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A MODEL OF THE INFLUENCE OF NEUTRAL AIR DYNAMICS ON THE SEASONAL VARIATION IN THE LOW IONOSPHERE

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INTRODUCTION

Recently it has become clear that the phenomena in the ionospheric D-region are determined to a great extent by dynamical processes in the strato-mesosphere and lower thermosphere. It is these processes that significantly influence the distribution of the minor neutral constituents and the whole ionic composition of the middle atmosphere, regulating the ionization-neutralization cycle of the D-region. In this respect much attention is paid to the study of the winter anomaly (WA) phenomenon on medium and short radiowaves, in which the meteorological character of the lower ionosphere is most prominent. Significant experimental data about the variations of the electron concentration, N, ion composition, temperature and dynamic regime during WA permit a better understanding of the character of the physical processes in the middle atmosphere. WA is recorded as an abrupt increase of absorption L of the middle and short radiowaves on different winter days caused by a significant growth of N predominantly in the 75-90 km region with a steep gradient in the 80-85 km interval. WA is usually observed at middle latitudes of both the Northern and Southern Hemispheres, and is best recorded in the frequency range $f_i = f \cos i$ = 1-2 MHz, where f_i is the equivalent frequency of the incident radiowave with a frequency f at angle i. This is confirmed by Figure 1 where we have presented the results of the absorption measurements in the Sofia ionospheric observatory, along the paths

Pristina-Sofia (1412 kHz/170 km; f_i = 1.1 MHz)
Greece-Sofia (4050 kHz/320 km; f_i = 2 MHz)

All measurements refer to a solar zenith angle $\chi = 78.5°$ ($\cos\chi = 0.2$).

The seasonal variation shown in Figure 1 allows a quantitative estimation of WA magnitude by calculating the excess of the winter absorption value, L_w, over the summer one, L_s. In this way the L_w/L_s ratio, obtained in experimental and theoretical ways, will serve as a criterion for comparing the different physical models of the middle atmosphere.

The purpose of the present report is to evaluate the influence of the neutral wind on the seasonal variation of the electron concentration N for the altitude interval $90 \le z \le 120$ km, where the ratio ν_{in}/ω_i, of the ion-neutral collision frequency, ν_{in} and the ion gyrofrequency, ω_i decreases from 40 to 1. CIRA-72 is used as a model of the zonal wind.

THEORETICAL MODEL

The distribution of the electron concentration N in ionospheric regions D- and E- is described by the balance equation:

$$\frac{\partial N}{\partial t} = q - \alpha N^2 - \mathrm{div}(N\,\vec{v}_e) \qquad (1)$$

where q is the electron production rate, α the recombination coefficient, and $\vec{v}_e$ the mean electron velocity. Due to the neutrality of electrical charge the following equation holds:

$$\mathrm{div}(N\,\vec{v}_e) = \mathrm{div}(N\,\vec{v}_i) \qquad (2)$$

Figure 1a.

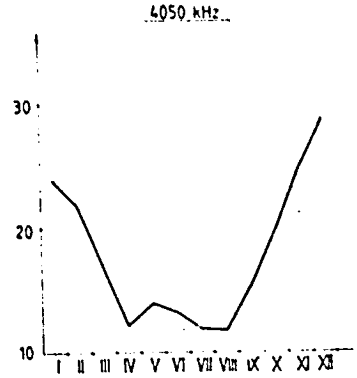

Figure 1b.

i.e. ion velocity should be investigated instead of electron velocity. Then, if we assume the three-component model of the ionosphere (neutrals, heavy positive ions and electrons) it becomes necessary to determine the ion drift velocity as a function of the neutral wind, geomagnetic field and $r_i = \nu_{in}/\omega_i$. If seasonal variations of the neutral wind for different altitudes are known and if the preliminary model for q and α is determined and conditioned by ionospheric chemistry in different months, then it is easy to find the seasonal distribution of the electron concentration.

If we neglect the pressure gradient and electric field forces, the macroscopic motion of heavy positive ions in a collisional ionosphere with geomagnetic field is described by

$$d\vec{v}_i/dt = \nu_{in}(\vec{v} - \vec{v}_i) + (e/m_i)[\vec{v}_i \times \vec{B}] \tag{3}$$

where e and m_i are the charge and mass of the ions, $\vec{v}_i$ the ion drift velocity, $\vec{v}$ the neutral gas velocity, and $\vec{B}$ the local geomagnetic field. Under quasi-steady conditions the ion velocity is determined from (3) as follows:

$$\vec{v}_i = (1 + r_i^2)^{-1} (r_i^2 \vec{v} + r_i[\vec{v} \times \vec{\gamma}] + (\vec{v} \cdot \vec{\gamma})\vec{\gamma}) \tag{4}$$

where $\vec{B} = B_0\vec{\gamma}$ and $\vec{\gamma} \cdot \vec{\gamma} = 1$. From (4) it follows that $\vec{v}_i$ is strongly influenced not only by the different neutral winds at different altitudes, but also by the strong altitude dependence of r_i. When $r_i >> 1$ there is a full drag of ions by the neutrals, and $\vec{v}_i \approx \vec{v}$. This holds between 60 and 90 km. In the height range from 90-100 to 140 km, r_i is of the order of 1, and this region is known as the transition region (AXFORD, 1963). There

$$\vec{v}_i = (1 + r_i^2)^{-1} (r_i^2 \vec{v} + r_i[\vec{v} \times \vec{\gamma}]). \tag{5}$$

Finally, when $r_i << 1$, then $\vec{v}_i = (\vec{v} \cdot \vec{\gamma})\vec{\gamma}$.

The quantity of the ion flow is measured by the convergence of the ion velocity. If $\vec{v}_i$ is given by (5), then

$$\begin{aligned} \text{div}\, \vec{v}_i &= (1 + r_i^2)^{-1} (r_i^2 \text{div}\, \vec{v} + r_i(\vec{\gamma} \cdot \text{rot}\, \vec{v})) + \\ &+ (1 + r_i^2)^{-2} (2r_i\vec{v} + (1 - r_i^2)[\vec{v} \times \vec{\gamma}])\, \nabla r_i \end{aligned} \tag{6}$$

Experiments indicate that the horizontal neutral wind components usually are an order of magnitude or so greater than the vertical component, while the vertical scales appear to be the same for the three components. Hence we can take $\vec{v} = (v_x, v_y, 0)$. We use the standard coordinate system in which the x-axis points to geomagnetic east, the y-axis to geomagnetic north, and the z-axis vertically upward. A horizontally stratified ionosphere is assumed. For altitudes 90-100 km, from (5) and (6) follows

$$v_{iz} = [r_i/(1 + r_i^2)]\, \gamma_y v_x \tag{7}$$

$$\partial v_{iz}/\partial z = [r_i/(1 + r_i^2)]\, \gamma_y\, \partial v_x/\partial z + [(1 - r_i^2)/(1 + r_i^2)^2]\, \gamma_y\, v_x\, \partial r_i/\partial z$$

As we are interested in the seasonal course of N for the respective altitudes, we have to solve the quasi-stationary equation (1):

$$\alpha N^2 + [\gamma_y/(1 + r_i^2)]\; r_i\; \partial v_x/\partial z + \frac{1-r_i^2}{1+r_i^2}\, v_x\, \partial r_i/\partial z + r_i p v_x\; N - q = 0 \tag{8}$$

where the altitude profile of N is taken in the form $N(z) = N_o \exp(pz)$ (VELINOV et al., 1974).

In the height range 115-120 km, the full ion velocity or the equation (5) is used. Then the redistribution of electron concentration is influenced not only by the zonal but also by the meridional wind. Then:

$$v_{iz} = [\gamma_y/(1 + r_i^2)](r_i v_x + \gamma_y v_y) \tag{9}$$

$$\frac{\partial v_{iz}}{\partial z} = \frac{\gamma_y}{1+r_i^2}\; \{r_i \frac{\partial v_x}{\partial z} + \gamma_y \frac{\partial v_y}{\partial z} + \frac{1}{1+r_i^2}\; \frac{dr_i}{dz}\, [(1-r_i^2)v_x - 2r_i\gamma_z\, v_y]\}$$

From (9) it follows that the influence of the meridional wind increases with increasing altitude.

NUMERICAL RESULTS

We are interested in the influence of the neutral wind on the seasonal course of absorption of short radiowaves, where the WA is clearly felt. Our attention is drawn by the courses, shown on Figure 1. As some of the radiowaves are reflected below 100 km, other between 100 km and 110 km, it becomes necessary to consider the altitude range from 90 km to 110 km, and to determine mainly the influence of the zonal winds on the seasonal course of N. Assuming that $L \sim N$, we can compare theoretical and experimental results. From CIRA-72 we obtain a monthly value of the zonal wind and wind shear for altitudes 90, 95, 100 and 110 km, for $\varphi = 40°$N. We include them in equation (8) and solve it. For altitudes 100 km and 110 km we used data for the meridional wind from MANSON et al. (1981), for January, February and March only. Their influence on N is slight. As the absorption is an integral characteristic, in order to be able to compare the seasonal course of L_{1412} with the course of N it is necessary to add the seasonal courses of N, influenced by the zonal wind for the altitudes 90, 95 and 100 km (Figure 2). The influence of the zonal wind on WA is obvious, conditioned by the model of α and q. If the latter increases the winter values of N over the summer values by 16%, then if we add the wind this increase is augmented up to 25%. The experimental results show an increase by 30%. Theoretically obtained seasonal course of N in Figure 2 clearly shows the experimentally obtained secondary maximum in May and minimum in July, as well as the sharp increases of N from October to November, known as "October-effect".

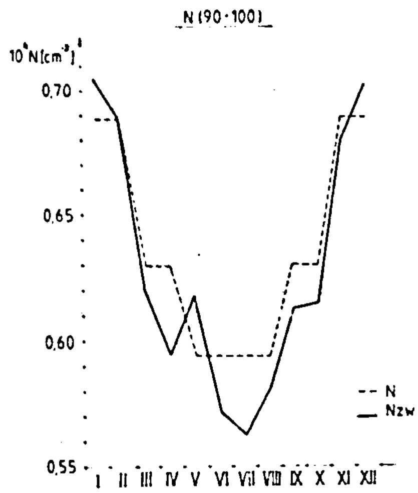

Figure 2.

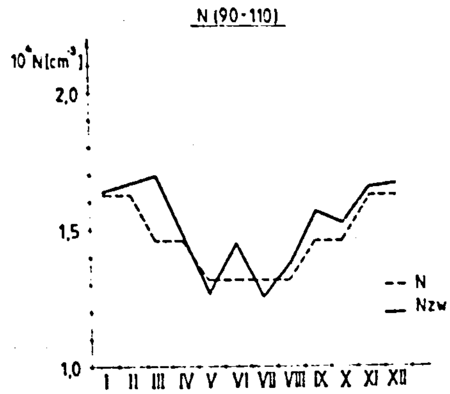

Figure 3.

The same is done for the second radio path, L_{4050}, and the influence of N by the altitude 110 km is added. Figure 3 shows the seasonal course of the average N. The increases of the winter values are about 38%, but considerably lower than experimentally observed. This may be due to the above mentioned model of q and α. The neglect of the influence of the electric field makes itself felt stronger at these altitudes. Figure 3 shows well the summer maximum, which is clearly seen on Figure 1b.

The theoretical results compared with experimental results are encouraging and clearly show the influence of the neutral atmosphere dynamics on the seasonal variation of the electron concentration, N.

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