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OROGRAPHIC DISTURBANCES OF UPPER ATMOSPHERE EMISSIONS

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ABSTRACT

There are some increases of the temperature of the hydroxyl emission ($\Delta T \sim 20$ K, $z \sim 90$ km) and of the intensity of the 6300 oxygen emission ($\Delta I/I \sim 20$ per cent, $z \sim 250$ km) for the lee of the mountains at distances about 150 km in the case of the latitudinal direction of the wind ($U \sim 10$ m/s) at the 3000 m level.

Airflow motions over mountains may be one of the possible processes of generation of wave disturbances penetrating into the upper atmosphere (HINES, 1974; LINDZEN, 1971). The purpose of this paper is to study the penetration of orographic disturbances into upper atmosphere. Airplane measurements of emission variations of hydroxyl and atomic oxygen 6300 A near the Northern Ural mountains were made. Several nocturnal flights have been carried out in March, 1980 and January - February, 1981 at heights about 3000 m along 64° northern latitude in the Ural region. Spectrographs SP-48 with electronic image converters registration for OH ((9,4) and (5,1) bands - 7700 - 8100 A) and OI (6300 A) emissions were used. The zenith region was observed, and exposure time was 2 minutes. This corresponds to averaging of the emission intensities along the airplane trace over a distance of 10 km. Simultaneous measurements of atmospheric temperature variations at the flight altitude have also been made. Data during 13 nights were obtained. The direction of wind was west to east during 5 nights, east to west during 4 nights, and nearly meridional during the rest of the cases. During some nights observations were disturbed and blended by aurora. These data were withdrawn.

In Figure 1 mean variations of the increments of hydroxyl emission rotational temperature ΔT_{90} and relative intensity of oxygen emission 6300 A for meridional and latitudinal wind directions are shown. A distinct increase of hydroxyl emission temperature (maximum intensity level is about 85-90 km) and of oxygen emission intensity (maximum intensity level is about 250-270 km) is clearly seen in the lee of the mountains at distances up to 300 km. There are wave disturbances in the atmospheric temperature with wavelengths about 10-50 km and amplitudes up to 0.4 K in the case of latitudinal direction of the wind at 3000 m level.

A comparison of various characteristics of the lower and upper atmospheric parameters of the measured data is shown in Figure 2. Circles represent the 1980 data, and dots those for 1981. It should be emphasized that numbers of dots and circles are not the same on all figures as for several flights there are no simultaneous data of hydroxyl and oxygen emissions.

Figure 2 shows that there are some connections between processes at the various altitudes. The horizontal distances, X_{90} and X_{250}, of the maximum disturbance of emissions are the same in many cases. However, there are some data of 1980 (circles) when X_{90} is greater than X_{250}. At these times orographic effects were observed to the west of the Ural because of westward wind. The observed decreases of the X_{250} compared to X_{90} may be explained by an eastward zonal component of the thermospheric wind for this season and diurnal period (HILLER, 1981; SEMENOV, 1982).

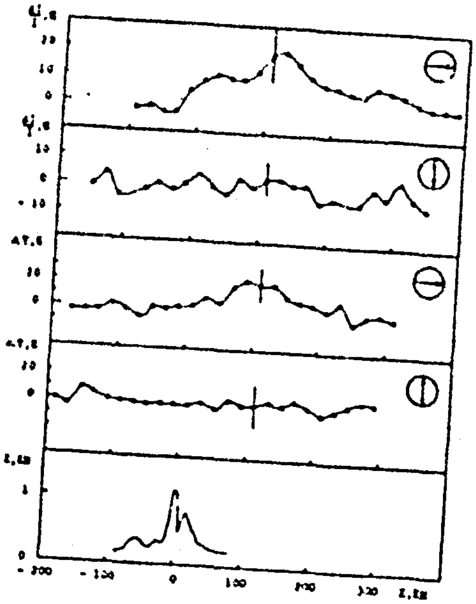

Figure 1. Mean variations of increments of hydroxyl rotational temperature ΔT_{90} near 90 km, and relative intensity $\Delta I/I$ of the 6300 A oxygen emission near 250 km, in dependence on the distance X from the Northern Ural mountains along the geographic parallel (about 64°N) for various wind directions (shown by arrows, $U \sim 10$ m/s). The vertical lines are error bars.

It is also notable that there are correlations between the velocity U of wind at the flight altitude and the measured values of X_{90} and X_{250} for the observed limits of U, as well as correlations between the halfwidths of the regions of disturbed emissions, ΔX_{90} and ΔX_{250}.

The dislocation of Ural mountains and a mesospheric temperature disturbance region reveal some criteria of interconnection between processes in the lower and upper atmosphere and allow to analyse the results of the temperature variation measurements at the flight altitude near 3000 m, distorted by the Doppler effect caused by airplane movement (velocity is about 60-90 m/s).

There are standing and propagating waves in the wave disturbances, as obtained by a harmonic analysis of measured time series of temperatures at the height of 3000 m. The existence of a standing wave has been revealed from the phase coincidence of the waves relative to ground surface during consecutive flights over the mountains. Amplitudes ΔT, horizontal wavelengths λ_x and periods τ for the standing waves observed on January 26, 1981 in the eastward lee of the mountains are given in Table 1. Some estimations of corresponding values of the vertical wavelengths λ_z for standing waves and also the group velocity components C_λ^E and C_z^N in the ground system obtained from the dispersion relation are also presented in Table 1. The wind velocity was 14 m/s. According to HINES (1974), wave disturbances with such parameters can penetrate up to the mesopause.

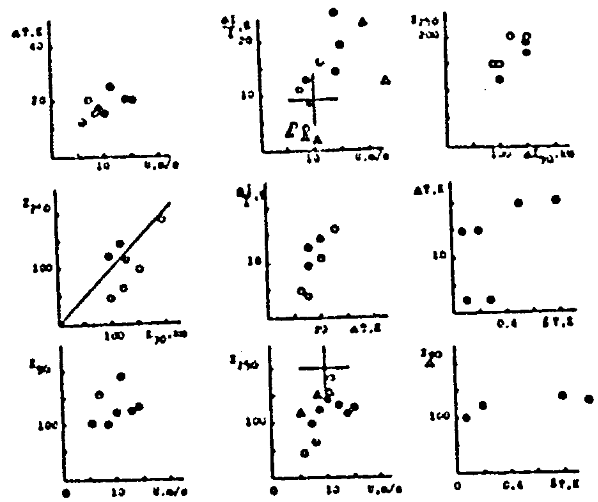

Figure 2. Comparisons of measured data: X, horizontal distance of the maximum increments of the upper atmospheric emission parameters, ΔX, halfwidth of disturbed region ΔT_{90} and ΔI/I, mean amplitudes, U wind velocity and ΔT temperature disturbance amplitudes at the 3000-m level. Circles represent data of 1980, dots are data of 1981. The cross is the result for the Hawaiian Islands, triangles are the same for Yokosuka (see the text).

For all these examples the horizontal components of the group velocity C_x^g are directed downwind, the vertical components C_z^g are directed upward and their values are almost equal. Such wave disturbances may reach the mesopause within 4-8 hours. According to the measurements, disturbances in the upper atmospheric emissions were observed during the whole period of night measurements (i.e., about 7 hours usually). The values of vertical energy fluxes $F_z = \Delta P \cdot \Delta W/2$ (ΔP and ΔW, being amplitudes of pressure and vertical velocity) for measured temperature amplitudes ΔT were computed on the basis of linear theory. The results are also included in Table 1.

It follows from Figure 2 that the values of disturbances at various levels of the atmosphere depend upon the wind velocity in the lower atmosphere. One can note that there is approximate proportionality between the temperature increments, ΔT_{90}, at the height near 90 km (and also the relative increments of the 6300 emission, ΔI/I) and the wind velocity, U. Such type of correlation is in agreement with theoretical estimations of the orographic disturbance amplitude in the upper atmosphere (near 90 km) in certain limits of used parameters (BLANK, 1980).

It is interesting to compare the obtained results with the observed data of the emission behaviour in the vicinity of other mountain ranges. Measurements of the spatial distribution of the 6300 emission intensity near the Hawaii Islands (ROACH et al., 1964; ROACH and GORDON, 1973) revealed a stable spot structure very similar to the geographic structure of these islands. From this fact Krassovsky suggested that the observed emission morphology is caused by

Table 1. Parameters of four wave trains of standing waves for January 26, 1981.

QUANTITY	1	2	3	4
Period τ, Min	12·2	13.5·2.5	19·3	23:4
Horizontal Wavelength λ_x, km	8.5·0.9	9.5·1.0	13.8:1.2	16.3·1.7
Vertical Wavelength λ_z, km	14.0·0.9	9.2·2.2	7.1·0.9	6.9·0.8
Horizontal Group Velocity C_x^g, m/s	7.3·3.3	5.8:2.5	2.9·1.1	2.0·0.9
Vertical Group Velocity C_z^g, m/s	5.5·1	5.8:1	5.0·1	4.3·1
Temperature Amplitude δT, K	0.05	0.09	0.08	0.08
Pressure Amplitude δP, Pa	1	2	2	2
Amplitude of Vertical Velocity δW, m/s	0.1	0.2	0.2	0.2
Vertical Flux of Energy F_z, erg/cm^{2}s	70	200	200	200

orographic disturbances penetrating into the upper atmosphere (SEMENOV et al., 1981). Supposing that the expected amplitude of such disturbances is proportional to the square of the mountain height (BLANK, 1980), the data for the Hawaii mountains (~3000, 4000 and 4200 m) after reduction to the altitude (~900 m) of the Ural have been represented by a cross in Figure 2. The size of the cross corresponds to the uncertainty of the data of the horizontal dimension of the observed emission region (ROACH et al., 1964, 1973), and of the wind in the troposphere, which was assumed to be equal to the seasonal mean for this geographic region. Nevertheless there is a satisfactory agreement between these data.

There are also some examples of simultaneous variations of emissions of atomic oxygen 5577 A (z ~100 km) and 6300 A, observed in Japan (MISAWA et al., 1981). Analyses of the meteorological situation in the troposphere showed that there was wind flowing from west to east with a velocity about 10 m/s at the isobaric levels 850 mbar and 700 mbar. In this case the relief of Honsu Island may be adopted as a single mountain with a height of about 3000 m and a horizontal dimension about 280 km, with its center located about 150 km from the observing station Yokosuka. The observed intensity variations of 6300 A emission can be reduced for mountain height in the same way as in the case of Hawaii Islands. The results are represented on Figure 2 by triangles. They are in satisfactory agreement with the other data. Of course it would be necessary to take into account the hydroxyl rotational temperature data which unfortunately are absent in the paper of MISAWA et al. (1981).

The nature of the 90 km level emission disturbance differs from that for the 250 km level. The increase of the temperature at 90 km is caused by dissipation of waves, generated by the airflow over the mountains. The disturbances at heights about 250 km rise possibly as a result of some dynamical transfer from 90 km up to 250 km level, maybe in consequence of a piston effect. The increase of the disturbance region dimension in case of the 6300 A emission intensity compared with that of the 90 km level is obviously the result of an increase of the diffusion rate at greater altitudes. According to the experimental data the mean ratio of the disturbance region dimensions is about 1.5.

Using the temperature increment at 90 km level one can conclude that the total energy of the disturbance observed along the X axis (in the case of 10 m/s velocity, Figure 2) is

$$\Delta E = \frac{3}{2} N k H \int T_{90}\, dx \sim 1.6 \cdot 10^{12} \text{ erg/cm}$$

where N is the particle density of the atmosphere, H is the thickness of the disturbance layer, k is Boltzmann's constant. With a disturbed region dimension of about 150 km, we get the energy density 10^5 erg/cm^2. The average time for generating a stable disturbance is obviously several hours. Since the recorded variations in the mesopause and thermosphere have existed during the whole period of the flight, the lifetime may be estimated to be 10^4s. Therefore the vertical energy flux of the orographic disturbance is about 10 erg/cm^{2}s. On the other hand, the measured amplitudes of the temperature variations at the flight level give a vertical energy flux of about 10^2 erg/cm^{2}s according to the linear wave theory (see Table 1). Such estimations show that heating of the layer near the mesopause takes a small part of mountain lee wave energy.

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