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INVESTIGATION OF AERONOMIC PROCESSES ON THE BASIS OF IONOSPHERIC SPORADIC E

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INTRODUCTION

The vertical sounding of the ionosphere by means of radio waves is even nowadays the most widely used method of ionospheric research. From the parameters scaled from ionograms only the characteristics of the sporadic E layers are directly related to dynamical processes. It is now generally accepted that at mid-latitudes E_s ionization is produced by wind-shear in the lower thermosphere.

It is also known that the lower thermosphere is an important region of the atmosphere, processes of which affect almost the whole atmosphere. Namely, in the lower thermosphere forms the "surface" called the homopause separating the homosphere from the heterosphere, respectively the turbopause, the boundary between the turbosphere and the diffusion region. The turbopause is defined as the altitude where the coefficient of turbulent diffusion equals the coefficient of molecular diffusion which increases with increasing height. The changes of the height of the turbopause result in variations of the composition of the neutral gas in the upper atmosphere. The altitude changes of the turbopause are mostly connected with variations of the turbulent diffusivity. At the same time, turbulent diffusion is also one of the factors establishing the vertical transport of atmospheric constituents produced in the thermosphere to the mesosphere and vice versa.

Realizing these two facts a method has been developed, by means of which the characteristics of turbulence can be determined on the basis of E_s parameters.

METHOD AND DATA USED IN THE INVESTIGATIONS

It has been assumed that the wind-shear theory of mid-latitude sporadic E is valid. First, the ion-convergence is computed by means of the formula (REDDY and MATSUSHITA, 1968)

$$\frac{dv_{iz}}{dz} = -\alpha_{eff}\, n_{max}\left(\frac{n_o^2}{n_{max}^2} - 1\right)$$

where v_{iz} is the vertical component of the ion drift velocity, α_{eff} is the effective recombination coefficient outside of the stratification, n_{max} and n_o are the maximum electron density within the layer, respectively the electron density in absence of the layer. The maximum electron density of the E_s layer is obtained from the measured blanketing frequency fbE_s. The electron density, appearing in the absence of the layer, that is the background electron density is computed by means of formulas defining ionospheric models (e.g., RAWER and RAMAKRISHNAN, 1972). In these relations the maximum electron density of the E layer was calculated from the simultaneously measured foE value, considering the measured virtual height $h'E_s$ as height of the E layer. Then, the vertical shear of the horizontal wind is determined by means of the equation

$$\frac{du}{dz} = -\frac{1+\left(\frac{\nu_{in}}{\omega_i}\right)^2}{\frac{\nu_{in}}{\omega_i}\cos I}\,\alpha_{eff}\,n_{max}\left(\frac{n_o^2}{n_{max}^2}-1\right)$$

where ν_{in} and ω_i are the ion-neutral collision frequency, respectively the ion gyrofrequency, I being the magnetic dip angle. The ion-neutral collision frequency has been computed from the formula given by CHAPMAN (1956). The number density of the neutral gas, necessary for the determination of the ion-neutral collision frequency has been obtained from CIRA (1972).

For the determination of turbulent parameters, first the gradient Richardson-number

$$R_i = \frac{g}{\theta}\,\frac{\frac{\partial\theta}{\partial z}}{\left(\frac{\partial u}{\partial z}\right)^2}$$

is computed, where the vertical gradient of the potential temperature has been determined on the basis of CIRA (1972). According to the investigations of WOODS (1969) turbulence sets in, if the value of the gradient Richardson number is less, than 0.25. Testing the fulfillment of this condition, the vertical turbulent wind w is calculated by means of the relation (DEACON, 1959)

$$w = [-0.15\,(R_i)^{1/2} + 0.08]u$$

where u is the horizontal wind velocity. The latter can be computed from the thermal wind equation using atmospheric models, or measured values are used.

The turbulent diffusivity is determined by means of the formula

$$K = \frac{\langle w^2\rangle}{\left(\frac{g}{\theta}\frac{\partial\theta}{\partial z}\right)^{1/2}}$$

given by ZIMMERMAN and MURPHY (1977). This relation is valid, if the vertical turbulent spectrum is inertial and limited to the buoyant limited scale. Since the dimensions of inhomogeneities in the E_s layer, inferred from the wave length of radio waves reflected from the layer are between 1.5 and 15 m, this condition is fulfilled also in our case.

For the illustration of the quality of the data, obtained by means of this method, the E_s parameters of the ionospheric stations Bekescsaba, Hungary (46°40'N, 21°10'E) and Juliusruh, GDR (54°38'N, 13°23'E) have been used. Further, the horizontal wind velocity has been determined from the drift velocity measured with the spaced receiver method in Kuhlungsborn, GDR (54°07'N, 11°46'E), extrapolating the data to the height of the E_s layer.

RESULTS AND CONCLUSIONS

Figure 1 shows the turbulent diffusivity deduced from ionospheric sporadic E in different altitudes above the ionospheric stations Bekescsaba and Juliusruh. Here the wind data measured by means of the spaced receiver method in Kuhlungsborn were used assuming that the horizontal wind velocity does not significantly change with latitude. In the Figure the values of turbulent diffusivity given by other authors are also given. It can be seen that our results agree very well with the data of other experiments.

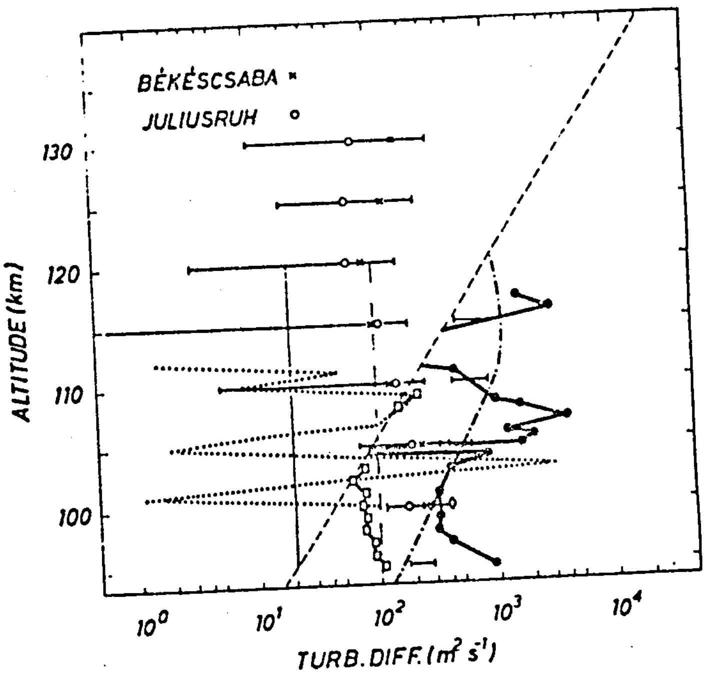

Figure 1.

In Figure 2 the seasonal variations of the turbulent diffusivity deduced from E_s parameters in different altitudes above the ionospheric station Bekescsaba are shown. Here again the drift velocity measured in Kuhlungsborn was used. The turbulent diffusivity displays below 115 km a seasonal variation with maxima in the equinoctial months and minima in summer respectively in winter. Above 115 km a seasonal variation with a minimum in summer and a maximum in winter appears. It should be mentioned that the seasonal variation of the turbulent diffusivit· agrees with that obtained by ROPER (1966) from radio-meteor wind-shear observations.

An attempt has also been made to determine the variation of the turbulent diffusivity during and after geomagnetic disturbances. The turbulent diffusivity was obtained this time by computing the horizontal wind velocity from the thermal wind equation. The mean variations of the turbulent diffusivity below and above the turbopause (100 km, resp. 120 km) during and after 12 geomagnetic disturbances of the year 1973 are shown in Figure 3. It can be stated that the change of the turbulent diffusivity below the turbopause is opposed to that appearing above the turbopause. Since the turbulent diffusivity increases below the turbopause during geomagnetic disturbances, this would indicate a rise of the turbopause and composition changes in accordance with theoretical considerations (SINHA and CHANDRA, 1974) and satellite measurements (PROLSS, 1980).

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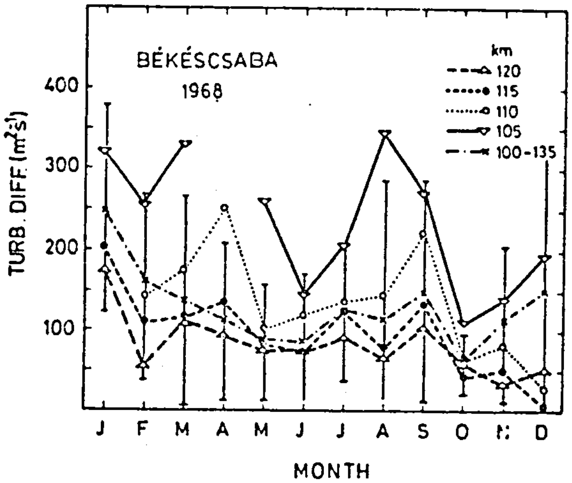

Figure 2.

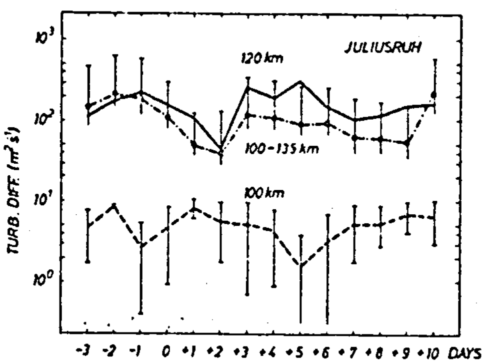

Figure 3.

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