INFLUENCE OF METEOROLOGICAL CONDITIONS ON VERTICAL PROFILES AND FLUXES OF ATMOSPHERIC TRACE SPECIES

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The physical and quantitative analysis of the influence of meteorological conditions on vertical distribution of trace gases has been performed on the basis of the relationships predicted by the similarity theory. The tables have been composed for the gas–exchange coefficients within a wide range of meteorological conditions and the characteristics of the Earth's surface. The gas–exchange coefficient and the turbulent flux of atmospheric trace species can vary by several tens times under the effect of the above factors.

Within the scope of recent studies of air pollution and monitoring of the environment the analysis of some peculiarities in the trace species distribution under different meteorological conditions is of urgent importance. Some physical and quantitative aspects of this analysis are discussed in this paper.

It is known that the main portion of anthropogenic species is concentrated in the atmospheric boundary layer at heights up to several hundreds of meters. Therefore we analyze here only the relationships obtained in Ref. 1 as a result of applying the similarity theory to the problem of atmospheric pollution.

From the equation of motion, heat inflow, and contaminant transfer in the boundary layer and by generalizing the theory for the case of nonequilibrium stratification, we derive the following expressions for a trace species concentration (q), air temperature (T), and wind velocity (c):

$$q(z) = q_1 + q_* \ln(\eta/\eta_1); \quad (1)$$

$$T(z) = T_2 + T_* \ln(\eta/\eta_2) - \gamma_a(z - z_2); \quad (2)$$

$$c(z) = (u_*/\varkappa)\ln(\eta/\eta_0). \quad (3)$$

Here q_* and T_* are the scales (characteristic values) of q and T variation in the boundary layer, connected with turbulent fluxes of species, $Q_q(0)$, and heat fluxes, $Q_T(0)$, by the relationships:

$$q_* = -Q_q(0)/(\varkappa \rho_1 u_*);\ T_* = -Q_T(0)/(\varkappa c_p \rho_1 u_*), \quad (4)$$

where η is new variable related to the height z by the relationship:

$$\eta = \exp(z/L_*) - 1; \quad (5)$$

q_1 and T_2 are the trace species concentration and air temperature at the levels z_1 and z_2, respectively; u_* is the dynamic velocity related to the wind velocity c_1 at the level z_1 by the expression:

$$u_* = \varkappa c_1/\ln(\eta_1/\eta_0); \quad (6)$$

η_1, η_2, and η_0 are the values of η at z_1, z_2, and z_0, respectively (z_0 is the parameter of roughness); L_* is the scale of the atmospheric boundary layer (Monin–Obukhov):

$$L_* = \frac{u_*^2 T_2}{\varkappa^2 g T_*} = -\frac{c_p \rho_1 u_*^3 T_2}{\varkappa g Q_T(0)}; \quad (7)$$

$\varkappa = 0.38$ is the constant; ρ_1 is the air density at z_1; g is the acceleration of gravity.

If formula (2) is written for a certain altitude z_3, at which the temperature T_3 is measured, and thus determined scale T_* is introduced into the relationship (7), then, taking into account Eq. (6), we derive the following expression for the scale L_*:

$$\frac{z_1}{L_*} = \frac{\ln^2(\eta_1/\eta_0)}{\ln(\eta_3/\eta_2)} B, \quad (8)$$

where the dimensionless parameter B is introduced to represent the finite-difference analog of the Richardson number:

$$B = \frac{g z_1}{T_2} \frac{(T_3 - T_2) - \gamma_a(z_3 - z_2)}{c_1^2}. \quad (9)$$

If the heights z_2, z_1, and z_3 are chosen so that $z_2 = z_1/n$ and $z_3 = n z_1$ (where n is an arbitrary number), then the right-hand side of Eq (8) proves to be dependent, in addition to B, only on z_1/L_* and z_0/z_1; the ratios z_3/L_*, z_2/L_*, and z_0/L_* involved in η_3, η_2 and η_0 should be written as:

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$z_3/L_* = n\, z_1/L_*; \quad z_2/L_* = z_1/(n\, L_*);$

$z_0/L_* = (z_0/z_1)(z_1/L_*).$

The results of z_1/L_* calculations at different z_0/z_1 and $n = 2$ (i.e., $z_3 = 2z_1$ and $z_2 = z_1/2$) are given in the figure.

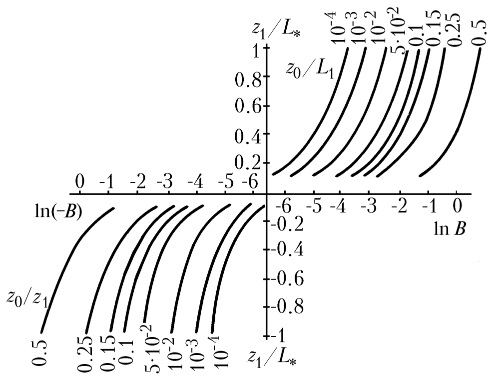

FIG. The dependence of z_1/L_* on B and z_0/z_1.

The exponential-logarithmic formulas (1)–(3) show the basic peculiarities of altitude distribution of the meteorological parameters (wind velocity, air temperature, mass ration of water vapor) and the species concentration.

We use the series expansion of the exponent

$$\eta = \exp(z/L_*) - 1 = z/L_* + z^2/(2L_*^2) + z^3/(6L_*^3) + \ldots \quad (10)$$

At low heights and at any values of the parameters B and L_* the distribution of q and c, as well as of T is described by the logarithmic formulas (laws):

$$q(z) = q_1 + q_* \ln(z/z_1), \quad (11)$$

$$c(z) = (u_*/\varkappa) \ln z/z_0, \quad (12)$$

to which Eqs. (1) and (3) are reduced if in the right-hand side of Eq. (4) only the first (linear) component remains.

If three terms remain in the expansion (10), the formulas (1) and (3) take the form:[1)]

$$q(z) = q_1 + q_* \left(\ln(z/z_1) + \frac{z - z_1}{2L_*} + \frac{z^2 - z_1^2}{6L_*^2} \right); \quad (13)$$

$$c(z) = \frac{u_*}{\varkappa} \left(\ln(z/z_0) + \frac{z - z_0}{2L_*} + \frac{z^2 - z_0^2}{6L_*^2} \right). \quad (14)$$

Analysis of the observational data has shown that for describing the wind velocity profile at stratification, essentially different from the equilibrium one (when $|L_*| \sim 5 \div 50$ m), in the right-hand sides of Eqs. (13) and (14) we should take into account not only the first and second terms (in this case the formula is called linear-logarithmic), but the third (square-law) term as well.

The profiles $q(z)$, $c(z)$, and $T(z)$ well (within the error less than 5%) agree with the logarithmic ones up to the height $z_* = 0.1\, |L_*|$. With the increase of the height these profiles deviate more and more from the logarithmic ones. The wind velocity grows with the increase of z more rapidly than $\ln z$, at a stable ($B > 0$, $L_* > 0$) stratification of the boundary layer and less rapidly at unstable ($B < 0$, $L_* < 0$) one.

During the last decades numerous data of the gradient measurements of c and T in different regions of the world have been analyzed. This analysis has shown that using Eq. (14) and more general formulas (2) and (3) the experimental data are approximated with high precision. As a result, we can conclude that formula (1) describes the altitude distribution q to a high accuracy.

The results of calculations of the dimensionless ratio β by formula (1)

$$\beta = \frac{q(z) - q_1}{q_*} = -\varkappa \rho_1 u_* \frac{q(z) - q_1}{Q_q(0)}$$

are given for several values of L_*/z_1 in Table I. Under conditions of usually observed positive fluxes ($Q_q(0) > 0$), the trace species concentration decreases with the height: $q(z) < q_1$ at $z/z_1 > 1$. However, the rate of the q decrease with height differs essentially at stable ($L_* > 0$) and unstable ($L_* < 0$) stratifications of the atmosphere.

TABLE I. The ratio $10^2\beta$.

L_*/z_1	z/z_1								
	0.25	0.5	2	5	10	20	40	100	200
10	−142	−72	74	182	279	411	623	1225	2225
25	−141	−70	71	169	249	340	457	718	1120
∞	−139	−69	69	161	230	300	369	460	530
−25	−137	−68	67	153	213	264	301	322	324
−10	−135	−67	64	142	189	221	233	235	235

At $L_* > 0$ the concentration decreases with the height much rapidly than at $L_* < 0$, i.e., at $z/z_1 = 100$ the ratio β amounts to 12.25 at $L_*/z_1 = 10$ and equals only 2.39 at $L_*/z_1 = -10$.

By means of the parameters L_* and u_* the difference $q_1 - q(z)$ rather complicatedly depends on the wind velocity, the temperature lapse rate ($\gamma = -\partial T/\partial z$), and the roughness parameter (z_0).

[1)] For this purpose η should be written as $\eta = x(1 + x/2 + x^2/6)$, where $x = z/L_*$. Having used the representation $\ln(1+y) \approx y$ at small value of $y = x/2 + x^2/6$, we obtain $\ln\eta = \ln x + x/2 + x^2/6$.

Let us determine the difference $q_1-q(z)$ for the two values of c_1, assuming that all other parameters (Q_T, Q_q, z_0) are the same. Assume that under a light breeze (c'_1) the scale L'_*/z_1 is close to 10 (at $B > 0$). If the wind velocity increases up to c''_1, being more than c'_1 by a factor of 5, then L_*, according to Eqs. (6) and (7) increases by about 125 times. At the value of $L_*/z_1 \approx 1250$ the height distribution of q is described by the logarithmic formula (11).

The values of the ratio $b = (q_1-q)'/(q_1-q)''$ of the differences q_1-q at c'_1 and $c''_1 = 5c'_1$ are

z/z_1	0.25	0.5	2	5	10	25	100	200
b	5.17	5.23	5.43	5.72	6.14	7.34	13.5	21.2

With the increase of wind velocity the difference q_1-q grows more rapidly than the ratio β.

According to Eq. (6), an increase in the roughness parameter z_0 affects the values β and b in the same manner as the growth of c_1. Let us give the values of the ratio $a = (q_1- q)'/(q_1-q)''$ of the difference q_1-q, at $z'_0/z_1 = 10^{-2}$ and $z''_0/z_1 = 0.25$:

z/z_1	0.25	0.5	2	5	10	25	100	200
a	3.43	3.51	3.60	3.80	4.08	4.87	8.95	14.1

The dependence of the difference q_1-q on γ, having a pronounced effect on the heat flux $Q_T(0)$, and owing to it, on L_*/z_1, follows immediately from the data of Table I. Thus, if at a certain value of $\gamma' < \gamma_a$ (stable stratification: $Q_T(0) < 0$) the scale $L_*/z_1 = 10$ while at the other value of $\gamma'' < \gamma_a$ ($Q_T(0) > 0$) the scale $L_*/z_1 = -10$, the ratio $(q_1-q)'/(q_1-q)''$ varies from 1.08 at $z/z_1 = 0.25$ to 9.31 at $z/z_1 = 200$.

It is readily seen that all the above regularities are explained by the influence of turbulent exchange on the transfer and altitude distribution of pollutants. In the model discussed the dependence of the turbulence coefficient k_z on the altitude is described by the following formula:

$$k_z = \varkappa\, u_*\, L_*(1 - \exp(-z/L_*)). \tag{15}$$

At a stable stratification ($B>0$, $\gamma < \gamma_a$) and at a certain altitude, k_z is the larger, the greater are L_* and u_*.

On the one hand, the pollutant flux $Q_q = -\rho k_z(\partial q/\partial z)$ is constant, and, on the other hand, this value is determined by the amount of pollutants emitted from different sources near the ground surface. Hence, the increase of the turbulent exchange intensity (the increase of c_1, z_0, and γ is useful in this case; c_1 and z_0, affect k_z in two ways through u_* and L_*) is accompanied by a decrease of $-\partial q/\partial z$; the slower the pollutant concentration decreases with height, the larger are c_1, z_0, and γ.

At the unstable stratification ($a<0$, $\gamma > \gamma_a$), when $L_* < 0$, at a given height k_z also grows with increasing L_* (i.e., with the decrease of L_* absolute value). Therefore in this case the pollutant concentration height decrease is the slower, the greater is γ. However, the dependence of the distribution of q on c_1 and z_0 at $\gamma > \gamma_a$ is not single-valued, as at $\gamma < \gamma_a$. At $\gamma > \gamma_a$ ($L_* < 0$) the growth of c_1 and z_0 results via u_* in the increase of k_z, and through L_* it results in the decrease of k_z. Thus, if at $L_*/z_1 = -10$ and $c'_1 = 1$ m/s the wind velocity increases up to $c''_1 = 5$ m/s, then k_z at 10 m height grows almost 5 times under the effect of u_* and decreases by a factor of 1.72 under the effect from L_*, i.e., as a result, k_z increases by a factor of 2.9. The above-mentioned ratio $L = (q_1-q)'/(q_1-q)''$ takes the following values at ($c''_1 = 5\, c'_1$, $L'_*/z_1 = -10$):

z/z_1	0.2	0.5	2	5	10	25	100	200
b	4.86	4.82	4.65	4.40	4.11	3.52	2.55	2.22

Assume that the turbulent flux of a trace species $Q_q(0)$ has the form:

$$Q_q(0) = -\rho_1\alpha_q c_1(q_3 - q_2), \tag{16}$$

where q_3 and q_2 are the species concentrations at the levels z_3 and z_2; c_1 is the wind velocity at z_1; α_q is the gas exchange coefficient.

At vertical distribution of $q(z)$ and $c(z)$ described by the formulas (1) and (3), for α_q we derive the ratio:

$$\alpha_q = \frac{\varkappa^2}{\ln(\eta_3/\eta_2)\ln(\eta_1/\eta_0)}\ , \tag{17}$$

where η_3, η_2, η_1, and η_0 are the values of the variable η at z_3, z_2, z_1, and z_0, and $z_3 = nz_1$; $z_2 = z_1/n$.

Results of α_q calculation at $n = 2$ are given in Table II. At a constant ratio z_0/z_1 the gas exchange coefficient increases by several times, when passing from a very stable stratification, when $L_* \sim (5.0 \div 1.0)z_1$, to a very unstable one, when $L_* \sim -(5.0 \div 1.0)z_1$. The value of α_q essentially depends on the roughness parameter, namely, at a constant L_* the value of the gas exchange coefficient increases by one order of magnitude with the growth of z_0/z_1 from between 10^{-4} and 10^{-3} to 0.25–0.50.

To calculate trace species flux by formula (16), one needs in addition to the wind velocity c_1 at the level z_1, to have the data on the species concentration at two levels, z_3 and z_2. In practice the measurements of q are normally performed only at one level. In this connection, along with the calculations at $n = 2$ the value of α_q is calculated at $n = 50$ (Table III). If in this case $z_1 = 10$ m, then $z_3 = 500$ m and $z_2 = 0.2$ m. The estimates show that at 500 m height the species oncentration q_3 is by about one order of magnitude less than its value q_2 close to the earth's surface. By this it is meant that the calculation of $Q_q(0)$ can be made with an error less than 10% only using the concentration q_2 measured at level z_2:

$$Q_q(0) = \rho_1 \alpha_q c_1 q_2, \tag{18}$$

where c_1 is the wind velocity at 10 m level.

TABLE II. The gas exchange coefficient $10^4\ \alpha_q$ (n = 2: $z_3 = 2z_1$; $z_2 = 0.5z_1$).

z_1/L_*	z_0/z_1							
	10^{-4}	10^{-3}	10^{-2}	$5\cdot10^{-2}$	0.1	0.15	0.25	0.50
1.0	64.7	84.7	123	179	226	267	351	648
0.8	71.9	94.5	137	203	257	310	405	811
0.6	80.1	106	154	232	293	358	468	893
0.4	89.6	119	172	264	340	410	545	1057
0.2	101	134	197	304	389	470	640	1258
0.1	107	142	204	319	409	500	692	1301
0.05	109	146	218	337	436	529	708	1437
0.025	110	149	223	343	446	542	739	1475
0	113	151	226	348	452	549	751	1503
−0.025	115	153	230	354	460	559	766	1536
−0.05	116	155	234	360	469	570	782	1571
−0.1	120	160	241	374	486	593	815	1646
−0.2	127	171	258	402	525	641	885	1804
−0.4	144	193	294	465	619	749	1046	2172
−0.6	163	220	337	540	711	879	1239	2628
−0.8	185	250	384	628	830	1032	1470	3183
−1.0	211	286	444	720	974	1218	1756	3889

The data given in Tables II and III show that the trace gas flux under the influence of meteorological conditions (thermal stratification, wind velocity) and roughness of the earth's surface may vary by several tens or even hundreds times.

TABLE III. The gas exchange coefficient $10^4\ \alpha_q$ (n = 50: $z_3 = 50z_1$; $z_2 = 0.02z_1$).

z_1/L_*	z_0/z_1							
	10^{-4}	10^{-3}	10^{-2}	$5\cdot10^{-2}$	0.1	0.15	0.25	0.50
1.0	2.75	3.59	5.22	7.59	9.55	11.3	14.9	27.5
0.8	3.40	4.46	6.47	9.59	12.1	14.6	19.1	38.3
0.6	4.40	5.83	8.46	12.8	16.1	19.2	25.7	49.1
0.4	6.17	8.20	11.8	18.2	23.4	28.3	37.6	72.8
0.2	10.0	13.3	19.6	30.2	38.2	46.7	63.6	125
0.1	14.0	18.5	26.6	41.6	53.4	65.2	83.8	170
0.05	16.6	22.3	33.3	51.5	66.6	80.8	108	220
0.025	18.3	24.6	36.8	56.6	73.6	89.4	122	243
0	20.0	26.7	40.0	61.6	80.0	97.2	133	266
−0.025	21.7	28.8	43.3	56.7	86.7	105	144	289
−0.05	22.9	30.6	46.3	71.2	92.7	113	155	311
−0.1	25.4	33.8	50.9	79.1	103	125	172	348
−0.2	28.6	38.5	58.1	90.5	118	144	199	406
−0.4	33.1	44.4	67.6	92.5	142	172	241	500
−0.6	36.5	49.3	75.5	121	159	197	277	588
−0.8	39.5	53.4	81.9	134	177	220	314	679
−1.0	42.2	57.4	89.1	144	196	244	352	781

REFERENCES

1. L.T. Matveev and S.A. Soldatenko, Atmos. Oceanic Opt. **7**, No. 2, 108–111 (1994).