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Highlights

Changes in turbulent processes caused by atmospheric gravity waves from troposphere

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- Changes in temperature and wind speed recorded in the Earth‘s upper atmosphere above hurricanes can be explained by the propagation of atmospheric gravity waves.
- Intensification of turbulent processes above tropospheric energy sources was recorded in the range of altitudes from 75 to 100 km.

Changes in turbulent processes caused by atmospheric gravity waves from troposphere

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Abstract

We have determined that changes in temperature and wind speed recorded in the Earth's upper atmosphere above tropospheric sources (hurricanes) can be explained by the propagation of atmospheric gravity waves (AGW). We carried out modeling of the propagation of AGW with a period of 65 minutes and $k_x = 10^{-5}$ m^{-1} using multi-layer methods in a non-homogeneous, non-isothermal atmosphere, taking into account viscosity and thermal conductivity. We obtained that disturbances in the horizontal component of the velocity are five times greater than the increase in the vertical component of the velocity, and temperature changes can reach 30 K. We should note that the disturbances of temperature and pressure as a result of AGW spreading are superimposed onto the usual view of changes of pressure and temperature with the altitude and reach the maximum amplitude in the range from 90 to 100 km. The obtained changes in the temperature of the upper atmosphere and the velocity with height as a result of the presence of AGW made it possible to estimate the values of the coefficients of turbulent viscosity and thermal conductivity in the upper atmosphere of the Earth above tropospheric energy sources. Intensification of turbulent processes was recorded in the range of altitudes from 75 to 100 km.

Keywords: troposphere-atmosphere interactions, neutral atmosphere, atmospheric-gravity waves, tropospheric energy source, turbulent processes
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1. Introduction

Turbulent layers, usually 3-5 km in thickness, at around 100 km above Earth's surface were detected as a result of an analysis of processes in the upper atmosphere. These turbulent layers could occur after the attenuation of high-altitude gravity waves (Kozak, 2002; Kozak et al., 2004). This was the first attempt to correlate changes in temperatures and dynamics of the upper atmosphere with the changes in the turbulent parameters in this region, where energy transmission is done by propagating waves, that flow into the neutral Earth atmosphere. The main goal of the investigation is to see if this effect is only the result of surface energy sources or if it is as well caused by other energy sources from the troposphere.

A lot of research was dedicated to turbulent processes in Earth's lower atmosphere. In book (Monin) has a detailed and finalized description of turbulent processes analysis based on the semi-empirical hypothesis of perturbation transmission in the inertial interval. The lifespan of turbulent layers in the troposphere is long, and they can exist long after the wave source of turbulence is caused (Hocking, 1999; Hines, 1988; Barenblatt, 2004).

Stationary equations for special functions of isotropic turbulent velocity fields and temperature-certified environments may be used for analysis because of the nature of the turbulence to adapt rapidly to the changing parameters of the atmosphere, caused by the propagation of low-frequency high-altitude gravity waves. These equations proved themselves well in the analysis of the lower Earth atmosphere (Monin). To "close" the equations for the spectral functions of the isotropic turbulent velocity and temperature fields, semi-empirical conditions of the Lamli-Shura theory are used (Monin; Gavrilov and Yudin, 1992). As a result, the expressions for the spectral density of the kinetic energy of pulsations ($E(k)$), and the spectral densities of changes in velocity ($E_{uz}(k)$) and temperature ($E_{Tz}(k)$) in the vertical direction and will be given by the following equations (Monin):

$$\begin{aligned}
&E(k) = C_E[W(k)]^{2/3}k^{-5/3},\\
&E_{uz} = -2C_u[W(k)]^{1/3}k^{-7/3}\partial\bar{u}/\partial z,\\
&E_{Tz} = -2C_T[W(k)]^{1/3}k^{-7/3}\left(\partial\bar{T}/\partial z + \gamma_a\right),
\end{aligned}$$

where C_E, C_u and C_T are positive constants, and $W(k)$ is the velocity of kinetic energy spectral transmission. For evolved turbulence ($k < k_\eta = (8l)^{-1}$, where $l = v^{3/4}{\varepsilon_d}^{-1/4}$ is Kolmogorov scale) specific speed of viscous dissipation of turbulent energy ε_d is conserved. According to the results by (Gavrilov and Shved, 1975; Fritts, 1984; Vinnichenko, 2013), we can state that turbulent vortexes are not larger than the thickness of the layer $L_0(10-20$ km).

The phenomenological coefficient of turbulent viscosity for a population of vortexes with wave numbers ranging from $k_0\,(k_0 = 2\pi/L_0)$ to k_η can be derived from the following correlations:

$$\begin{aligned}
\varepsilon_d &= W(k) \approx \Omega^3{k_0}^{-2},\\
E(k) &\approx C_E{\varepsilon_d}^{2/3}k^{-5/3}\\
K_V(k) &\approx \alpha{\varepsilon_d}^{1/3}k^{-4/3},
\end{aligned} \tag{1}$$

where

$$\alpha = 3C_u/2, \quad \Omega^2 = C_u\left(\frac{\partial\bar{u}}{\partial z}\right)^2 - C_T\frac{g}{\bar{T}}\left(\frac{\partial\bar{T}}{\partial z} + \gamma_a\right).$$

These equations represent the Kolmogorov-Obuhov and Richardson's laws for $E(k)$ and $K_V(k)$ in the inertial interval. The coefficients of turbulent heat transmission and turbulent viscosity are correlated by the relation $K_T = K_V\mathrm{Pr}^{-1}$. According to experimental data by (Monin), constants in these equations are equal to $C_E \approx 1.4$ and $\alpha \approx 0.1-0.2$, which gives $C_u \approx 0.07-0.13$. Thus, the constant C_T is determined by the Prandtl number, which depends on the certification of the flow (for unstable certification $Pr \sim 0.8-1$, as for strong stability the increase to $Pr \sim 10$ is observed (Izakov, 2007). The edge of the inertial interval (which depends on altitude) can be determined by

$$k_\eta = \frac{1}{8}\left[\frac{g\left(\partial\bar{T}/\partial z + \gamma_a\right)}{v\bar{T}}\right]^{1/2}. \tag{2}$$

The following formula is only applicable for $\Omega^2 > 0$. This corresponds to the condition $Ri < C_u/C_T$ or $Rf < 1$, where

$$Ri = g\bar{T}^{-1}\left(\partial\bar{T}/\partial z + \gamma_a\right)/(\partial\bar{u}/\partial z)^2,$$
$$Rf = Ri/Pr.$$

Here, Rf and Ri are the gradient and dynamic Richardson's numbers, respectively.

The impact of vertical wind shifts and temperature on the turbulent mode in aforementioned correlations is described by Ω, meanwhile $\bar{u}$ and $\bar{T}$ can be presented as $\bar{u} = u_0 + u', \bar{T} = T_0 + T'$, where u_0, T_0 are background values and u', T' are perturbations caused by the passing wave.

Thereby the analysis of the effect of high-altitude gravity waves on turbulence results in finding the changes in wind and temperature due to interaction with the waves.

2. Result of numerical simulation of temperature and wind changes during AGW propagation

During analysis of changes in temperature and wind velocity using data from satellites UARS and TIMED above 9 hurricanes increase in temperature to $25 - 40$ K at the mesopause level was obtained (Kozak et al., 2015; Pylypenko and Kozak, 2010). The storms category 4 and 5 on the Saffir-Simpson hurricane scale was considered. Taking into account the localization of the perturbation, the transmission of energy to the upper parts of the atmosphere can be from atmospheric gravity waves (AGWs). While AGW ascends in adiabatic mode, its density decreases and wave amplitude increases (Gavrilov and Kshevetskii, 2013; Imamura and Ogawa, 1995; Dzubenko et al., 2003; Kozak et al., 2004), at the same time it is deviating from adiabatic conditions. Such an effect often results in a decrease in waves' stability, as a result, they fragment and create a system of large vortexes. At the same time, turbulent layers emerged that were mainly observed in regions with greatly deformed vertical profiles of temperature and wind speed (Kozak, 2002; Kozak et al., 2004).

Despite a great number of studies, to this day there is no clarity on which mechanisms of wave damping dominate at different altitudes. It is worth mentioning that the problem of instability, saturation and dissipation of AGW is actively discussed in scientific publications based on linear, quasi-linear and non-linear theories (Fritts and Alexander, 2003; Khomich et al., 2008; Rapoport et al., 2004; Cheremnykh et al., 2021; Fedorenko et al., 2021; Cao and Liu, 2023). This is due to the constant change of atmospheric parameters, its inhomogeneous nature and other causes which greatly complicate the analysis of AGW in a real dissipating environment.

For numerical modeling of AGW propagation in the Earth's atmosphere, we used the method which is based mainly on the method of solution of the Navier-Stokes equations, described in (Francis, 1973, 1975). It is similar to the multi-layer methods, which were firstly considered by Midgley and Liemohn (Midgley and Liemohn, 1966). AGW, while propagating in an inhomogeneous atmosphere, can dissipate its energy both by self-damping and by redistribution via various dissipative processes (viscosity, thermal conductivity etc.). Calculations of Midgley and Liemohn (1966) are based on the assumption that energy redistribution between gravity waves and dissipative processes in the lower atmosphere is negligibly small so that waves can be considered as of gravity type only. The iterative scheme used in this method is valid until dissipative processes dump much faster than atmospheric-gravity waves. Volland (1969) showed that viscosity and thermal conductivity may be important at upper atmospheric levels. He admits that the gravity-wave dominated solution used for lower altitudes, will be gravity-wave dominated at high altitudes also. In this work, we solve the Navier-Stokes equation taking into account dissipative processes. We consider a plane-parallel atmosphere consisting of homogeneous layers with constant temperature T_0, mass M, adiabatic constant γ, gravity g, viscosity to density ratio μ/ρ_0 and thermal conductivity to density ratio λ/ρ_0. We linearize the system of equations relative to the unperturbed steady state of the atmosphere:

$$\begin{cases} \rho_0 \frac{\partial u_i'}{\partial t} = -\frac{\partial p'}{\partial x_i} + \rho' g_i + \frac{\partial}{\partial x_i}\left[\mu\left(\frac{\partial u_i'}{\partial x_j} + \frac{\partial u_j'}{\partial x_i} - \frac{2}{3}\delta_{ij}\nabla\cdot\mathbf{u}'\right)\right], \\ \frac{\partial \rho'}{\partial t} + \nabla\cdot(\rho_0\mathbf{u}') = 0, \\ \frac{\rho_0 R_a}{(\gamma-1)M}\frac{\partial T'}{\partial t} = \nabla\cdot(\lambda\nabla T') - \rho_0\nabla\cdot\mathbf{u}', \end{cases} \quad (3)$$

where u', p', ρ' - denote a 1st-order perturbation of velocity, pressure, and density, caused by the propagation of the wave, R_a is universal gas constant,

$$p' = \frac{\rho' R_a T_0}{M} + \frac{\rho_0 R_a T'}{M}.$$

We search for the solution in a plane mode:

$$\frac{p'}{A_p} = \frac{T'}{A_T} = \frac{u_z'}{A_z} = \frac{u_x'}{A_x} \propto \exp\left(i\omega t - ik_x x - ik_z z + \frac{z}{2H}\right). \quad (4)$$

A_p, A_T, A_z, and A_x are scaling factors, $H = R_a T_0/g$ is an atmospheric scale height. The horizontal wave number k_x and the real frequency ω are assumed to be constant throughout the atmosphere since the atmosphere depends neither on spatial coordinate x nor on time t. On the other hand, vertical wave number k_z varies through the different atmospheric layers.

Substituting a plain mode solution (4) into the (3) we turn the system of differential equations into a system of algebraic equations:

$$\begin{pmatrix} 1 & k - i\alpha & -1 & 1 \\ 1 & k & 0 & -\frac{1}{\gamma-1} + iR \\ 4\eta - \beta - 3i\eta\alpha k + 3\eta k^2 & \eta k - 3i\eta\alpha & 1 & 0 \\ 2i\eta\alpha + \eta k & -\beta - 4i\eta\alpha k + 4\eta k^2 + 3\eta & k & -i\alpha \end{pmatrix} \cdot \begin{pmatrix} \frac{k_x u'_x}{\omega} \\ \frac{k_z u'_z}{\omega} \\ \frac{p'}{p_0} \\ \frac{T}{T_0} \end{pmatrix} = 0 \tag{5}$$

where the dimensionless parameters are: $k = (k_z + \ i/2H)/k_x$, $\quad R = k^2 - i\alpha k + 1, \alpha = 1/k_x H$, $\quad \beta = \omega^2/gk_x^2 H, \eta = i\omega\mu/3p_0, \nu = i\lambda T_0 k_x^2/\omega p_0$. The coefficients of viscosity μ and thermal conductivity λ are specified via the density of main atmospheric parameters and proportional to the square root of the temperature.

A system of algebraic equations has a non-trivial solution if the determinant of the matrix of coefficients is equal to zero. Then we get the dispersion relation:

$$C_3 R^3 + C_2 R^2 + C_1 R + C_0 = 0 \tag{6}$$

where

$$C_3 = -3\eta\nu(1 + 4\eta),$$

$$C_2 = \frac{3\eta\nu(1 + 4\eta)}{\gamma - 1} + \nu\beta(1 + 7\eta) + 3\eta,$$

$$C_1 = -\left[\beta^2 - 2\eta\alpha^2(1 + 3\eta)\right]\nu - \frac{\beta(1 + 7\eta)}{\gamma - 1} - \beta,$$

$$C_0 = \frac{\beta^2 - 2\eta\alpha^2(1 + 3\eta)}{\gamma - 1} + \alpha^2(1 + 3\eta).$$

For a given frequency ω and horizontal wave number k_x, this dispersion relation gives solutions for R and, accordingly, for k_z.

We join the solutions for waves in different adjacent layers by assuming continuity for vertical velocity and moment flux. So, for a certain ω and a horizontal wave number k_x, the scale parameters A_z, A_x, A_p, and A_T are defined by the following formula:

$$A_z = \frac{\omega}{k_x}\left[(1+\eta)k - 2i\eta\alpha + \frac{k - i\alpha}{(\gamma-1)^{-1} - \nu R}\right] - \\ -\frac{\omega}{k_x}\left[1 + \eta - \beta + 3\eta R + \frac{1}{(\gamma-1)^{-1} - \nu R}\right], \qquad (7)$$

$$A_x = \frac{\omega}{k_x}\left[(1+\eta)k - i\alpha(1+3\eta) + \frac{k}{(\gamma-1)^{-1} - \nu R}\right] - \\ \frac{\omega}{k_x}\left[(1+4\eta)R - \eta - \beta - 1 + \frac{R-1}{(\gamma-1)^{-1} - \nu R}\right], \qquad (8)$$

$$A_T = \frac{T_0 k_x}{\omega}\left[\frac{A_x + kA_z}{(\gamma-1)^{-1} - \nu R}\right], \qquad (9)$$

$$A_p = \frac{p_0 k_x}{\omega}\left[A_x + A_z(k - i\alpha)\right] + p_0\frac{A_T}{T_0} \qquad (10)$$

If the values u'_x, u'_z, p' and T' are valid for Eq.4, then they are valid for the Navier-Stokes equations. In the course of numerical modeling while using Eqs.7-10 we have calculated the amplitudes of velocity disturbances (both vertical $u_z{}'$ and horizontal components $u_x{}'$), pressure p', and altitude measurements of temperature T' caused by the motion of AGW with a period of 65 minutes and horizontal component of the wave number $k_x = 10^{-5}$ m^{-1} (Grigor'ev, 1999). The damping rate $\delta \approx k_x^2/\omega^2 \approx 3.8 \times 10^{-5} s^{-1}$, which corresponds to 7.2 hours. At the period and k_x changing the scales of processes are also changing, however, the regularity of changes of atmosphere parameters in the case of AGW is still the same (Kozak et al., 2015).

For the analysis of the investigated parameters the initial conditions (temperature profiles, concentrations of all components, altitude of uniform atmosphere) were calculated using the model of MSIS (Hedin, 1991) for the days of maximal intensities of considered storms 2005 (Haitang, 18 July

2005; Katrin, 28 August 2005; Wilma, 19 October 2005. Hurricanes Katrina and Wilma developed above the Atlantic Ocean, Haitang above the Pacific Ocean). In the course of the analysis, we considered an atmosphere that is non-isothermal and is stratification in terms of density and concentration of main components while taking into account viscosity and heat-conductivity.

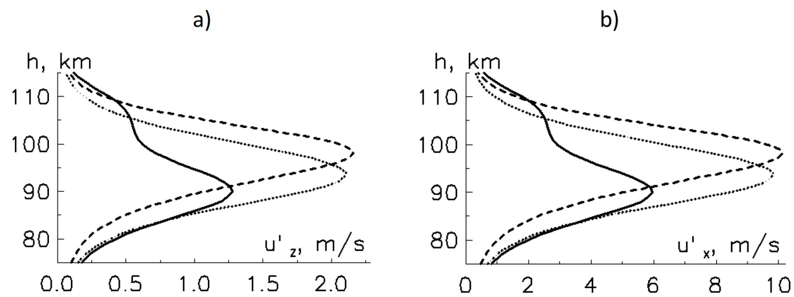

Figure 1: Change in vertical $u_z{}'$ (a) and horizontal $u_x{}'$ (b) component of velocity with altitude, as a result of passing of an AGW with a period of 65 minutes and $k_x = 10^{-5}$ m^{-1} (solid line -18 July 2005, dashed line - 28 August 2005, dotted line - 19 October 2005).

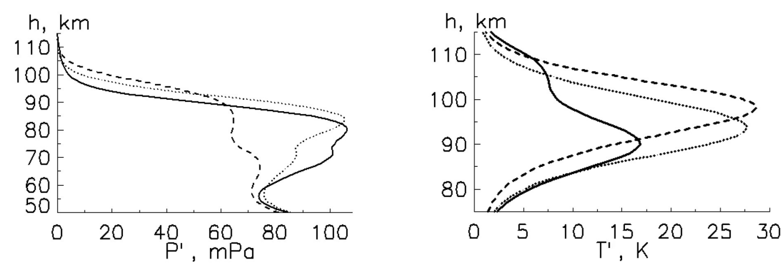

Figure 2: Change in pressure p' (a) and temperature T' (b) with the altitude as a result of passing of an AGW with a period of 65 minutes and $k_x = 10^{-5}$ m^{-1} (solid line - 18 July 2005, dashed line - 28 August 2005, dotted line - 19 October 2005).

The results of numerical modelling of the change with the altitude of vertical and horizontal components of AGW velocity, pressure, and temperature, as a result of waves passing, are shown in Figures 1-2. We should note that the disturbances of temperature and pressure as a result of AGW spreading are put onto the usual view of changes of pressure and temperature with the altitude. It can be seen that for the chosen set of modeling parameters, the waves propagate to heights of approximately 115 km and reach a maximum amplitude in the range of 90 to 100 km. In addition, disturbance in the horizontal component of velocity exceeds changes in the vertical one approximately in five times. Corresponding values are $6 - 10$ m/s and $1.3 - 2.1$ m/s. Temperature changes resulting from the transmission of AGW can reach 30 K.

The results of numerical modeling align with previous studies on the amplitude of temperature variations at heights of $80 - 90$ km above mountain systems (orographic effect). Thus, according to the work of Gavrilov and Koval (2013), temperature variations are $10 - 30$ K, which is in order of magnitude consistent with experimental estimates (Sukhodoev and Yarov, 1998; Shefov et al., 1999). Numerical modeling of the temperature disturbance from the tsunami shows a temperature increase of $10 - 12$ K, wave periods ~ 30 min (Artru et al., 2005).

3. Estimation of changes in turbulent viscosity and thermal conduction as a result of a spreading AGW

Calculated changes in temperature of the upper atmosphere and velocity with the altitude caused by AGW gives us coefficient values for turbulent viscosity and thermal conductivity using formulas (1) and (2). Measurements from TIMED satellite were used as background temperature and dynamics changes.

Calculated coefficients of turbulent viscosity are shown in Figure 3. Prandtl number value $\mathrm{Pr} = 6$ and $\mathrm{Cu} = 0.1$ (Izakov, 2007) were used in calculations. As could be seen from the figure, AGW causes the occurrence of turbulent regions within 95 to 105 km.

4. Conclusions

The parameters of the upper atmosphere changes significantly during the transmission of atmospheric gravity waves. AGW in the upper atmosphere results not only in temperature changes but also, due to upcoming impulses, result in a strong change in its dynamic conditions. Atmospheric gravity waves dissipate and transmit impulses, inducing a localized in altitude flat and parallel flows.

The physical mechanism of AGW is their turbulence generation, which in the end, results in the heating of the upper atmosphere due to the dissipation of turbulent energy, as well as in the occurrence of enhanced dynamics, mostly of horizontal, processes. This is evidence of the generation of small-scale turbulence by atmospheric gravity waves. The mechanism of this generation is related to vertical velocity shifts of horizontal wind and increase in temperature as a result of gravity waves.

In addition, turbulent regions will cause local heating and wind motions. Such regions could be the sources of secondary AGW, which propagate down and upwards from the mesosphere.

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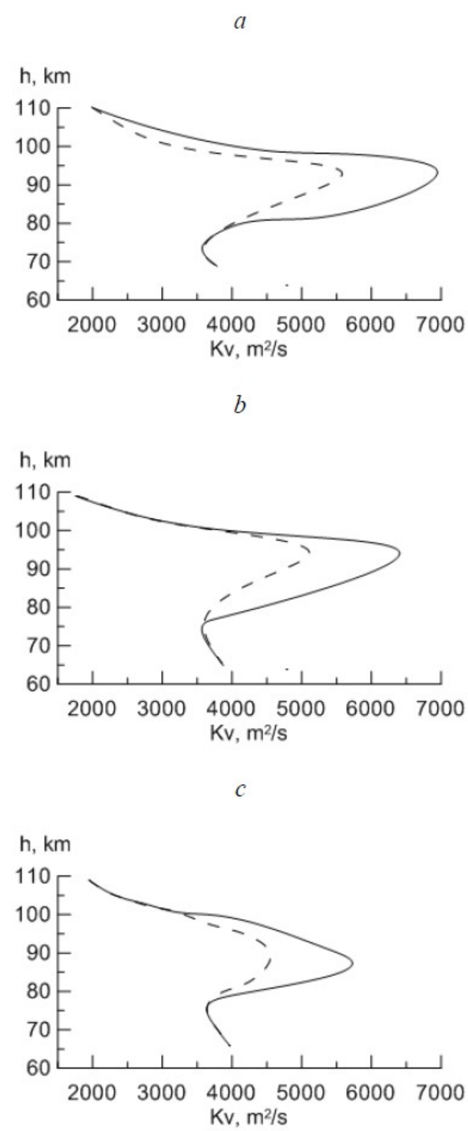

Figure 3: Turbulent viscosity coefficient: dash line - without atmospheric waves; solid line - disturbances caused by waves are accounted (a - 28 August 2005, b - 19 October 2005, c - 18 July 2005).