RESULTS OF INVESTIGATIONS OF THE ATMOSPHERIC WATER VAPOR COLUMN DENSITY USING AN OPTICAL HYGROMETRY METHOD. PART II. CHARACTERISTICS OF THE WATER VAPOR COLUMN DENSITY VARIABILITY

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The paper describes the characteristics of temporal variability of the water vapor column density of the atmosphere for several seasons of 1992 to 1994 in the city of Tomsk. The amplitudes of the short-term variations of the water vapor column density are shown to be close to the magnitude of its seasonal variations. The regular component of diurnal run weakly manifests itself and is minimal at noon. The correlation between water vapor column density and surface humidity is analyzed. The height of the atmosphere homogeneous in humidity is shown to have a meaning of its reciprocal altitude profile. Monthly mean and seasonal values of the water vapor column density agree well with the data of long-term aerological observations in the West Siberia.

1. INTRODUCTION

In the first part of this article[1] we have analyzed the use of the spectroscopic method of solar photometry with different calibrations. The results obtained have made it possible to consider the optical method to be an effective instrument for investigating sufficiently small variations of the atmospheric water vapor column density $W(\sigma_W < 0.07\ \mathrm{g/cm^2})$.

It is well known[2–5] that the seasonal variations of W are clearly defined and characterized for mid-latitudes by no less than a three-fold change of the water vapor column density from winter to summer. Although the data on variations of W with periods less than a seasonal one are few in number,[5–7] we can conclude that there exists a synoptic variability with the periods of 10–11 and 3 days as well as the ambiguity in a manifestation of variations with smaller periods.

The present paper describes the results of investigations of variations of the water vapor column density of the atmosphere for several seasons in 1992–1994 in the vicinity of Tomsk. Some results were obtained in the forest area, at the scientific base of IAO SB RAS 70 km apart from the city. The measurements were made practically every day with the periodicity of 30 to 60 minutes when the sun was not screened with clouds. To improve the reliability of the data analyzed, the results of single measurements (from two to ten measurements per hour) were beforehand averaged over the one hour periods. The hourly averaged values W_h were used for estimating the character of diurnal variability of the water vapor column density, and slower variations were analyzed using the diurnal mean values $W_\mathrm{d} = \sum_{i=1}^{n} W_\mathrm{h}/n$, where $n = 5$–10 is the number of hourly averaged values of the water vapor column density per day.

2. STATISTICS OF W_d FOR DIFFERENT SEASONS OF 1992–1994

General character of variation of the water vapor column density W_d for all periods of measurements is shown in Fig. 1. For better estimate of seasonal transitions we smoothed out the initial data on W_d, i.e., calculated the average running values of the water vapor column density with the averaging period of 5 days. The data presented show the variations of synoptic scale and seasonal variations when going from spring to summer. The differences in the time of transition to higher summer water vapor column density turned out to be equal to about 19 days (from June 25 in 1992 to June 6 in 1993).

Analysis of the results of W measurements in the forest area (July 11 to July 26, 1992, and July 20 to July 28, 1994) enabled us to draw the conclusion about the lack of differences with the city area in both the average values of the water vapor column density and in the character of its variation. That is, under plain conditions and when there are no big water bodies close to the measurement site, local peculiarities proved to have insignificant influence on W.

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Table I gives the report of statistical parameters for different months of observations and selected seasons, including average, minimal, and maximal values of W, the standard deviations σ_W, the variation coefficients V_W, and the number of measurement days N. As would be expected, the amplitude of seasonal variations shows the multiple change of W. Thus, on the basis of monthly mean data of 1992, the water vapor column density in July was 5.7 times as large as in December. Similar values of W for the Tomsk area are presented in Ref. 4: 2.7 in July and 0.5 in December; a 5.4-fold decrease is observed. The statistical characteristics obtained (Table I) are in good agreement with the many-year observations of the water vapor column density at the nearest station of aerological sounding located in Novosibirsk.[2]

From the tabulated data the conclusion follows that along with the seasonal increase of W from spring to summer the amplitude of short-period variability also increases, and the variation coefficient $V_W = \sigma_W / \overline{W}$ remains within the limits of 35–45%. It should be noted that the maximal amplitudes of day-to-day variations of the water vapor column density reach almost the value of seasonal variations.

Figure 2 shows the repetition histograms for W_d. In the spring period, the histogram is of unimodular character with the maximum close to the mean value for the season. The histograms for summer periods are more diffuse, i.e., their view reflects the fact that in the wide range of values $1.5 < W_d < 4$ the water vapor column density occurs with the same repetition frequency.

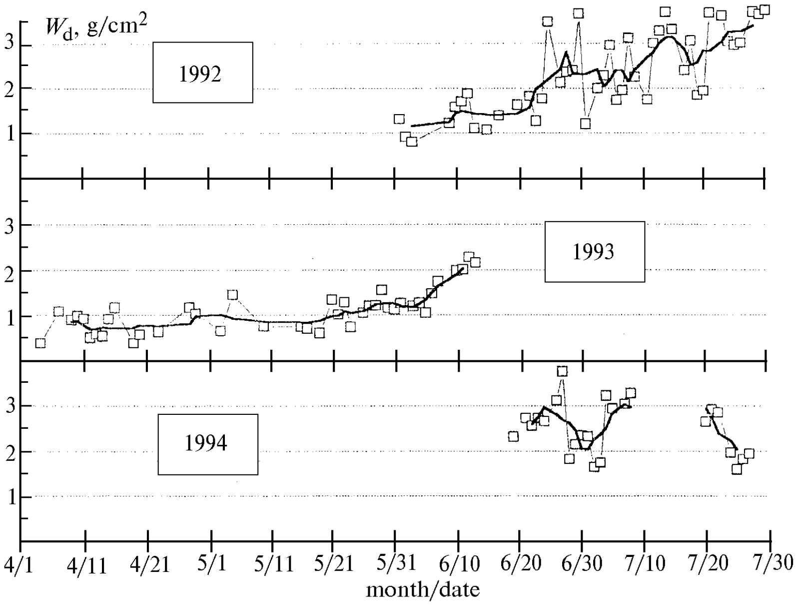

FIG. 1. Time behavior of the water vapor column density of the atmosphere W_d during different periods of investigations (the mean running values of W are shown by a heavy line).

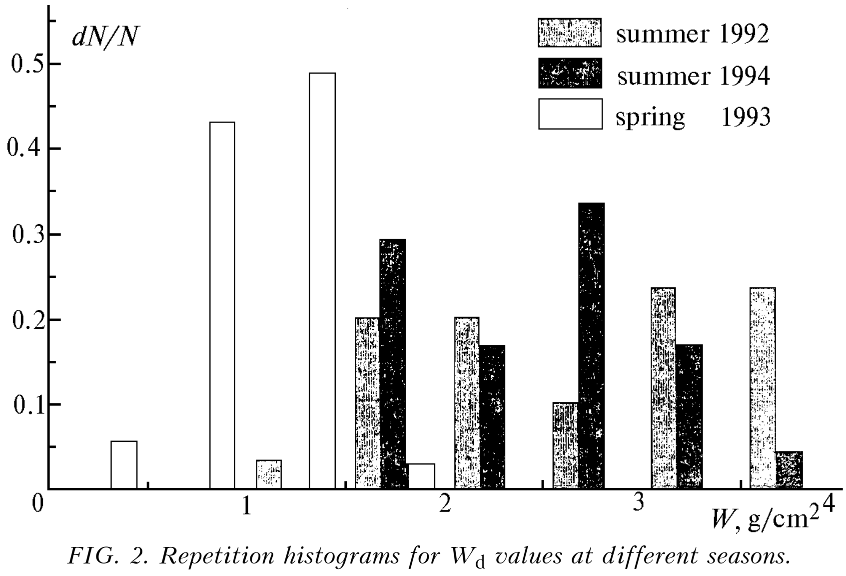

FIG. 2. Repetition histograms for W_d values at different seasons.

TABLE I. Statistical parameters of W_d variations for different months and seasons.

Month, season	Year, region	W_{mean}	Min	Max	σ	V	N
April	Tomsk, 1993	0.783	0.378	1.168	0.281	0.359	15
	Novosibirsk	0.775	0.142	1.770	0.358	0.461	–
May	Tomsk, 1993	1.036	0.604	1.552	0.302	0.292	16
June	Tomsk, 1992	1.751	0.790	3.655	0.783	0.447	19
	Tomsk, 1993	1.636	1.048	2.267	0.443	0.271	10
	Tomsk, 1994	2.618	1.818	3.763	0.540	0.206	10
July	Tomsk, 1992	2.760	1.191	3.737	0.767	0.278	25
	Tomsk, 1994	2.424	1.539	3.282	0.628	0.259	14
	Novosibirsk	2.859	0.582	3.851	0.708	0.248	–
December	Tomsk, 1992	0.482	0.264	0.710	0.164	0.340	8
January	Novosibirsk	0.384	0.063	1.013	0.203	0.529	–
Spring	Tomsk 92, 93	1.069	0.378	1.869	0.374	0.350	49
Summer	Tomsk 92–94	2.578	0.692	3.767	1.191	0.462	60

3. DIURNAL VARIABILITY OF WATER VAPOR COLUMN DENSITY

The character of diurnal variations of W_h is characterized by a great variety, namely, the increase and decay of W by noon and evening, approximately constant level during daytime, and so on.

The repetition histogram (Fig. 3) of maximal diurnal relative changes of water vapor column density $\varepsilon_W = (W_{max} - W_{min})/W_d$ shows that in the summer period no less than 70 per cent of diurnal variations are within the range of ε_W from 0 to 0.3, and the maximal value of ε_W is 0.7. In spring the amplitudes of diurnal variations of W are distributed in the wider range with the maximal $\varepsilon_W = 1$. The most probable value of $\varepsilon_W = 0.15$ is the same for the two seasons. In absolute values, the maximal probability corresponds to the amplitude of diurnal variations 0.16 g/cm^2 in spring and 0.35 g/cm^2 in summer.

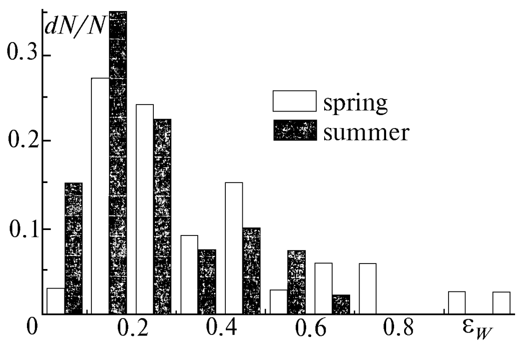

FIG. 3. Repetition histograms for relative diurnal variability ε_W of the water vapor column density.

The wide range of diurnal variations of W is the result of manifestation (or overlapping) of larger–scales variations due to synoptic processes. For revealing the proper diurnal variation of the water vapor column density, we calculated the mean values and variances of W_h for every hour of observations. In this case it was assumed that the averaging enables us to minimize the synoptic scale effect. Figure 4 presents the normalized values of (W_h/W_d) indicating, in the diurnal variation, a slight minimum of the water vapor column density with relative depth about 10 per cent. The difference in the value of W in the periods from 10:00 to 13:00 and from 13:00 to 14:00 L.T. is confirmed with a large confidence coefficient α, calculated by the Student t-test. Meridian minimum of the water vapor column density was also mentioned in Refs. 5 and 12 for the stable anticyclone situations in Havana as well as for specific atmospheric conditions of Issyk Kul basin. The presence of minimum in the W diurnal run in our experiments is yet to be explained. One can assume that during measuring periods the atmospheric situations with a developed convection prevailed, resulting in the water vapor condensation when it rises and the moisture outflow to cloudiness. (For example, the formation of only one cloud with water storage of 1.5 kg/m^2 results in the decrease of water vapor column density of about 7 per cent).

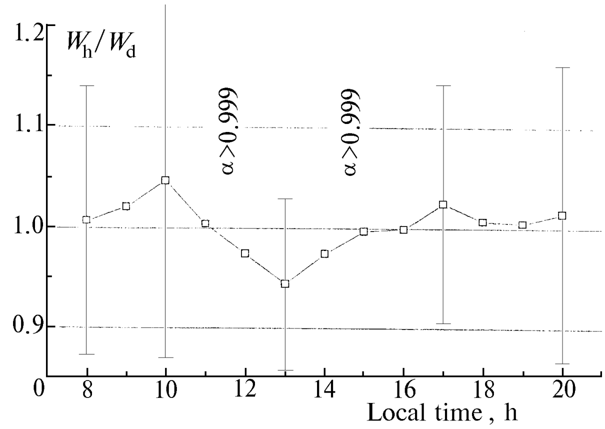

FIG. 4. Average diurnal variation of normalized values of water vapor column density during spring–summer observation periods.

4. CORRELATION BETWEEN THE INTEGRAL AND SURFACE HUMIDITY CHARACTERISTICS

The correlation between the water vapor column density of the atmosphere and the surface moisture was analyzed by many authors.[2,9,10] According to these investigations, the correlation is close to a linear one, and the coefficient of cross-correlation is within the limits from 0.7 to 0.98. Account must be taken of the difficulty encountered in comparing the results of correlation because of the use of different characteristics of moisture and different scales of averaging.

When evaluating we took the hourly mean values W_h and water vapor pressure e_0. Figure 5 shows the dependence obtained for all observation periods. The correlation coefficient is found to be equal to 0.878, and the regression equation is written in the form:

$$W = (0.101\pm0.003)e_0 + (0.420\pm0.034). \tag{1}$$

In this figure for comparison the approximation dependence, obtained in Ref. 10,

$$W = 0.15e_0 \,, \tag{1a}$$

is given.

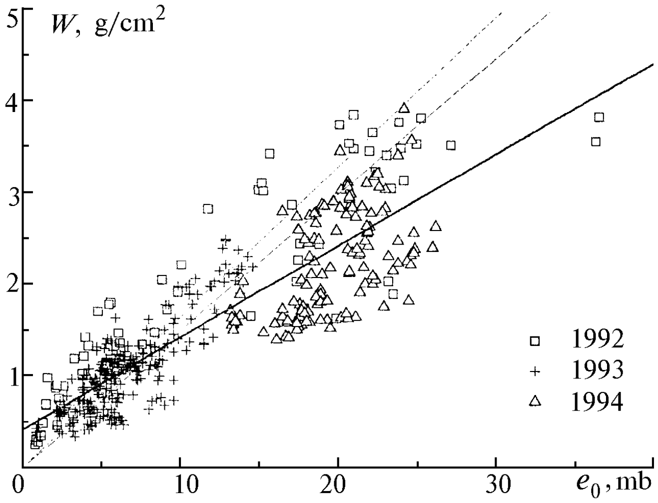

FIG. 5. Water vapor column density W_h vs. the water vapor pressure in the surface atmospheric layer: the approximation dependence[10] (dashed line), the approximation by Eq. (8) with the use of many-year data[4–8] (dot-and-dash line).

Although the regression coefficients in the equations cited are different, it should be noted that in the basic part of the range (e_0 = 5–25 mb) the both approximations give the results being in close agreement. A comparison of the results for separate years indicates that the data of 1994 are peculiar, because the common values e_0 correspond to lesser column density. It is believed that the above disagreement was due either to peculiar features of meteorological conditions over a short period of measurements in 1994 (24 days) or to a systematic error of 10–20% of calibration in 1994 for low column density (interpolation error in the range of $W > 1.5$ g/cm^2 in Ref. 1). The lack of radiosonde observations in 1994 has made it impossible so far to draw a definite conclusion.

A knowledge of the water vapor column density W and humidity in the surface layer enables one to estimate basically its altitude distribution. For this purpose we use the concept of the height of homogeneous atmosphere for the humidity H_0, which is introduced in the form[11]:

$$H_0 = W/a_0, \tag{2}$$

and determines the equivalent thickness of the atmosphere with constant, in height, absolute humidity, being equal to the surface one. The altitude profile of humidity in the troposphere is well described, on the average, by the exponential dependence.[3] (In this case, it should be remembered that it is precisely the smoothed averaged profile without local inhomogeneities or inversions.) Using, for example, the known Hann formula[10] for absolute humidity one can write:

$$a(h) \approx \frac{217}{T(h)}\, e_0 \cdot 10^{-h/\beta'} = \frac{217}{T(h)}\, e_0 \exp(-\beta h), \tag{3}$$

where e_0 is the water vapor pressure at the ground; β is the index of altitude humidity variation. The altitude variation of temperature in the troposphere is generally described by the linear function as $T(h) = T_0(1 - \alpha' h)$. Taking into account a slight influence of $T(h)$ on $a(h)$, for convenience of further consideration, this dependence can be represented as an exponential function with the exponent α:

$$T(h) \approx T_0 \exp(-\alpha h). \quad (4)$$

Substituting Eq. (4) into Eq. (3) gives

$$a(h) = a_0 \exp[(\alpha - \beta)h]. \quad (5)$$

Having integrated Eq. (5) over height, we obtain the expression for water vapor column density of the atmosphere:

$$W(H) = [a_0/(\beta - \alpha)][1 - e^{(\alpha-\beta)H}]. \quad (6)$$

When H tends to infinity and, actually, already from the altitude of several kilometers, the exponential function in Eq. (6) vanishes, and instead of Eq. (2) we can finally write:

$$H_0 \approx (\beta - \alpha)^{-1}, \quad (7)$$

in this case $\beta \gg \alpha$.

Thus, the height of homogeneous atmosphere (2) simultaneously has a meaning of the inverse value of the index of altitude variation of humidity and its profile can be estimated by the measured values of W/a_0. It should be noted that the use of more accurate but cumbersome expressions for $a(h)$ does not change the essence of the connection considered, especially as the variety of real situations in the atmosphere prevents the description of $a(h)$ as single universal dependence.

Note that the considered connection of the integral parameter – water vapor column density – with the index β is not a surprising peculiarity of the humidity characteristics, but the result of the property of the exponential function. As applied to the atmospheric parameters, the manifestation of this property can be briefly formulated as follows. If the vertical distribution of the atmospheric parameter (humidity, extinction coefficient, etc.) can be expressed as the exponential function, then accurate to this representation we can consider that the index of the vertical profile β (or the height H_0^{-1}) is equal to the ratio of the parameter value in the surface layer to its altitude integral.

Our study reveals that the coefficient of the regressive equation in the version (1*a*) can also be expressed in terms of H_0 or $(\beta - \alpha)$:

$$k = W/e_0 = 217 H_0/T_0 = a_0/e_0(\beta - \alpha). \quad (8)$$

The height H_0 calculated from the results of local meteorological observations of e_0 and W for the southern areas of the Tomsk region[4,8] can be estimated within the range from 2 km in warm period and to 3.8 km in winter.

TABLE II.

Period	$H_{0\text{mean}}$	$H_{0\text{min}}$	$H_{0\text{max}}$	σ_{n}	V_{n}	N
Spring, 1992, 1993	2.55	1.21	5.06	0.83	0.32	44
Summer 1992–1994	1.93	1.20	3.06	0.43	0.22	41
December 1992	3.47	2.28	4.66	0.81	0.23	7

From the characteristics of H_0 variability in our investigations (Table II) it follows that in the spring–summer period of 1992–1994 the mean value of the height of homogeneous atmosphere was about 2.2 km and varied moderately (V = 20–30 per cent). Using the cited value of H_0, one can estimate the mean coefficient of linear regression $k = 0.165$ by Eq. (8). The corresponding linear approximation is given in Fig. 5 by a dot-and-dash line and, on the average, it is in the good agreement with the results of joint measurements of W and e_0. A more detailed analysis of variability of the height H_0 and its use for estimating the humidity altitude profile is beyond the scope of this paper.

Thus, the above investigations lent support to the validity of the method of optical hygrometry and enabled us to estimate the characteristics of short-period variations of the water vapor column density of the atmosphere and to determine the character of correlation between the parameters H_0, $(\beta - \alpha)$, and k.

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