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ESTIMATES OF BOUNDARY LAYER PARAMETERS IN PLANETARY ATMOSPHERES OF THE TERRESTRIAL GROUP

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SUMMARY

The similarity theory of atmosphere boundary layer is applied to the estimate of the form of vertical profiles of average wind velocity and potential temperature in the atmospheres of planets of the terrestrial group in day and nighttime conditions.

It is then considered, as also for the Earth, that the magnitude of the turbulent heat flow q_T in daytime is about 0.1 of $q(1 - A)$, where q is the solar constant for the planet A is its albedo, and in nighttime q_T is still by several factors less. The dynamic velocity U^* is taken equal to 2 — 5 percent (depending upon the stratification) of the mean wind velocity in the free atmosphere, which was adopted after the computations of the work [5].

The boundary layers in the atmospheres of Mars and Venus and in the hypothetical atmosphere of Mercury are examined in detail. Sharp temperature drops are characterstic of Mars within the bounds of a few tens of meters, attaining several tens of degrees, which is caused by the low density of its atmosphere.

For Venus, owing to very high atmosphere density, the stratification is close to neutral, i.e., the temperature profile is close to the adiabatic one.

Owing to high winds, the stratification on Mercury must also be close to neutral with respect to the wind (the profile being close to the logarithmic), but because of low density, such temperature drops may be very great.

* * *

1. The theory of atmosphere's boundary layer has been by now sufficiently well worked out [1—4]. A broad emprical material has been assembled in terrestrial conditions, particularly in the lower part of the boundary layer, that is, the near-ground layer, corroborating the conclusions of theory. For the terrestrial atmosphere the main direction of research is the obtaining of estimates of turbulent flows of the quantity of motion $\underline{\tau}$ and heat q_T according to measurement data of mean velocity u(z) and temperature T(z) profiles for the near-ground layer or after the data on the geographic wind velocity U_g and on the

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potential temperature drop $\delta\theta$ for the boundary layer of the atmosphere, in which Coriolis forces already exert a substantial influence. It would be interesting to obtain for other planets if only rough estimates of the course in their boundary layers of mean temperature and velocity profiles.

Do we dispose at the present time of data, required for such approximate estimates of the structure of the boundary layer on other planets ? It seems to us that such data are already available. The mean charcteristic velocities of motions in the atmospheres of planets have already been estimated in the work [5]. Consequently, we may evaluate the dynamic velocity $u_* = \sqrt{\tau / \rho}$, where ρ is the density. Depending upon the stratification [4], in the terrestrial atmosphere we have $u_* / U_g \simeq 2 - 5$ percent (the first numeral being related to strong stability, and the second to strong instability, i. e. convection).

On the magnitude of the second parameter, determining the structure of the boundary layer, i. e. on the turbulent heat flow, there exists a limitation from above: it cannot exceed $q(1 - A) = q_A$, where $\underline{q}$ is the solar constant for the planet, A is its albedo. For the Earth, even in conditions of developed convection, the ratio q_T/q_A is of the order of 0.1. At stable stratification, when the atmosphere is warmer than the ground, which is usually observed at night, $q_T < 0$, i. e., the heat flow is directed toward the soil and the modulus of the ratio q_T/q_A is generally by several factors smaller than in daytime. For other planets (Mars, Venus, and, perhaps, Mercury) there is no basis to expect too substantial deflections from the regularities, achieved for the terrestial atmosphere. Moreover, one may qualitatively estimate toward which side such deflections could be acting in these planets.

Therefore, there stands a problem before us, which in a certain sense is inverse by comparison with the terrestrial atmosphere: having some kind of re presentation on the magnitude of fluxes of quantity of motion and heat, one must estimate the thickness of the boundary layer and determine the mean vertical profiled of velocity and temperature.

2. According to the general theory [1 - 3], the structure of turbulence in a temperturewise stratified medium is determined by the following parameters: $q' = q_T / c_p$ the normalized turbulent heat flow, $u_* = \sqrt{\tau / \rho}$ — the dynamic velocity, and the buoyancy parameter $\underline{g\beta}$, where $\underline{g}$ is the gravitation acceleration, β is the volumetric expansion coefficient, equal to $1/T_0$ for an ideal gas, where

T_0 is the characteristic temperature of the medium. From these parameters one may construct the scale of length

$$L = -u_*^3 / (\kappa g \beta q_T / c_p \rho), \tag{1}$$

usually called the 'Monin-Obukhov' scale, and the scale of temperature

$$T_* = q_T / c_p \rho \kappa u_*, \tag{2}$$

where κ is the Karman constant.

The vertical profiles of the mean velocity and potential temperature $\theta = T + \gamma_a z$, where γ_a is the adiabatic temperature gradient, are universal functions of dimensionless height $\zeta = z / L$, whereupon

$$u(z) = \kappa^{-1} u_* \, [f_u(z / L) - f_u(z_0 / L], \tag{3}$$

$$\theta(z) = \theta_0 + T_*[f_\theta(z / L) - f_\theta(z_0 / L], \tag{4}$$

where z_0 is the height of roughness. For the universal functions f_u and f_0 we have the following expressions [2, 3]:

$$f_u(\zeta) = f_0(\zeta) = \begin{cases} \ln \zeta + \beta\zeta, & 0 < \zeta, \\ \ln|\zeta| + \beta'\zeta, & \zeta_1 \leqslant \zeta \leqslant 0, \\ a + C\zeta^{-1/3}, & \zeta < \zeta_1. \end{cases} \tag{5}$$

According to careful statistical processing of a vast empirical material, conducted in [6], $\kappa = 0.43$; $\beta = 9.9$; $\beta' = 1.45$; $\zeta_1 = -0.16$; $a = 0.24$; $C = 1.25$.

This formulas are valid for the near-ground layer, where one may neglect the variation of τ and q_T with height. An estimate is given in [2] of the thickness H of the near-ground layer, conducted from that view point:

$$H < \alpha u_*^2(0) / l U_g, \tag{6}$$

where $\alpha = [u_*^2(0) - u_*^2(H)] / u_*^2(0)$ is the relative variation of friction stress τ, l is the Coriolis parameter, U_g is the geostrophic wind velocity. For the terrestrial atmosphere, we obtain at $\alpha = 20\%$ and $u_* / U_g \approx 5\%$, $H \approx 50$ m. For Mars, with the same allowance for α, we obtain $H \approx 100 - 200$ m, since the Coriolis parameter has the same value, while the mean wind velocities are two to four times higher [5]. For the slowly-rotating Venus and Mercury we may take for thickness of the near-ground, or, to be more precise, of the boundary layer, the altitude, at which the wind velocity is comparable with that in the free atmosphere. Usually, as will be seen below, this thickness is of the order of a few units of Monin-Obukhov's scale L.

For the Earth and Mars, one may determine the planetary boundary layer, inside which the wind velocity varies in modulus by comparison with the near-ground layer, but where, owing to the action of Coriolis forces, a notable wind turn with altitude takes place. The thickness of this layer may be determined as [4]:

$$L_* = \kappa u_* / l. \qquad (7)$$

For the Earth, L_* is of the order of 1 km, for Mars, it is 2 to 4 times greater. The wind's rotation angle with height depends on the stratification parameter $\mu = L_* / L = \kappa^2 \beta T_* / l u_*$. In terrestrial conditions [4], the total angle of wind rotation with height is of the order of several degrees at convection, and attains approximately 40° at strong stability (rise of potential temperature with altitude).

A parameter, entirely unknown for other planets, enters into formulas (3) and (4), which is the height of dynamic roughness z_0 of planet's surface. Fortunately, it enters under logarithmic sign, and, for this reason, for our objectives if only an approximate estimate of its magnitude is sufficient. In terrestrial conditions we have, on the average, for dry land $z_0 \approx 1$ cm, for oceans, depending upon swells, it may be substantially lower; even for the forest $z_0 \lesssim 1$ m. Bearing in mind that for other planets there are neither oceans nor forests, we shall assume that there $z_0 \approx 1$ cm.

Being aware of temperature and velocity profiles, we may determine the stability parameter, namely, the Richardson number

$$R_i = g\beta(d\theta / dz) / (du / dz)^2 = \zeta\phi(\zeta), \qquad (8)$$

where the universal function $\phi(\zeta)$ is determined as

$$\phi(\zeta) = \kappa z u_*^{-1} du / dz = z T_*^{-1} d\theta / dz \qquad (9)$$

At the same time, it is considered that the coefficients of turbulent exch change for the momentum K and heat K_T, introduced according to equalities

$$\tau = \rho K(du / dz), \quad q_T = -c_p \rho K_T(d\theta / dz),$$

are identical. Note that at strong stability this is specifically not so, and then, one must introduce in the denominator of the right-hand part of formula (8) the multiplier α, which is the inverse Prandtl turbulence number ($\alpha = K_T/K$). The universal function for the velocity and temperature will also differ by that multiplier ($f_u = \alpha f_\theta$, see [3]). In view of the great uncertainty of a series of other factors and estimatory character of the present work, we shall not take here this effect into account.

The turbulent exchange coefficient $K = \kappa u_* L$ Ri is expressed by the following formulas:

$$K = \kappa u_* z, \qquad |L| \to \infty, \tag{10}$$

$$K = \kappa u_* z(1 + \beta z / L)^{-1}, \qquad |L| < \infty, \tag{11}$$

$$K = 3C^{-1} u_* z(z / L)^{1_3}, \qquad \zeta = Z / L < \zeta_1. \tag{12}$$

3. Compiled in Table 1 are the values of solar energy flux q_A, arriving to planet's surface: these values are for Mars, Venus and Mercury, and for the sake of comparison, for the Earth, and of characterstic scale of temperature T_* and velocity of atmospheric motions U, borrowed from [5], of normalized turbulent heat flow $q_T' = q_T / c_p\rho$, of buoyancy parameter g / T_0 and of Monin-Obukhov scale L. The value of q_T / q_A was taken equal to 0.1, which,,as was earlier noted, is valid in the case of terrestrial atmosphere for noon time in conditions of strong convection. During the night q_T and T_* will be several times smaller, with another sign, and $|L|$ will be by several factors greater. In the morning and evening the stratification becomes close to indifferent, and then $L \to \infty$, i. e. the boundary layer becomes logarithmic. For Mars we assumed the minimum atmosphere model with pressure at its surface $p_0 = 5$ mb, and for Venus — with $p_0 = 100$ atm. For hypothetical atmosphere of Mercury we considered that $p_0 = 1$ mb. The ratio u_* / U was taken equal to 3 percent.

TABLE 1

PLANET	q_A, cal/cm·min	U, m/sec	q'_T, deg.cm/s	T_*, deg	$g\beta$, cm/sec^2.deg	−L, m
MARS	0.6	40	600	10	2	50
VENUS	0.9	3	0.05	0.01	1.2	500
MERCURY	12	200	10	40	1	600
EARTH	1.2	10	7	1	3.3	20

The data of Table 1 show that the basic parameters determining the structure of the near-ground layer, the dynamic velocity u, and particularly the scale of temperature T_* for the planet under consideration differ strongly owing to sharp differences of the fundamental atmosphere parameters, and, in the first place, of density, so that the near-ground layer on each planet must have its own well expressed singularities.

Let us now pass to detailed examination of these singularities.

MARS

The dynamic and, more particularly, the thermic structure of the lower part of the atmosphere of Mars was considered at fairly great length in [7]. Here vertical profiles of temperature and the convection conditions were computed in detail for various latitudes and seasons, and even times of the day. Incidently, the very same estimate of the mean wind velocity of 40 m/sec for a model atmosphere with $p_0 = 5$ mb was obtained there by another method than in [5]. However, the vertical profiles of the mean wind in the numerical model considered there could not be found, and the authors limited themselves to a very rough estimate of Richrdson numbers for various conditions.

In conditions of convection the "logarithmic + the linear law" for wind and temperature profiles (5) is valid to values $\zeta_1 = -0.16$, i.e. at L = -50 to 8 m altitude from planet's surface. At the same time ($z_0 = 1$ cm)

$$u(z) = 3[\log(100z) - z / 35],$$

$$O(z) \quad T(z) = T_0 + 10°[\log(100z) - z / 35],$$

where u(z) is expressed in m/sec and z in m. At 8 km altitude, $u \approx 20$ m/sec, and $\Delta T = T(0) - T(8\text{ m}) \approx 70°$. At the same time the number $Ri \approx -0.03$. Therefore, over the extension of an atmosphere layer of the order of 10 m in all, the velocity attains about one half of its value, characteristic of the free atmosphere, while the temperature jump reaches 70° ! (here the difference between the usual temperature T and the potential, equal to $\theta = T + \gamma_a z$, where for Mars $\gamma_a \approx 4.7°\ \text{km}^{-1}$ is entirely insignificant). Very sharp, though somewhat lesser temperature variations in daytime and lowermost atmosphere layer were also found in [7]. Note that such sharp variations as in our case could not be obtained there, for in the calculation the adopted vertical spacing was there of 100 m.

Above 8 m it was necessary to make use of the last formula (5), describing the condition of free convection. At the same time the mean velocity approaches asymptotically its limiting value of wind velocity in the free atmosphere. According to [3], this takes place for $|\zeta| \approx 5$, i. e. $z \approx 250$ m. The calculation by the last formula (5) shows that for $\zeta \approx 3$ the velocity reaches about its magnitude at infinity. The turbulent exchange coefficient rises rapidly with height. For $z \approx 250$ m we shall have, according to (12), $K \approx 10^7$ cm^2/sec. In [7], for the lower kilometer layer a value $K \approx 10^8$ cm^2/sec was obtained. We may see that both these estimates are to some degree in accord.

At stable stratification (night) we shall assume $L = 250$ m, $T_* = 2°$. Then the velocity 40 m/sec will be attained at an altitude of about 200 m, and the temperature drop will then be 40°. The turbulent exchange coefficient will be of the order of 10^5 cm^2/sec, as is shown by [11]. According to [7], at such a value of K effects of radiation attenuation of temperature can prove to be already substantial. This must lead to a certain decrease of q_T, i. e. to the increase of L, and, by the same token, to a decrease of temperature drop for the given altitude interval. Note that in [7], the temperature profiles during the night were determined from purely radiational computations, completely ignoring the turbulence.

In both cases, by modulus, the velocity above the level of about 200 m will already vary little, but the wind will turn with the altitude, approaching at 2 – 4 km altitude the geostrophic one. In daytime, the total rotation angle is small (a few degrees), while in nighttime it may reach several tens of degrees.

V E N U S

Owing to great values of $|L|$ one should expect that profiles of potential temperature and velocity should be close to logarithmic. Let us estimate at the outset at what altitude the velocity, computed by formula $u(z) = \kappa^{-1} u_* \log(z/z_0)$ is comparable with the mean velocity of 3 m/sec. This altitude is estimated bu the formula $z = z_0 \exp(\kappa u / u_*)$. Hence, for $z_0 \approx 1$ cm, we have $z \approx 1.5$ km. At the same time, the variation of potential temperature at a distance of the order of 1.5 km will constitute an entirely insignificant quantity $\Delta\theta \approx 0.1°$, i. e. the temperature profile must be adiabatic with a high degree of precision. At finite value of L a certain departure from purely logarithmic profiles still can be observed. It will be manifest mainly in the lowering of the altitude, at which the velocity of 3 m/sec is attained. Thus, at stability (night), when $|q_T|$ is several times smaller than in daytime, say 4 times, $L \simeq 2000$ m and the velocity profile will have the form $u(z) = 0.25(\log 100\, z + z/200)$, where z is expressed in m and u(z) in m/sec. The velocity $u(z) = 3$ m/sec will be attained for $z \simeq 350$ m. At the same time, $\Delta\theta \approx 0.03°$.

At instability (daytime) $L = -500$ m and the "logarithmic + linear law" will be observed to $z = -0.16\, L \approx 80$ m. At this altitude the velocity will reach 2.2 m/sec. Above this level the velocity profile (and that of potential temperature) will be described by the law $z^{-1/3}$, and if one considers that the constant value of velocity is attained for $\zeta \approx 5$, we shall have for the thickness of the layer encompassed by convection $z \approx 2.5$. At that distance the variation of potential temperature will be of the order of 0.1°.

Therefore, owing to great thickness of Venus' atmosphere at relatively small flux of incident solar radiation, the state of the atmosphere must be close to neutral, i. e. the profile of temperature must be adiabatic and that of velocity — logarithmic. The situation will change little even if we take the limiting, and entirely unrealistic case $q_T = q_A$. But, generally speaking, one must bear in mind that if part of the visible radiation is absorbed in the atmosphere of Venus, it is quite probable that our assertion on the closeness of stratification to neutral will be the more so correct. But if all radiation is absorbed in planet's atmosphere, the boundary layer will be purely logarithmic.

The coefficient of vertical turbulent exchange at the altitude of the order of 1 km will be of the order $K \approx \kappa u_* z \approx 4 \cdot 10^5$ cm^2/sec. Let us note that for the scale $L = 1$ km we obtained by the Richardson-Obukhv formula a quantity of same order: $K \approx 0.1\, \varepsilon^{1/3} L^{4/3}$, if we had taken $\varepsilon \sim 0.1$ cm^2/sec^3, which is the quantity obtained in [5]. from global estimates of the effectiveness of Venus' atmosphere as a whole in the reprocessing of the power of solar energy, arriving to the planet into a mechanical power, i. e. into the generation of kinetic energy of atmospheric motions. Then $K \approx 2 \cdot 10^5$ cm^2/sec.

M E R C U R Y

If does indeed exist on that planet, the processes in it will prove to be the most exotic among other planets of the Solar system. The enormous wind velocities, of the order of 200 m/sec, the differences in temperatures between the dark and illuminated sides of the planet of the order of 500°, the length of the day and night corresponding to 180 terrestrial days, all this is instrumental in rendering the boundary layer specific too. Owing to great friction, say, $u_* \approx 12$ m/sec ! the characteristic Monin-Obukhov scale L is proved to be very great also. This is why the velocity profile is found to be close to logarithmic although certain small deflections from the purely logarithmic profile may take place. The velocity reaches a magnitude of the order of 200 m/sec over the extent of the lower 100 — 200 meters. At the same time, the high value of T_* results in high temperature drops, Thus, in daytime, the jump of potential temperature, determined by formulas (3) - (5), using the parameters of Table 1, yields $\Delta T \approx 400°$ over the extent of the lower 100 m (at surface temperature of the order of 650°); in the nighttime, with surface temperature of the order of 150°, this drop is of the order of 50° for the lower 200 m. These estimates appear to be extremes, for owing to a very long duration of the day and of the night, the radiation may , apparently, strongly diminish the temperature drops, i. e., on Mercury, the fraction of q_T / q_A is probably notably less than 0.1. At the same time, the scale of L will be still greater, i. e. the velocity profile will be still closer to logarithmic, and the boundary layer will be correspondingly thicker.

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