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The influence of the polar stratospheric vortex dynamics on circulation in troposphere*

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Abstract. Currently, there is observed an increase of the polar vortex as well as an increase in the temperature gradient. This paper considers the influence of the stratosphere circulation on the troposphere. It has been found that an enhancement of cooling in the polar stratosphere brings about changes in the tropospheric processes, and this means a positive annular mode response in the troposphere.

Introduction

Dynamic interaction between the stratosphere and the troposphere, which is associated, in particular, with planetary waves, can essentially affect on variations of the tropospheric circulation with time scales from several days up to several months. The climatology of extratropical stratosphere is defined by the intensity of a polar vortex, which is connected with gradients rate of temperature. Large gradients generate strong eddies which transfer a sufficient amount of heat and tends to decrease gradients. Small temperature gradients cause weak small eddies which carry little heat and do not prevent from baroclinicity by diabatic processes. All the studies of the general circulation of the atmosphere are based on the dividing the circulation to the motion averaged by latitude (or zonal mean) and deviations from the mean or eddies. In the stratosphere and the mesosphere main disturbances represents waves of planetary scale which are described by linear theory with enough accuracy. So, the splitting to the zonal and the eddy components is an effective method of theoretical analysis of the stratospheric dynamics. Recently, there is observed a decrease in temperature of the top layers of the atmosphere, a contrast between a tropical and a polar stratosphere and, hence, a temperature gradient also decreases. Therefore, the question on how changes of a polar vortex affect conditions of the bottom atmosphere layers is of an essential interest. However, we have no deep understanding of the mechanism of interaction of the atmosphere and the stratosphere. This paper is a numerical research into changes of the dynamical field in the atmosphere, particularly, in the low troposphere.

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1. A polar vortex dynamics and a circulation in the troposphere

M. Ambaum and B. Hoskins (2003), using the daily and the monthly average data, established a connection between variations of surface pressure, strength of a polar vortex and tropopause altitude, an increase of the polar vortex is connected with a low tropopause over Iceland and a high tropopause over the Arctic. A high tropopause altitude is connected with strength and twisting of a column of the air and a small drop of surface pressure in the polar region.

A model of the atmosphere. To describe the dynamics of the atmosphere, spectral general circulation model proposed in [1], was used with the following notations: $\mu = \sin\varphi$, φ is latitude; λ is longitude; $\sigma = p/p_s$ is vertical coordinate.

A vertical component of the potential vorticity $\zeta = \xi + f$, the horizontal divergence D, the temperature T and the surface pressure p_s are taken as prognostic variables.

When modeling the dynamics of the atmosphere, parametrization of the radiative heat fluxes developed in [2] was used. Radiative heating in a layer between 30 km and 70 km was determined by absorption of the solar ultraviolet radiation by ozone and infrared radiation by CO_2, and these processes can be represented by a simplified model, in which a radiative heat inflow depends only on temperature.

The vorticity equation:

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

The divergence equation:

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

The thermodynamical equation:

$$\frac{\partial T'}{\partial t} = -\frac{1}{1-\mu^2}\frac{\partial}{\partial\lambda}(uT') - \frac{\partial}{\partial\mu}(vT') + D\cdot T' - \dot\sigma\frac{\partial T}{\partial\sigma} + \kappa\frac{T\omega}{p} + \frac{T_R - T}{\tau_R} - k(-1)^n\nabla^{2n}T'. \tag{3}$$

The continuity equation:

$$\frac{\partial \ln p_s}{\partial t} = -\frac{U}{1-\mu^2}\frac{\partial \ln p_s}{\partial\lambda} - V\frac{\partial \ln p_s}{\partial\mu} - D - \frac{\partial\dot\sigma}{\partial\sigma}. \tag{4}$$

The quasistatic equation:

$$\frac{\partial \Phi}{\partial \ln \sigma} = -T. \tag{5}$$

The vertical velocity (in terms of $\omega = \frac{dp}{dt}$) is found from the equation

$$\frac{\omega}{p} = \vec{V} \cdot \nabla \ln p_s - \frac{1}{\sigma} \int_0^\sigma \left(D + \vec{V} \cdot \nabla \ln p_s \right) d\sigma. \tag{6}$$

Here u is zonal velocity, v is meridional velocity, $\vec{V} = (u, v)$ ia a velocity vector, $U = u\sqrt{1-\mu^2}$, $V = v\sqrt{1-\mu^2}$,

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

$T = T_0 + T'$ Φ is a geopotential, $f = 2\Omega \sin\varphi$ is the Coriolis parameter, k is a coefficient of superdiffusion, n is a superdiffusion degree, κ is an adiabatic coefficient, $\tau_R = 1/\alpha_R$ is a scale of the radiative cooling, τ_f is a scale of the surface friction, considered to be infinite everywhere except the bottom surface.

In the thermodynamical equation, the term $\frac{T_R - T}{\tau_R}$ is responsible for the radiative heating.

2. Numerical experiment and analysis of the results

Two numerical experiments were conducted in which the radiative equality temperature T_R varied.

In the first experiment, as T_R, the following function was taken ([3]):

$$T_R^1(\sigma, \varphi) = T_r(\sigma) + h(\sigma)(\Delta T_{SN}\frac{\mu}{2} - \Delta T_{EP}(\mu^2 - \frac{1}{3})),$$
$$h(\sigma) = \begin{cases} \sin\left(\frac{\pi}{2}\frac{\sigma - \sigma_T}{1-\sigma_T}\right), & \sigma > \sigma_T, \\ 0, & \sigma \le \sigma_T, \end{cases}$$

$\Delta T_{EP} = 60$ K, $\Delta T_{SN} = 0$, $\sigma_T \approx 0.2$ is a value of σ on a tropopause, $T_r(\sigma)$ is a standard temperature. In this case, the horizontal temperature gradient is equal to zero ($\Gamma = 0$).

In the second case of the temperature profile in the stratosphere

$$T_R^2(\sigma, \varphi) = (1 - \omega(\varphi))T_0 + \omega(\varphi)T_1(\sigma),$$

$\omega = \sqrt{1 - \sin\varphi}$, $T_1(\sigma) = T_0 + \Gamma H \ln(\sigma/\sigma_T)$, $T_0 \approx 210$ K, where Γ is a temperature gradient, $H = 7$ km is the height of the stratosphere offered in [3]. In the troposphere, the temperature of the radiative equality does not change.

Temperature and wind speed fields for these background temperature distributions are shown in Figures 1 and 2, respectively. In the first case, the zonal velocity field is symmetric relative to the equator. We can see from the figure that in the moderate latitudes, the western winds blow with a maximum of speed equal to 15 m/s in the troposphere and 25 m/s in the stratosphere. In the region of the equator, the wind is eastern and its speed is 15 m/s. In the case of $\Gamma = 2.25$, in the stratosphere of the south hemisphere, there are eastern winds.

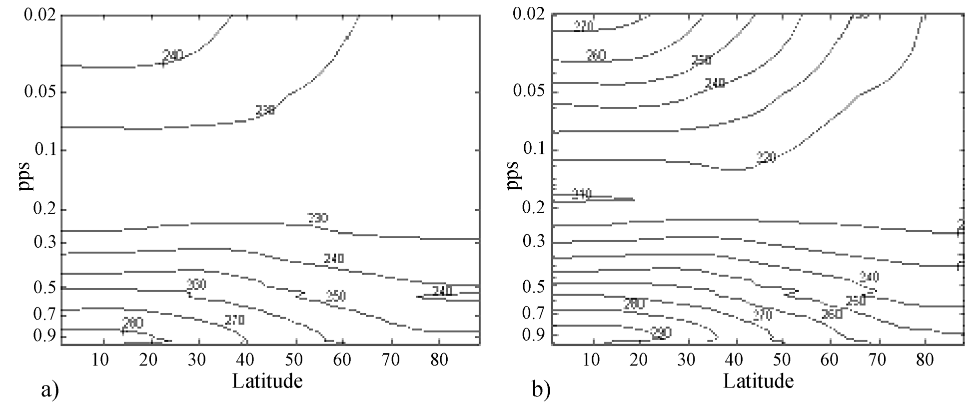

Figure 1. Latitudinal-altitudinal distributions of temperature (K): a) $\Gamma = 2.25$; b) $\Gamma = 4$

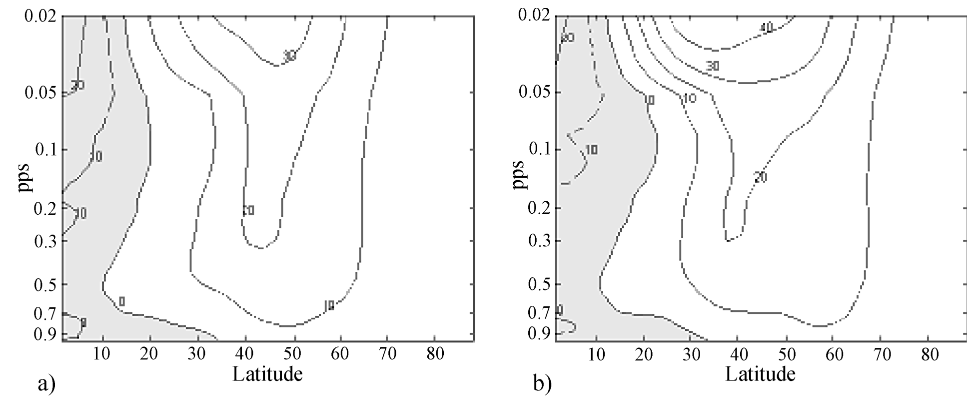

Figure 2. Latitudinal-altitudinal distributions of zonal wind (m/s): a) $\Gamma = 2.25$; b) $\Gamma = 4$. The regions with the eastern wind are shown in a darker color

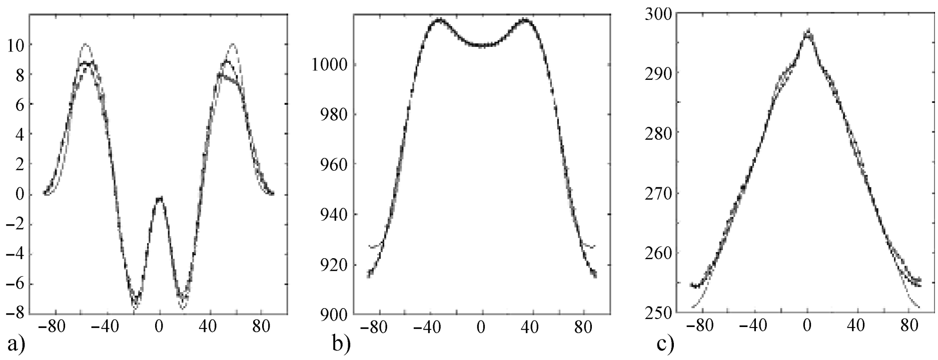

Figure 3. Dynamic fields in the ground layer: a) mean-zonal wind speed (m/s); b) mean-zonal atmosphere pressure; c) mean-zonal ground temperature

In Figure 3, the dynamic fields on the Earth's surface depending on latitude are presented in these figures, graphs obtained with the radiative balance temperature T_R^1 are shown by the solid line. In the high latitudes, when the temperature in the stratosphere increases, a decrease in pressure occurs in low layers of the polar troposphere. According to quasi-geostrophic formulas, there is an increase in the western winds in the polar region. At the same time, a warming, which is significant within a polar cap and insignificant out of it, is observed in the troposphere. In the quasi-geostrophic theory [4], the concept of the potential vorticity, whose value fits diagnostics of the processes taking place in the atmosphere because when the motion is adiabatic it determines the dynamics of the atmosphere and acts like a passive tracer. We may do some suggestions about a character of the air mass motion and, also, about its properties such as its humidity and concentrations of the tracers. A meridional vorticity flux $Q = \overline{v'q'}$ is an indicator to the eddy activity transport.

In usual hydrostatic conditions, the PV amount per volume unit is equal to the absolute "isentropic" vorticity which can be written down in the form $\zeta_{a\theta} = \zeta_\theta + f$, where $\zeta_\theta = \dfrac{\partial v}{\partial x} - \dfrac{\partial u}{\partial y}$, f is a Coriolis parameter, the derivatives being taken with a constant potential temperature. In the medium stratosphere, the isentropic surfaces are almost equal to isobaric, so the maps of the potential vorticity are almost equal to those of the absolute vorticity, and absolute vorticity fluxes may give a notion about a character of interaction between different atmosphere layers as PV fluxes. Meridional fluxes are positive in the medium latitudes (30–65°) and negative in the tropics. This means that in the medium latitudes, there is a transport of the air masses with high vorticity to the poles. One of features of the circulation with a large temperature gradient is a zone of intensive concentration of vorticity in the high stratosphere of the south hemisphere. In the case of a large vorticity gradient, the zone, in which the transport of vorticity to the poles is essential. In [5, 6], the formula was obtained for an eddy flux of the quasi-geostrophic vortex:

$$\overline{v'q'} = -\overline{u'v'}_y + f_0 \left(\frac{\overline{v'\theta'}}{\overline{\Theta}_p} \right)_p$$

This relation can be written down as: $\nabla \cdot E = \overline{v'q'}$, where E is a vector of the Eliassen–Palm flux determined on the surface (y, p) and written in a form $E = \left(-\overline{u'v'}, f_0 \frac{\overline{v'\theta'}}{\overline{\Theta}_p} \right)$.

Convergence of the total flux of wave energy determines the energy exchange between waves and a zonal flow. The direction of the Eliassen–Palm flux is an indicator to relative maintenance of the heat flux and the momentum flux. The vertical component of E near the boundary shows the heat flux through this boundary.

This can be shown when the notion of group velocity c is applicable: $E = cA$, where $A = \frac{1}{2} \frac{\overline{q'^2}}{\bar{q}_y}$ is wave activity. So, E is a measure of a resulting wave spreading in the broad sense. It is obvious from the Eliassen–Palm cross-section that the momentum transport in the high latitudes is greater at non-uniform temperature distribution than in the case when temperature does not depend on latitude.

In Figure 4, there are Eliassen–Palm cross sections, including Eliassen–Palm flux and its divergence. The region, in which a horizontal component of the EP-flux is large, at large gradients is localized near to 50° latitude. Also, weakening of the momentum transport in a subtropical bottom layer and changing its direction in the polar cap region should be noted. So,

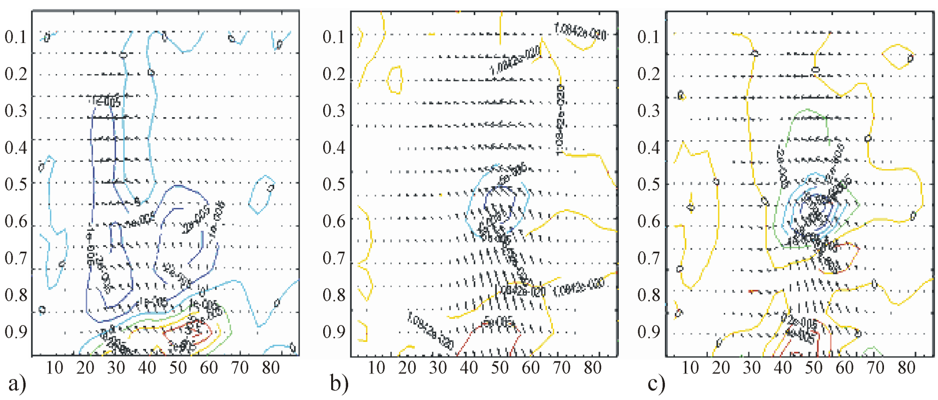

Figure 4. Eliassen–Palm cross section: a) $\Gamma = 0$; b) $\Gamma = 4$; c) difference between a) and b)

a conclusion can be made that at large temperature gradients, the wave activity transport by horizontal fluxes is weaker than at small ones.

In the standard atmosphere, there is a sufficiently large region at the medium latitudes of the troposphere, where the Eliassen–Palm flux is directed upwards. In the case when $\Gamma = 2.25$, such a region is much narrower and is located near to 50° in the north hemisphere and near to 60° in the south hemisphere.

When a polar vortex is strong, the wave fluxes tend to be directed equator-ward within the troposphere and to converge in the subtropical troposphere, whereas when the vortex is weak, the wave fluxes tend to be directed upward from the troposphere to the stratosphere and to converge, implying an anomalous westward wave force, in the mid- and in the high-latitude stratosphere.

Conclusion

The sensitivity of the tropospheric circulation to thermal perturbations of the polar stratosphere is examined using a dry general circulation model (GCM) with zonally symmetric forcing and boundary conditions.

We reveal from a simple GCM that some part of the tropospheric sensitivity to stratospheric cooling is attributed to the interaction between the zonal-mean flow and transient eddies on both the synoptic and the planetary scales.

We attribute this to the relatively weak eddy forcing of the stratosphere by the troposphere, this characteristic might be sensitive to the presence of topography and other changes to the model formulation.

For sufficiently strong cooling of the polar stratosphere, the tropospheric jet shifts polewards at the surface. This is accompanied by a drop in the surface pressure at high latitudes, and this means a positive annular mode response in the troposphere.

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