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SOME ASPECTS OF GENERAL CIRCULATION AND TIDES IN THE MIDDLE ATMOSPHERE

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From 80 km altitude upwards, a change in the character of mean zonal circulation compared with stratospheric and mesospheric circulation has been observed by many investigators. Firstly, in radio meteor sounding data from an average level near 95 km in middle latitudes, a semi-annual wind variation was found. While in the main seasons (summer and winter) a westerly wind is observed here, in the interseasonal periods the wind weakens sharply or even reverses into easterly wind. The circulation features of the meteor zone have been discussed in detail in the monograph by PORTNYAGIN and SPRENGER (1978). The recently available possibilities to obtain vertical profiles of the wind from radio meteor data in the altitude range 80-110 km have shown that the features of annual variation at 95 km reflect a sign reversal of the circulation near 100 km. While below 80 km a westerly circulation is found in winter, and an easterly circulation in summer, above 100 km the circulation is easterly in winter and westerly in summer. This can be clearly seen in Figure 1, taken from the monograph by KOKIN and CAIGEROV (1981). The fact of this circulation reversal in itself was not surprising, since it had been mentioned in many papers and monographs. However, it is surprising inasmuch as in none of the theoretical models (e.g., SCHOEBERL and STROBEL, 1978; CLUSHAKOV et al., 1979; DICKINSON et al., 1977) this circulation reversal was obtained.

Analysis of mean daily winds at altitudes above 200 km, determined from radiowave incoherent scatter data, again showed a normal monsoon circulation, as seen in Figure 2 (EMERY, 1978). Thus, one may come to the conclusion that between 100 and 200 km, possibly nearer to 100 km, a layer of anomalous circulation must be situated, at least in middle latitudes.

The understanding of mechanisms which are responsible for the circulation "anomaly" in this height region appears to be of extraordinary interest and importance. Obviously, one had to take into account mechanisms which are usually not considered in hydrodynamics. Let us write the equation of motion of the atmospheric gas, taking into account the presence of a charged component of the atmospheric gas, the presence of the Earth's magnetic field, and the existence of the electric potential field (DOKUCHAEV, 1959; CLUSHAKOV, 1975):

$$2\rho[\vec{\omega}\vec{v}] = -\nabla p - \vec{g}\rho + \frac{1}{c}[\vec{j}\cdot\vec{B}] \tag{1}$$

The appearance of a quantity which is new for hydrodynamics, viz., the current density, $\vec{j}$, requires additional equations to close the system of hydrodynamic equations. Usually this can be done by writing $\vec{j}$ in Ohm's law under conditions of anisotropic conductivity due to the Earth's magnetic field:

$$\vec{j} = \hat{\sigma}\vec{E}' \tag{2}$$

$$\vec{E}' = \nabla\phi + \frac{1}{c}[\vec{v}\cdot\vec{B}] \qquad \hat{\sigma} \text{ – conductivity tensor}$$

ϕ is the electric field potential, the nature of which will be discussed below, and $\frac{1}{c}[\vec{v}\times\vec{B}]$ is the electric induction. For large-scale and rather slow processes ($T > 1$ hour), space charge may be considered as quasi-stationary, i.e.,

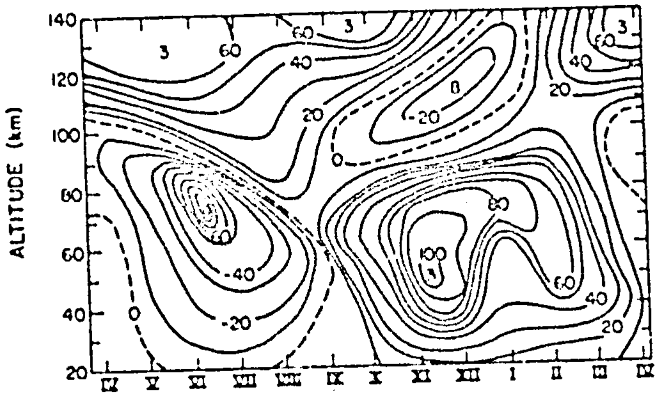

Figure 1. Height-season variation of the prevailing zonal wind in middle latitudes. Solid curve-computed data, dashed curve-experiment.

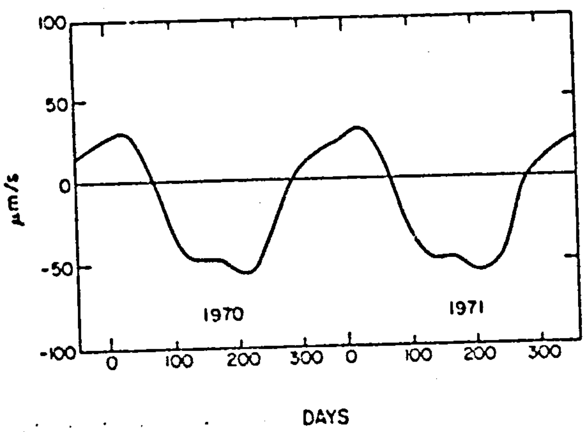

Figure 2. Annual variation of prevailing wind at 300 km altitude over Millstone Hill from incoherent scatter data

$$\mathrm{div}\,\vec{j} = 0 \tag{3}$$

This equation appears to close the system. DOKUCHAEV (1959), neglecting the electric potential field, has shown that the electrodynamic force, $\frac{1}{c}[\vec{j}\times\vec{B}]$, can be divided into the ion friction, in the form of dry (Rayleigh) friction with coefficients $\frac{\sigma_1 B_o^2}{c^2}$ for zonal and $\frac{\sigma_1 B_z^2}{c^2}$ for meridional components, and the magnetic rotary force, coinciding in its sense with the Coriolis force, so that the effective rotation of the Earth may be written in:

$$\omega' = \omega - \frac{\sigma_2 B_o B_z}{2\rho c^2 \cos\theta}$$

Here, σ_1 is Pedersen conductivity, σ_2 is Hall conductivity, and

$$B_o^2 = B_z^2 + B_\theta^2$$

The σ_2 value has a distinct maximum at 110-120 km altitude, and a considerable diurnal variation. In daytime, according to some data, the sign of ω' in the 110-130 km region can become opposite to the sign of ω. If this reversal of the sign ω' took place both in day and night, then, according to the geostrophic relation for large-scale slow processes, with the sign of pressure gradient remaining unchanged, the wind sign would reverse in this altitude region. This is the simplest hypothesis to explain the circulation "anomaly". But, unfortunately, what is most possible, the Coriolis force compensation takes place far from always, and the "anomaly" level of 110-130 km is too high to correspond to experimental facts.

Another hypothesis is connected with peculiarities of the solar UV heating rate altitude profile (see Figure 3). As seen from this figure, the volume heating source decreases with altitude everywhere except for the 90-105 km altitude region where it increases. This anomaly in the heating source profile draws one's attention to an attempt to explain the circulation "anomaly".

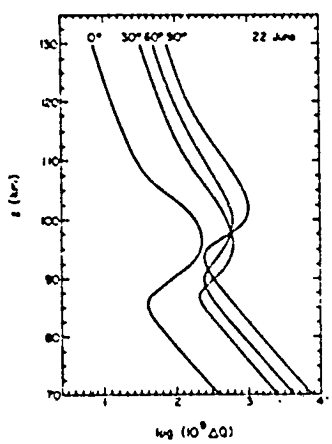

Figure 3. Vertical variation of UV heating rate in the atmosphere for summer solstice conditions at different latitudes.

Having written the system of hydrodynamic equations for a zonal case (without longitudinal dependence) and represented viscosity and heat conduction terms in the form of Rayleigh friction and Newtonian cooling, by means of substitutions one can obtain the equation for pressure latitudinal variations as a function of height

$$\frac{d^2p}{dx^2} + \frac{dp}{dx} - Ip = \frac{\gamma - 1}{\gamma\chi} \frac{dq}{dx} \qquad (4)$$

Where $dx = \frac{dz}{H}$, and q is a volume UV radiation heating source.

$$I = \frac{\eta}{\chi} \frac{gH}{a^2 f_o^2} \lambda \frac{R}{\mu g} \left(\frac{\partial T}{\partial z} + \Gamma\right)$$

η is the Rayleigh friction coefficient, χ is the Newtonian cooling coefficient,

$t_0 = 2$, other symbols are the standard ones. By substitution of $p = Pe^{-x/2}$, equation (4) is obtained in the form:

$$\frac{d^2 p}{dx^2} - b^2 p = \frac{\gamma - 1}{\gamma \chi} e^{x/2} \frac{dq}{dx} \; ; \quad b^2 = \frac{1}{4} + I$$

In the simplified case of absence of the Earth's influence. Green's function of the left-hand side operator is equal to:

$$G(x, \xi) = \frac{1}{b} \exp(-b|x-\xi|)$$

Then the solution is written in the form:

$$P(x) = \frac{\gamma - 1}{\gamma \chi_0} \int_{-\infty}^{\infty} e^{-b|x-\xi|} e^{\xi/2} \frac{dq}{d\xi} d\xi$$

For sufficiently large values of b, the integral kernel assumes the character of a delta function, and P(x) will alter its sign together with the alteration of the sign of $\frac{dq}{dx}$. The sign alteration of the latter quantity, as mentioned above, takes place in the 90-105 km altitude region. Together with this, the circulation sign will naturally change. In the hypothesis suggested the question of the value of b is very important. In order that this value is adequately large, it is necessary that the Rayleigh friction coefficient should be much greater than the Newtonian cooling coefficient. The high level of thermal stability also contributes to an increase of b. It is possible that the large value of b is a specific quality of the mesosphere and thermosphere. Obviously, the mechanism suggested at least partly explains the appearance of the "anomaly" circulation near 100 km.

However, one does not manage to explain the "anomaly" up to the 120-130 km altitude by means of this mechanism, and we will address again the equation (1), taking into account the electric potential field.

Strange as it may seem, the complex case of self-consistent calculation of hydrodynamic characteristics jointly with the electric field for tidal waves has already been done by GLUSHAKOV et al. (1979, 1980, 1981), but the simpler case of zonal stationary flow has not been examined. Let us fill the gap by presenting the results in this paper, in a simplified form only.

Integrating (3) over z in infinite limits (in fact, in a thin ionospheric layer) supposing that $\vec{j}$ vanishes outside the ionospheric layer, we obtain the expression

$$\frac{1}{a \sin\theta} \frac{\partial}{\partial\theta} \sin\theta \int_{-\infty}^{\infty} j_\theta \, dz = 0 \tag{6}$$

It follows from this that $\sin\theta \int_{-\infty}^{\infty} j_\theta \, dz$ C, but the integral current must not flow over the Pole, because this would mean non-zonality. Therefore, C must be zero, i.e.,

$$\int_{-\infty}^{\infty} j_\theta \, dz = 0.$$

Expressing j_θ in terms of Ohm's law, taking into account zonality of the problem, and making some simplifications conditioned by numerical estimates for altitudes less than 130 km, one can obtain

$$E_\theta = \frac{B_p \vec{\bar{U}}_r}{c} \cos\theta$$

$$\bar{U}_r = \frac{1}{\Sigma_1} \int_{-\infty}^{\infty} \sigma_1 U_r dz; \; \Sigma_1 = \int_{-\infty}^{\infty} \sigma_1 dz \tag{7}$$

Here geostrophic wind is determined by

$$U_r = \frac{1}{ca} \frac{1}{\rho} \frac{\partial p}{\partial \theta}$$

For $\bar{U}_r = 30$ m/s and $B_p = 0.5$

$$E_0 = 1.5 \cos\theta \qquad [\frac{mV}{m}]$$

This electric field is adequate to influence strongly the wind. Noncomplex, but expanded calculations yield the expressions

$$U = U_r - \frac{2}{\sqrt{1 + 3\mu^2}} \frac{\Delta l}{l_o} \vec{\bar{U}}_r \tag{8}$$

$$V = -\frac{l_o \nu}{l_o^2 \mu^2 + \nu^2} (U_r - (1 + \frac{\Delta l}{l_o} \frac{2}{\sqrt{1 + 3\mu^2}}) \vec{\bar{U}}_r)$$

Here, $l_o = 2\omega'$ (ω' - the Earth's effective rotation),

$$\nu = \frac{\sigma_1 B_o^2}{\rho c^2}; \quad \Delta l = \frac{\sigma_2 B_o^2}{\rho c^2}; \quad \mu = \cos\theta$$

Assume that U_r increases with height. $\bar{U}_r$ is the result of averaging U_r over height, with $\sigma_1(z)$ as a weighting function having a pronounced maximum near 140 km; so that $\bar{U}_r \simeq U_r(z = 140$ km). Apparently, below this level, as seen from (8), the sign of U and V may reverse with respect to U_r.

Let us now discuss the current state of tidal theory. Historically it has develcped in a way that investigations were made independently in two altitude regions: below 100 km, where dissipative forces are negligible and above 100 km, where these forces are essential.

By 1970 the theory for the region below 100 km was summarized in the monograph of CHAPMAN and LINDZEN (1970). All calculations were based on tidal energy sources connected with the solar UV radiation absorption computed by BUTLER and SMALL (1963). Though explaining many experimental facts, the theory appeared to be incompetent to explain facts connected with semidiurnal tide seasonal variations in mid-latitudes.

In the last decade, theory of tidal seasonal variation has been greatly developed. IVANOVSKY and SEMENOVSKY (1971) pointed out that for the mean zonal circulation, different from the air shell rotation as a whole, the so-called "equivalent depth" in the classic tidal theory depends in a complex way on the seasonal variation of the circulation index. It is important to emphasize that the circulation character in the stratopause region (50 km) influences the tidal seasonal variations in the meteor zone.

Further development of the theory was aimed at improvements of the numerical technique of solving linear equations, taking into account high-latitude profiles of the background zonal wind and temperature by means of the perturbation method (LINDZEN, 1974; WALTERSCHEID et al., 1979, 1980). These calculations qualitatively confirmed the results of Ivanovsky and Semenovsky, but did not address solving the tidal seasonal variation problem.

A revision of calculations of tidal energy sources and their seasonal variations appeared to be extremely important in the theory (FORBES and GARRETT, 1978, 1979). The maximum of the source appeared to be greater than that of BUTLER and SMALL (1963), and the altitude region of maximum values was narrower. While in BUTLER and SMALL's (1963) sources the Hough mode (2,2) prevailed for the new sources, in certain seasons, the modes (2,4) and (2,5) are essential.

WALTERSCHEID et al. (1980) made tidal calculations for a new source already. So it seemed that all the necessary work had been done to complete the tidal seasonal variation theory in the altitude region up to 100 km.

A paper by KAIDALOV (1979) disproved this opinion. Kaidalov drew attention to the fact that the perturbation method used by LINDZEN (1974) and WALTERSCHEID et al. (1979, 1980) is good for tidal modes which do not change their character when propagating upward or being trapped. In contrast, the main modes of the semidiurnal tide are on the boundary between propagating and trapped ones, being trapped in one altitude region but propagating in others, the boundaries between these regions being dependent just on the height-latitude distribution of mean zonal wind and temperature. Zero-order approximation in the perturbation theory. not taking into account effects of background wind and temperature for all approximations, defines the tidal waves character as trapped or propagating and leads to inadmissibly large errors in the tide calculation. KAIDALOV (1979) developed a numerical technique to solve the tidal model, taking into account seasonal variations of background wind and temperature without the short-comings of the perturbation method.

Considering tides above 100 km, we will limit ourselves to a few problems arising here. In its most complete form, modelling of tidal variations in the thermosphere is based on the equation of motion (1), Ohm's law (2), the equation of quasistationary space charge (3), together with the equations of energy and continuity. Usually, all variations are examined in linear approximation. Dissipative forces are molecular and turbulent heat conductivity, and ion friction. The contribution of molecular and turbulent friction is estimated to be insignificant, at least, up to the 200-250 km height.

In a majority of theoretical investigations the electric potential field is excluded and, as a rule, a torsional effect by the magnetic rotary force compensating the Coriolis force is neglected. This means that, instead of self-consistent examination, combining hydrodynamic equations with electrodynamic equations, a purely hydrodynamic approach is used (e.g., GARRETT and FORBES (1978)).

Taking into account the electric field and magnetic rotary force generates a series of new effects. Estimation of conclusions with regard to these effects, using experimental data, permits one to evaluate more reasonably their contribution to the tide formation in the thermosphere.

Partial or even complete compensation of the Coriolis force by the magnetic rotary force at 110-130 km altitudes would result in the formation of sharp wind maxima at these altitudes. In numerical experiments this maximum is well pronounced, as it can be seen in Figure 4, taken from theoretical papers by GLUSHAKOV et al., (1979, 1981). After excluding the magnetic rotary force the effect vanishes. These wind maxima are also well observed in luminous trail experiments (ANDREEVA et al., 1981).

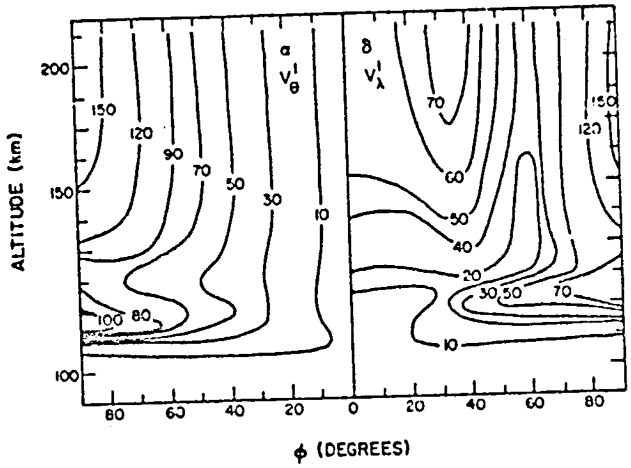

Figure 4. Height-latitude cross-section of the amplitude of zonal (a) and meridional (b) tidal wind diurnal component. In the 120 - 130 km altitude region at high latitudes tidal wind amplitude maxima are clearly seen.

It is of interest to connect the sporadic layer E_s appearance with this wind maxima. This is especially interesting as, according to the calculation, the height of wind maximum increases from lower towards higher latitudes, as it is observed for the height of the layer E_s.

Calculations show that taking into account the electric field may appear to be essential in the tidal wind pattern at altitudes above 130-140 km. This can be seen from Figure 5. It should, however, be mentioned that the electric field contribution in the tidal wind formation essentially depends on the electron concentration model assumed. The increase of electric field contribution to the tidal wind formation is caused by the fast increase with height of the electric forces term $\frac{\sigma_1}{\rho c}$ [E×B]. A sharp growth of the σ_1/ρ value with height begins near 125-130 km, and this quantity reaches approximately a constant value near 160-180 km. On the other hand, the forces term connected quasi-linearly with the pressure gradient also increases with height, prevailing over the electric force below 140 km, but being inferior to it at altitudes above 140 km.

Calculations by GLUSHAKOV et al. (1979, 1981) show that in the formation of semidiurnal tidal winds in the region from 100 to 200 km, tidal oscillations propagating from the mesosphere prevail. These propagating oscillations influence also the rather considerable variations of relative temperature at the altitude near 110 km both for diurnal and, especially, for semidiurnal tides. Estimations of the electric field contribution may be considered to be valid only, if not only the thermodynamic parameters, but also the calculated electric field corresponds to direct measurements. Estimations of the electric field connected with the dynamo-effect have been made from drift velocities of ions in the ionospheric F_2 layer. In Figure 6 a comparison of the electric fields calculated theoretically and derived from drifts measurements according to data of incoherent radiowave scatter data (RICHMOND et al., 1980) is presented. It shows qualitative similarity of these fields.

Calculations made and their comparison with experiment remove a shadow of doubt about the considerable contribution of the electric field in the formation

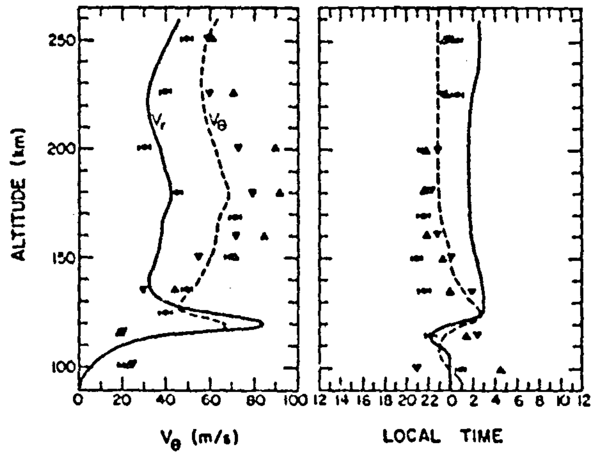

Figure 5. Vertical profile of tidal wind diurnal component amplitude (a) and phase (b):

________ without electric field accounted for

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Δ, ∇, ↔ experimental data

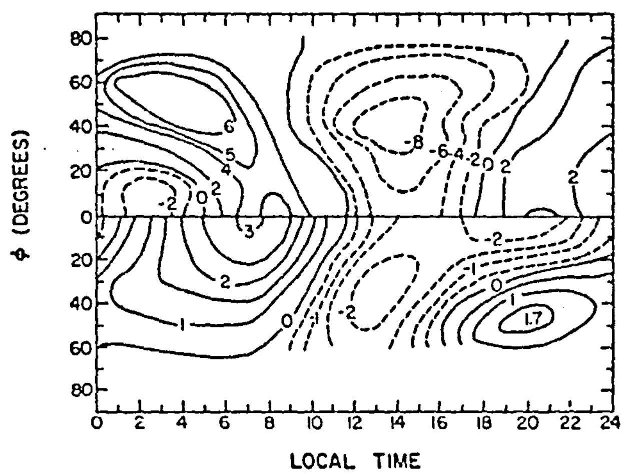

Figure 6. Time-latitude variation of electric field. At the top: theory, at the bottom: empirical model of RICHMOND et al. (1980).

of tidal oscillations of thermospheric parameters. However, only first steps have been made in the investigation of these new effects, and further efforts are needed in this direction.

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