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Influence of radiative equilibrium baroclinicity variation on stratification of troposphere

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Goals and objectives

- Now in connection with increase concentration of GHG and warming in troposphere which is accompanied with cooling stratospheres, and, accordingly, with change in dynamics of a polar vortex, the question is key, how changes of a stratospheric vortex affect troposphere cuirculation

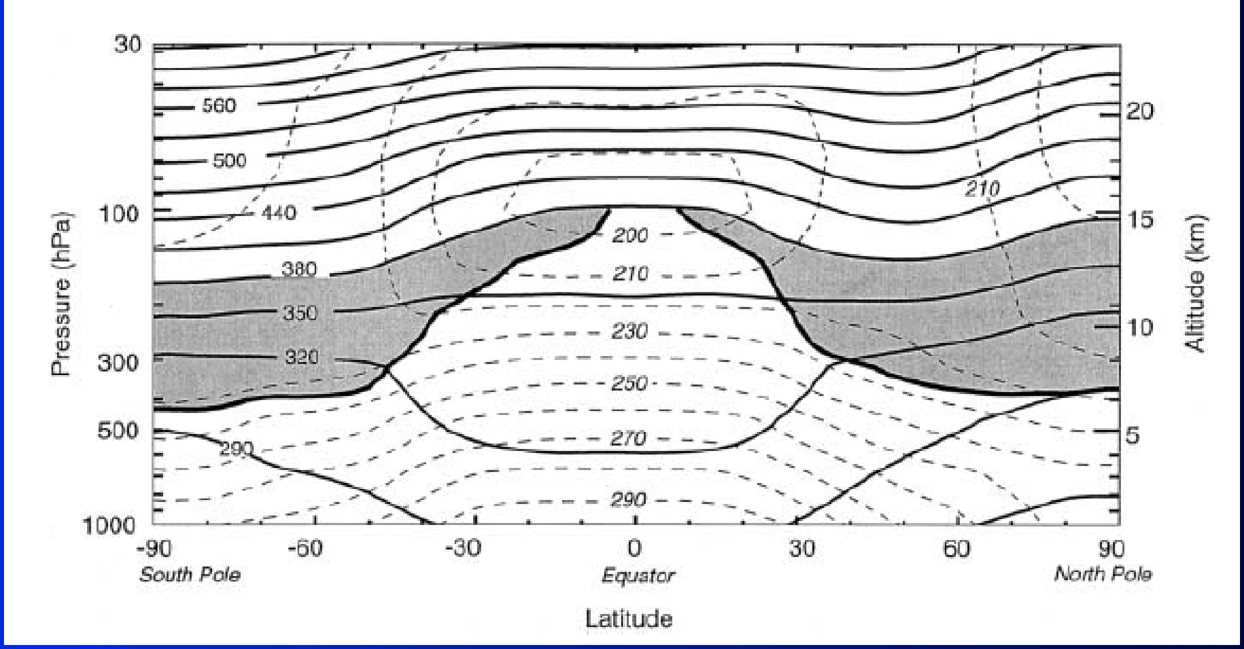
- The report was focused upon some studies about how radiative and dynamic processes interact to maintain the static stability and mean meridional potential temperature gradient in extratropic of Troposphere

- It is very important to establish relationships between these quantities

- In this report idealized dry primitive equation model on the sphere and particular the classic two-layer quasi-geostrophic (QG) model are used

Contents

- Introduction
- The stratosphere and troposphere
- Simulation scenario with dry primitive equation model
- Theory for baroclinic turbulence in frame of two-layer quasi-geostrophic (QG) model
- Sensitivity estimation of isentropic surfaces slope in the medium latitudes
- Conclusion



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).

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

In this paper, 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

$$T_R\left(\sigma,\varphi\right)=T_r\left(\sigma\right)+h\left(\sigma\right)$$ **Radiative equilibrium temperature**

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

In the stratosphere

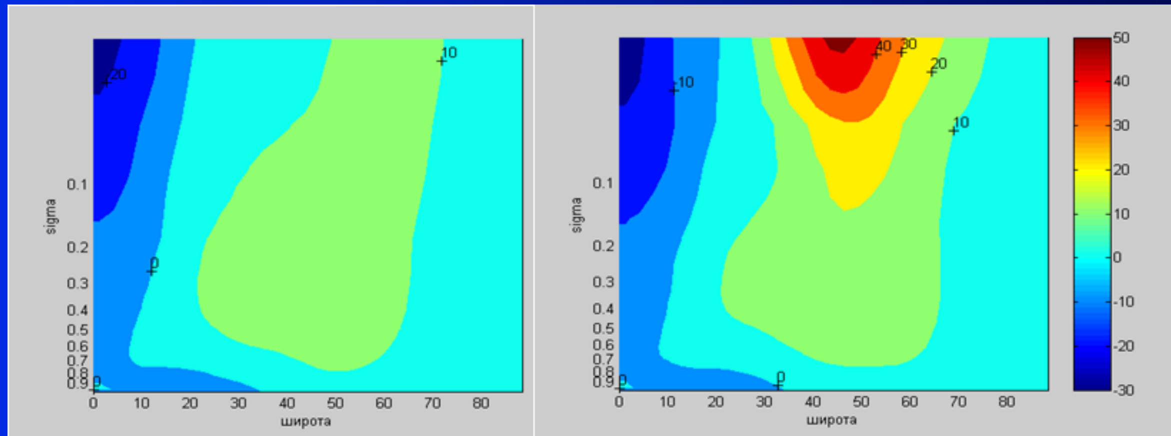
$$T_r(\sigma) = T_{tr} + (\Gamma_{max} - \Gamma) H \ln\left(\frac{\sigma}{\sigma_T}\right)$$

where T_{tr} =210K.

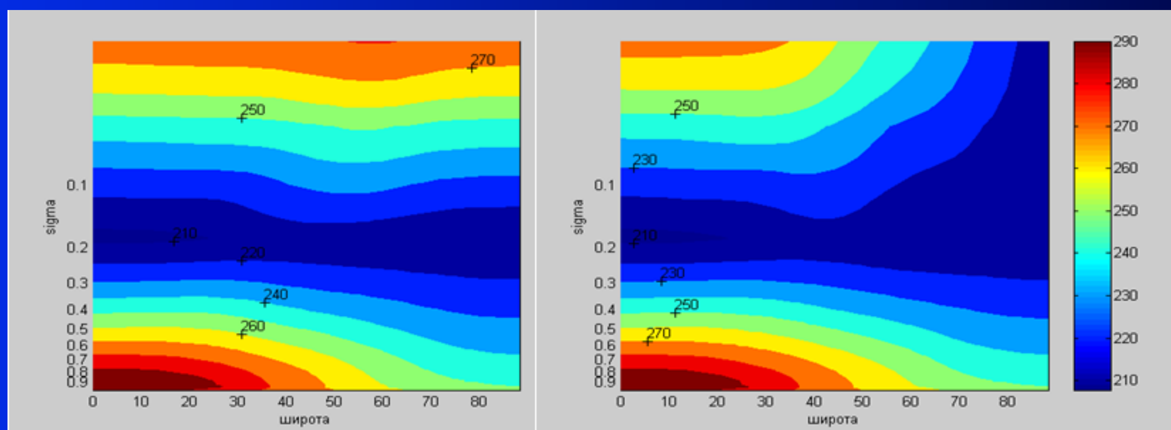
In this equation, the parameter Γ defines a temperature gradient.

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) b)



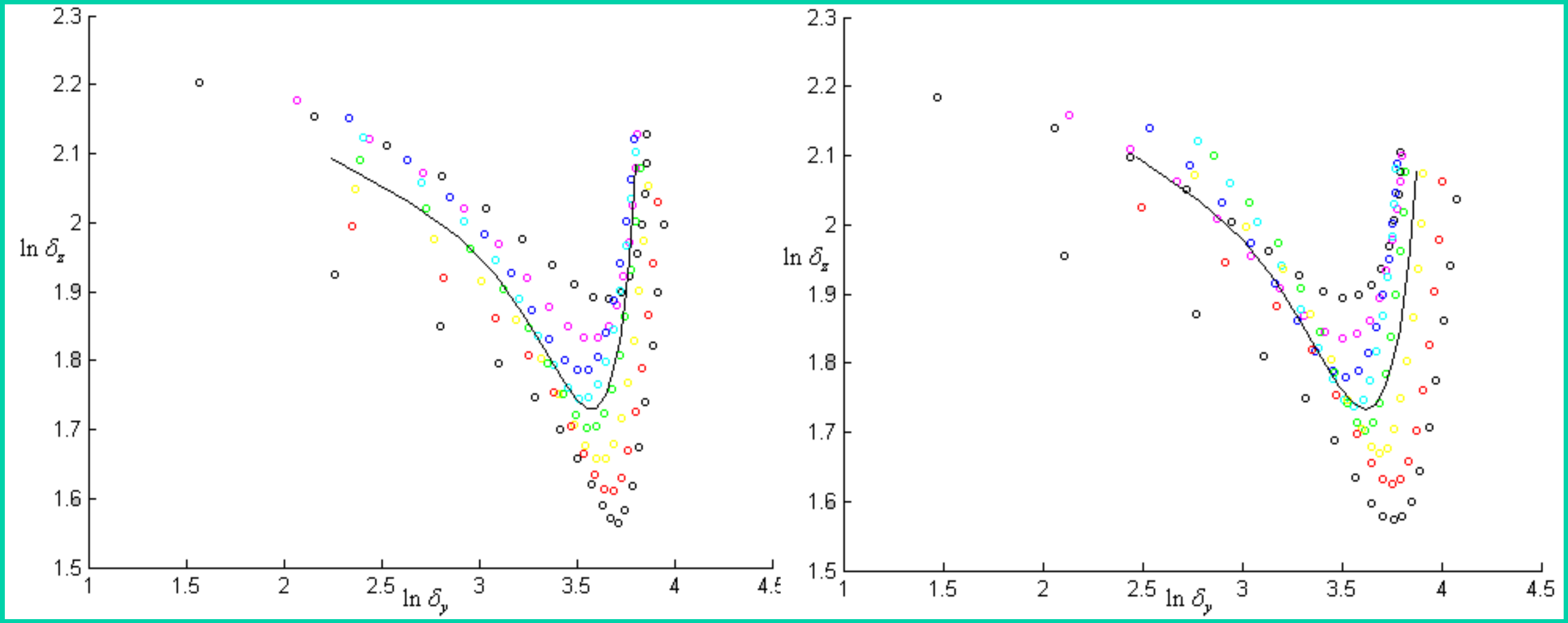
c) d)

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

The slope of isentropes in extratropical troposphere

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

$\Gamma=0$ $\Gamma=4$

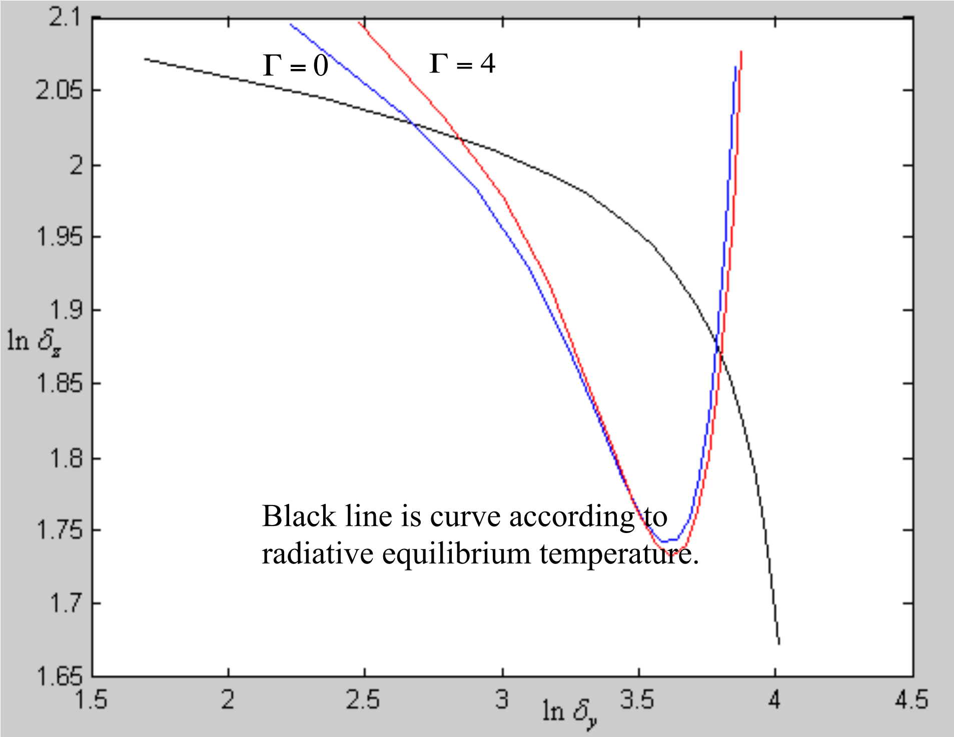


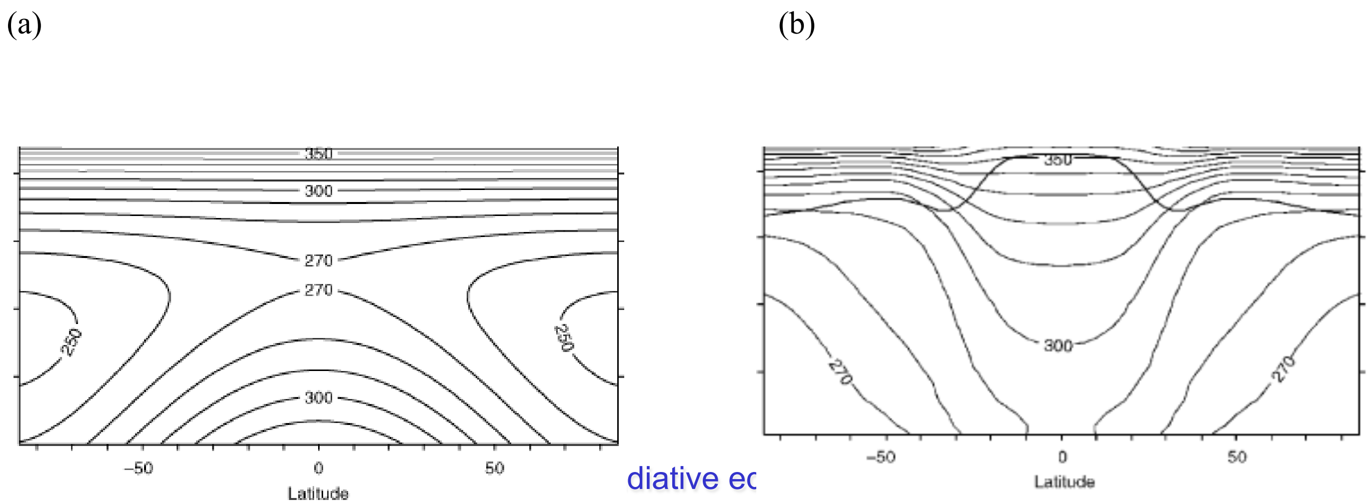
$$\delta_v : \delta_h^{11/7} \quad \rightarrow \quad I_\theta = \frac{\delta_h}{\delta_v} : \delta_h^{-4/7}$$

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.





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(b) potential temperature (K) in the reference simulation.

Theory for baroclinic turbulence in framework of two-layer quasi-geostrophic (QG) model (for example, I.Held, 2005)

- 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 ow in two layers of uid, meant to represent the flow in the upper φ_1 and lower φ_2 troposphere.

QG potential vorticity:

$$q_k = \nabla^2 \psi_k + (-1)^k \lambda^{-2} (\psi_1 - \psi_2) + \beta y; \;\; k = 1, 2$$

and λ is the radius of deformation, defined by,

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

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 ow in the lower layer and zonal ow 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 (N. Phillips) :

$$S_c = \frac{\Delta U}{(\beta\lambda^2)}$$

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 dierence between the two layers is fixed ????

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

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

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

EDDY CLOSURE IN THE TWO-LAYER QG MODEL

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

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

$$S_c = \frac{\Delta U}{(\beta \lambda^2)}$$

$S_c > 1$ - 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 diusivity due to baroclinic eddies (P. Stone, 1972), we can try to develop a theory for the static stability.

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

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

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

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

Modelling:

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

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?

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)

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

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

Concluding remarks

We investigated the relevance of the generalized baroclinic adjustment for this dry primitive equation model on the sphere with standard Newtonian forcing.

When the radiative equilibrium is changed, our results the model equilibrates at a preferred isentropic slope in troposphere, even for significant perturbations in the heating.

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