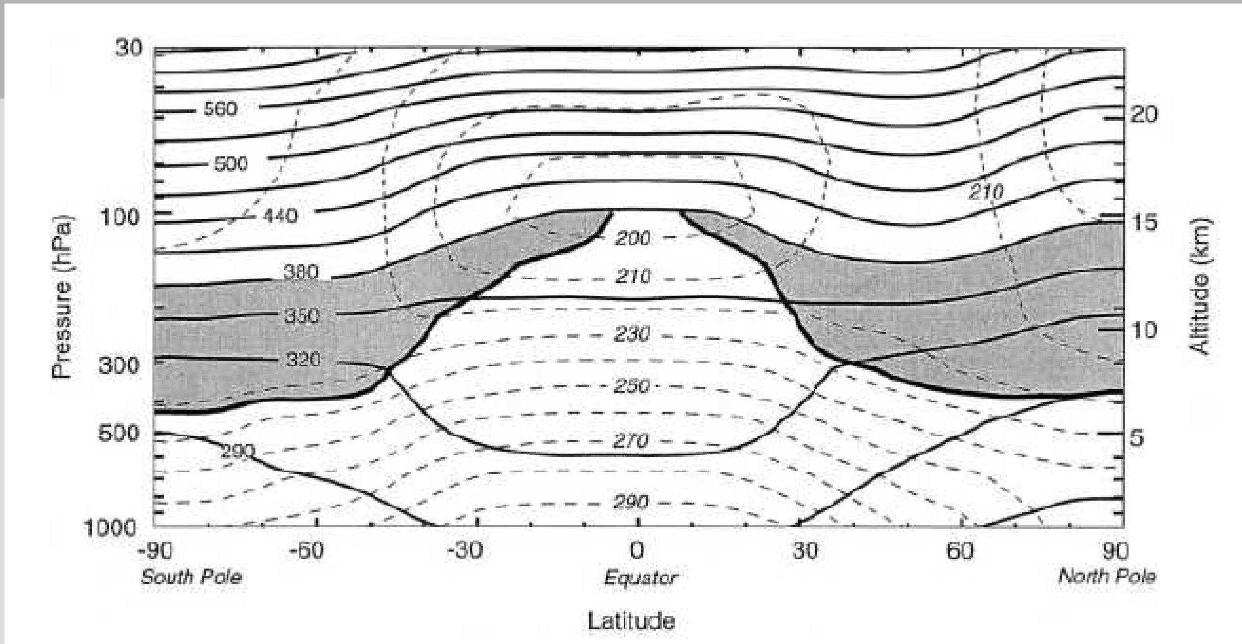
1. Is snow active or passive in driving seasonal variability of the winter tropospheric circulation?

- Snow forces atmospheric response.
- Local anomalies damped by circulation response after ~15 days
- **Low zonal wavenumber eastward moving wave trains**.

2. Can autumn snow drive upward propagating wave activity from the surface into the stratosphere?

- Still not clear. Analysis of WAF response in progress…
- Problems: **weather noise (baroclinic turbulence), model sensitivity at midlatitudes**.

Stratification and baroclinic turbulence



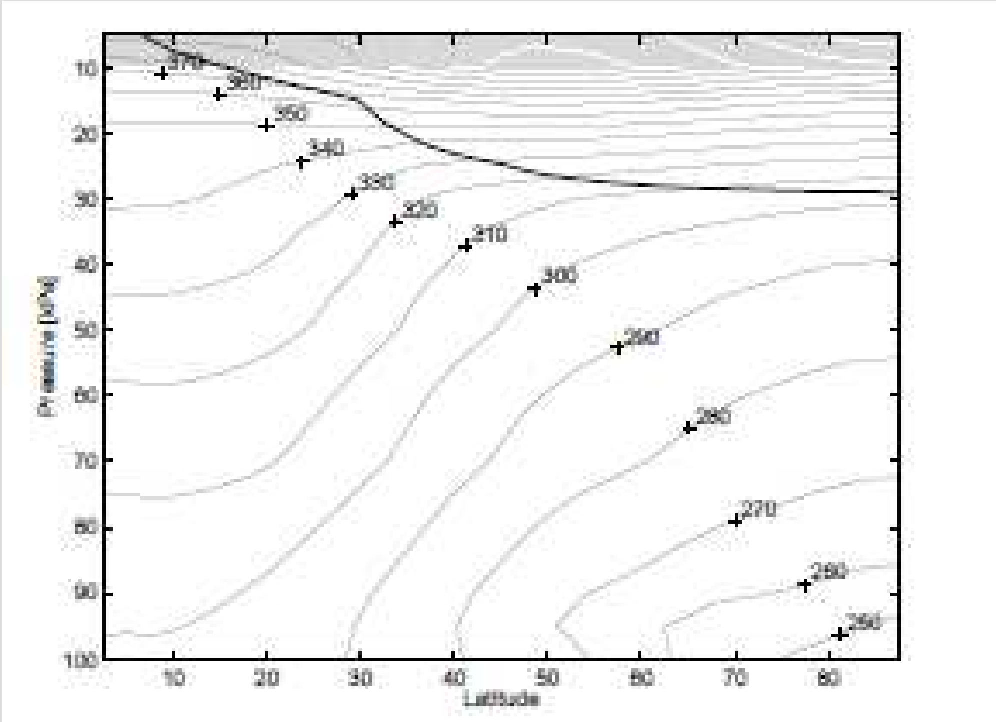
Annual and zonal mean distribution of potential temperature (solid) and temperature (dashed), in degrees K. The thick line denotes the thermal tropopause. The shaded regions denote the ‘‘lowermost stratosphere’’, which is that part of the stratosphere ventilated by the troposphere along isentropic surfaces, wherein stratosphere-troposphere exchange can be particularly rapid. (Holton et al., 1995).

A key question in general circulation theory is whether or not the slope of the mean isentropes in the troposphere is strongly constrained.

The observed slope is close to the aspect ratio of the troposphere: an isentropic surface that is near the ground in the tropics rises to the tropopause in polar latitudes.

Is this a coincidence, or is this particular slope feature?

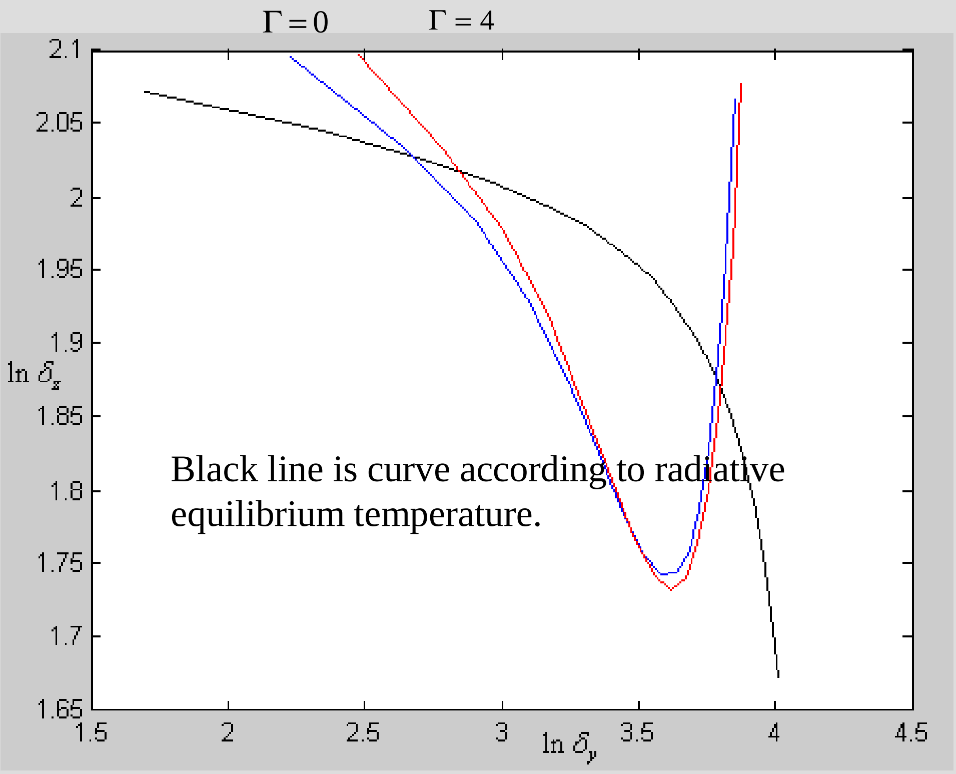
Longitudinally averaged potential temperature (K) in NH, January. Heavy line 3.5.PVU
(T.Schneider, I. Held, 1998)



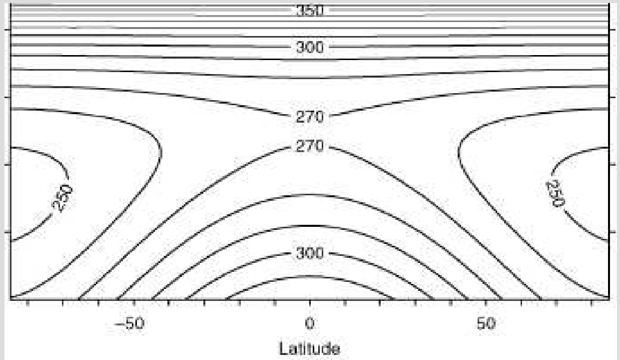
There are distinguishing between radiative and dynamical constraints on the thermal stratification:

•Dynamical constraints express balance conditions based on dynamical considerations, such as that moist convection maintains the thermal stratification close to a moist adiabat or that baroclinic eddy fluxes satisfy balance conditions derived from the mean entropy and zonal momentum balances

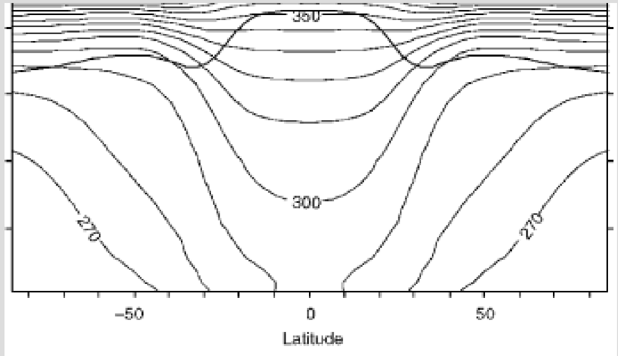
•Radiative constraints express the balance of incoming and outgoing radiant energy fluxes in atmospheric columns, plus any dynamical energy flux divergences in the columns.



(a)



(b)



(a) potential temperature (K) in radiative equilibrium of the reference simulation.
(b) potential temperature (K) in the reference simulation.

Theory for baroclinic turbulence in framework of two-layer quasi-geostrophic (QG) model

- The two-layer QG system provides us with what may be our simplest turbulent "climate" model. The state of this model is determined by the streamfunctions for the non-divergent component of the horizontal flow in two layers of fluid, meant to represent the flow in the upper φ_1 and lower φ_2 troposphere.

  QG potential vorticity:

$$q_k = \Delta\psi_k + (-1)L_R^{-2}(\psi_1 - \psi_2) + \beta y$$

and L_R is the radius of deformation Rossby - Obukhova, defined by,

$$L_R^2 = g\,\frac{(\theta_1 - \theta_2)}{\theta_0}\cdot\frac{H}{f^2}$$

with H the resting depth of the two layers

- A simple way of creating a statistically steady state is to force the system with mass exchange between the two layers, this model's version of radiative heating, arranged so as to relax the interface to a "radiative equilibrium" shape with a zonally symmetric meridional slope.

- Radiative equilibrium is a solution of these equations, with no flow in the lower layer and zonal flow in the upper layer, with the Coriolis force acting on the vertical shear $\Delta U = u_1 - u_2$ between the two layers balancing the pressure gradients created by the radiative equilibrium interface slope.

This flow is unstable, in the absence of the dissipative terms, when the isentropic slope is large enough and reverse the sign of the north-south potential vorticity gradient in one of the layers.

If the relative vorticity gradient of the zonal flow is negligible as compared to β , the criterion is classic one (N. Phillips) :

$$Ph_c = \frac{\Delta U}{L_R^2 \beta} > 1$$

(I. Held, 2005)

The existence of this critical slope presents us with a problem, since analogous models of inviscid baroclinic instability in continuously stratied atmospheres are unstable for any non-zero vertical shear (or isentropic slope).

The most fundamental limitation of QG dynamics is that it assumes a reference static stability; in this two-layer model the potential temperature difference between the two layers is fixed ????

But we can develop theories for the QG fluxes, and then use these outside of the QG framework

$$\Rightarrow \quad S_c = Ph_c = \frac{\Delta U}{L_R^2 \beta}$$

Scaling: $$D \sim \varepsilon^{3/5} \cdot \beta^{-4/5} \qquad \varepsilon = \frac{D}{\tau^2} \; (g^* H \rightarrow (NH)^2)$$

$$\tau = \frac{NH}{\Delta U \cdot f} \Rightarrow D \sim \frac{1}{\beta^2 \tau^3} \Rightarrow S_c{}^3 \sim \frac{D}{\beta \lambda^3} \; (I.Held, V.Larichev, 1996)$$

EDDY CLOSURE IN THE TWO-LAYER QG MODEL

Rate of transfer of energy through the spectrum – ε; vorticity gradient - β

$$\delta_y = -a\frac{\partial\overline{\theta}}{\partial y} \qquad \delta_z = H\frac{\partial\overline{\theta}}{\partial z}$$

$$S_c = Ph_c = \frac{\Delta U}{L_R^2\beta} \qquad S_c > 1$$ - Phillips's Criterion for 2-layer model

$$\lambda^2 = g\frac{(\theta_1 - \theta_2)}{\theta_0}H / f^2 = g^* H / f^2$$ - Radius of deformation

The potential energy extracted from the environment can be written in terms of the eddy potential vorticity flux in either layer:

$$\varepsilon = \Delta U \cdot P_1 = -\Delta U \cdot P_2 = \Delta U D_1 \beta(1 + S_c) = \Delta U D_2 \beta(1 - S_c)$$

$\beta(1+S_c)$ - Mean potential vorticity gradient of upper layer

$\beta(1-S_c)$ - Mean potential vorticity gradient of lower layer

Using the scaling for the diffusivity due to baroclinic eddies (P. Stone, 1972), we can try to develop a theory for the static stability.

Diffusivity in each layer is defined as the eddy potential vorticity flux divided by the mean potential vorticity gradient.

$$D \sim \delta_h^3 \cdot \delta_v^{-3/2} \qquad F_V \cdot \delta_v \sim F_H \cdot \delta_h \Rightarrow F_V \sim \delta_h^{5} \cdot \delta_v^{-5/2}$$

F_V – vertical eddy heat flux, F_H – horizontal eddy heat flux

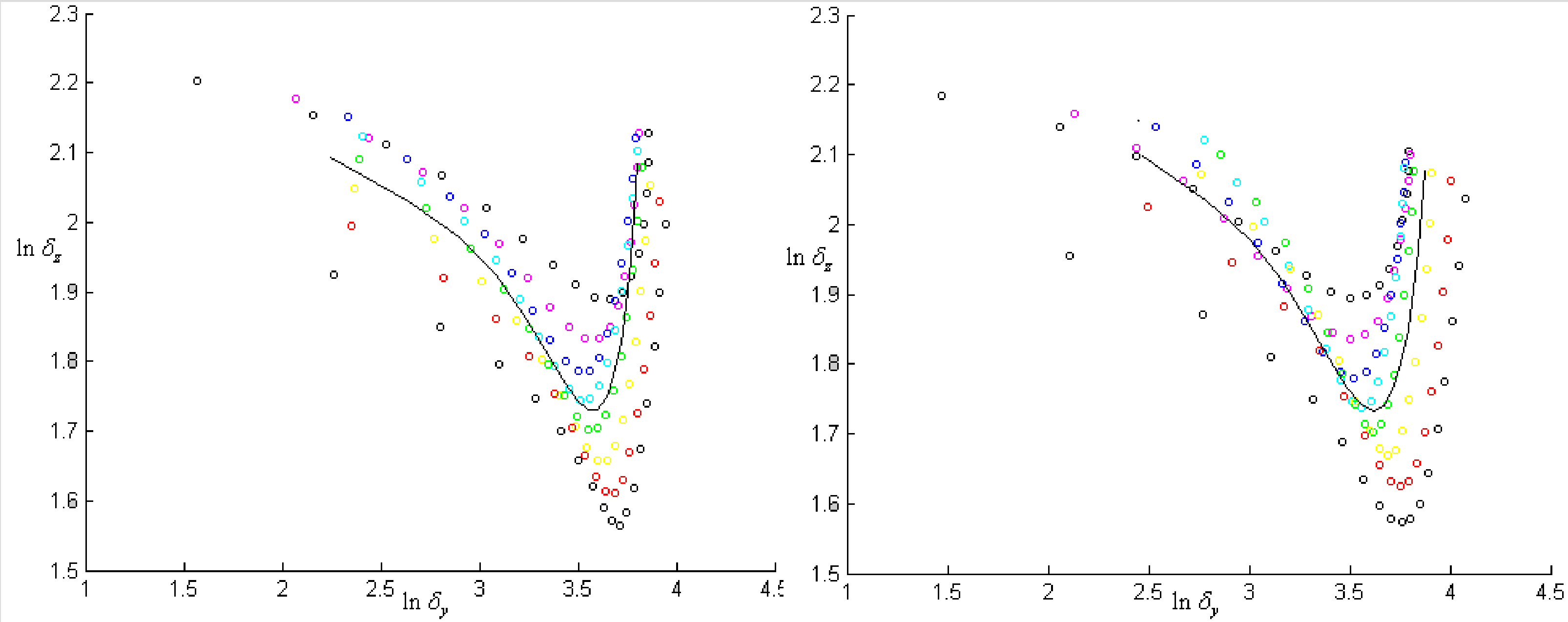
! $$F_V \sim \delta_v \Rightarrow \delta_v \sim \delta_h^{10/7},\ D \sim \delta_h^{6/7} \Longrightarrow I_\theta = \frac{\delta_h}{\delta_v} \sim \delta_h^{-3/7}$$

Modelling: $$\delta_v \sim \delta_h^{11/7} \rightarrow I_\theta = \frac{\delta_h}{\delta_v} \sim \delta_h^{-4/7}$$

The slope of isentropes in extratropical troposphere

$$\delta_y = -a\frac{\partial\bar{\theta}}{\partial y} \qquad \delta_z = H\frac{\partial\bar{\theta}}{\partial z}$$

Γ=0 Γ=4



$$\delta_v \sim \delta_h^{11/7} \quad \rightarrow \quad I_\theta = \frac{\delta_h}{\delta_v} \sim \delta_h^{-4/7}$$

Some remarks about sensitivity of baroclinic turbulence to heating…

The determination of the mean extratropic **thermal structure** is a long standing problem in the general circulation of the atmosphere. The equilibrium extratropical climate arises from the competition between **diabatic heating and dynamical transport.**

We have faced with hard problems: closure problem requires relating these eddy fluxes to the mean state. **It could be even ill posed problem**.

- One possible closer is baroclinic adjustment (Stone, 1978)
- Another approach to closure problems is turbulent diffusion (Held, Larichev, 1996)

Assuming that inverse cascade is stopped by beta effect, they obtain explicit Analytical prediction for eddy scales as function of mean flow:

$$\overline{(v'\theta')} \approx D\cdot\partial_y\overline{\theta} \sim \beta\cdot L_R^3\cdot\xi^3\cdot\partial_y\overline{\theta} \qquad \xi = -(\frac{f}{\beta H})\cdot\frac{\partial_y\overline{\theta}}{\partial_z\overline{\theta}}$$

- measure of isentropic slope

$$Q_H \approx \frac{\partial_z\overline{\theta}H\beta^2L_R^3}{f}\cdot\xi^4$$

Balance equation for $\overline{(\theta'^2)}$

$$\overline{(v'\theta')}\cdot\partial_y\overline{\theta} + \overline{(w'\theta')}\cdot\partial_z\overline{\theta} \approx \overline{(Q'\theta')}$$

Let's eddies are adiabatic Q'=0 → $\overline{(w'\theta')}\Big/\overline{(v'\theta')} \sim -\partial_y\overline{\theta}\Big/\partial_z\overline{\theta}$

$$\overline{(w'\theta')} \approx \frac{\theta_0}{g}\cdot\beta^3\cdot L_R^5\cdot\xi^5$$

Therefore, it's possible relate the heating and mean state as follows

$$Q_V \sim \overline{(w'\theta')} \sim \frac{\theta_0}{g} \cdot \beta^3 \cdot L_R^5 \cdot \xi^5 \qquad \frac{Q_V}{Q_H} \sim \frac{\beta H}{f} \xi$$

$$\xi \sim \frac{f}{\beta H} \cdot \frac{Q_V}{Q_H} \qquad L_R \sim (\frac{g}{\theta_0})^{1/5} \cdot \frac{\beta^{2/5} \cdot H}{f} \cdot Q_H \cdot Q_V^{-4/5}$$

$$\xi \cdot L_R \sim Q_V^{1/5}$$

- The eddy length scale

At last, it is possible to express gradients of temperature in terms of heating

$$\partial_y \bar{\theta} \sim \left[\frac{\theta_0}{g}\right]^{3/5} \beta^{4/5} \cdot Q_H \cdot Q_V^{-3/5}$$

$$\partial_z \bar{\theta} \sim \left[\frac{\theta_0}{g}\right]^{3/5} \beta^{4/5} \cdot Q_H^{2} \cdot Q_V^{-8/5}$$

Conclusion

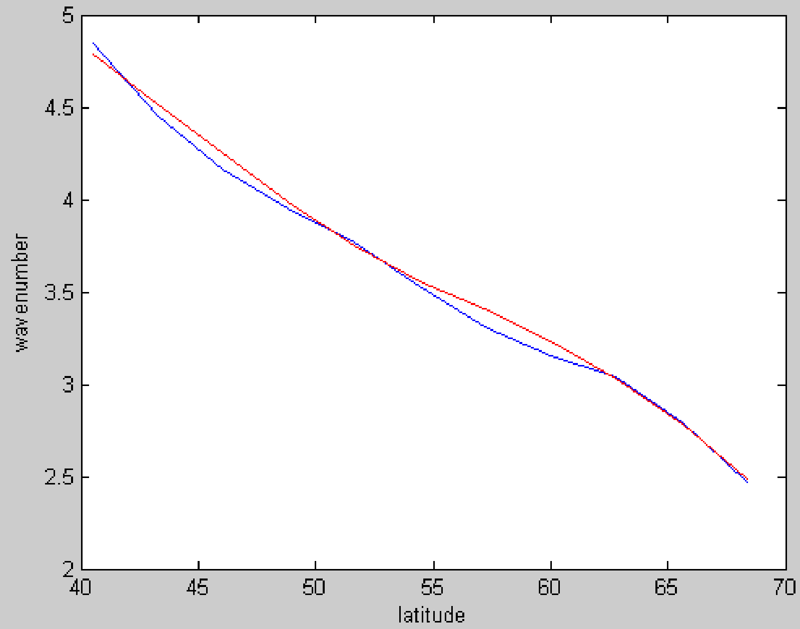
Knowledge of heating determines the local thermal structure.

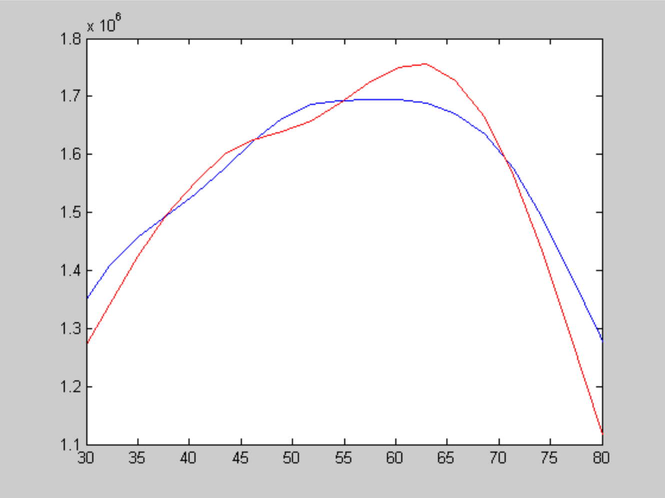
But the heating is not known *a priori. Heating is coupled to dynamics.*

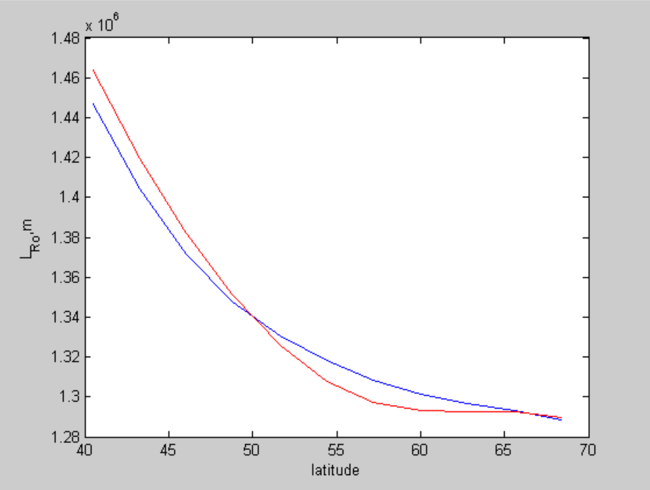
Before we have explored sensitivity of mean state on heating (in stratosphere and on surface)

Eddy Scales in the GCM

$$\bar{m}(\varphi) = \frac{\sum_k k \left| v^*(\varphi,k) \right|^2}{\sum_k \left| v^*(\varphi,k) \right|^2}, \quad \bar{\lambda}_e(\varphi) = 2\pi a \cos(\varphi) / \bar{m}(\varphi) \rightarrow \Delta\lambda$$







$$L_{Rn} = 2\pi \frac{\sqrt[4]{KE_{ed}}}{\sqrt{\beta}}$$

$$L_R^2 = g \frac{(\theta_1 - \theta_2)}{\theta_0} \cdot \frac{H}{f^2}$$

At scale smaller than Rossby radius the flow becomes barotropic and downscale cascad baroclinic turbulence halted at Rossby radius. The Rhines scale define transition between linear and nonlinear dynamics and, hence, inverse cascad borotropic flow halted at Rhines scale.

$$L_{Rh} = \left(\frac{u^{'}}{\beta} \right)^{\frac{1}{2}} \qquad L_{Ro} = \left(\frac{g}{\overline{\theta}} \frac{\partial \overline{\theta}}{\partial z} \right)^{\frac{1}{2}} \frac{H}{f}$$

- Rhince scale and Rossby radius

$$u^{'} = \left(\frac{L_{Rh}}{L_{Ro}} \right) \cdot \overline{u}$$

(I. Held, V. Larichev, 1996)

$$L_{Rh} = L_{Ro} \left(\frac{\delta_h}{\delta_v} \right) \cdot tg\varphi \qquad \varphi - latitude$$

Concluding remarks

When difference between Rhines scale and Rossby radius tend to zero then for baroclinic turbulence leaves no space for inverse energy cascade !

This fact limits application of the theory two-layer baroclinc turbulence!!!