

WATER VAPOR IN THE STRATOSPHERE

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1. INTRODUCTION

THE need for solving several practical problems and explaining theoretically the physical processes that occur in the upper layers of the atmosphere has stimulated during the last decade intense research on the stratosphere. The availability of new methods and means of research, particularly balloons, high-altitude airplanes, rockets, and artificial earth satellites, which have made it possible to lift the measuring apparatus to high altitudes, has greatly contributed to the acquisition of new factual data concerning the structure of the stratosphere, its thermal and dynamic states, its optical properties, and other important characteristics of this atmospheric layer.

During the initial period of these investigations, principal attention was paid (for perfectly understandable reasons) to measurements of the main characteristics of the stratosphere such as chemical composition, temperature, pressure, density and wind velocity at high altitudes. In most cases the data obtained have radically changed the previously prevailing notions concerning these characteristics and concerning the physical processes occurring in the stratosphere. At the same time, the obtained data have in turn served as the starting point for the formulation of new problems.

A clear example illustrating this premise is the discovery that the stratosphere has a complex thermal structure with many peculiarities. It has been established relatively long ago that in the lower stratosphere (10–20 km) where the temperature reaches 200–220°K, isothermy gives way to a temperature rise in the lower layer of the mesosphere (20–50 km) with maximum values on the order of 270–300°K at levels 45–50 km (mesopeak); above the mesopeak a decrease in the temperature is observed, with a minimum on the order of 180–200°K at the upper boundary of the mesosphere, at an altitude of 75–80 km (mesopause)*. However, no satis-

factory physical explanation for these peculiarities in the vertical profile of the temperature has been found as yet. For example, the temperature rise in the mesosphere and the existence of the mesopeak are attributed to the absorption of ultraviolet radiation from the sun in the ozone layer which lies between 15 and 50 km. Yet precisely the same temperature variation is observed in the mesosphere under conditions of the polar night, when this source of heat is completely absent. Nor has the origin of the temperature minimum in the mesopause region been determined. As is well known, along with the low temperatures one observes in this region also the formation of silver clouds, concerning the nature of which there are at the present two different opinions (the water and dust hypothesis of the origin of the silver clouds). The formation of silver clouds is apparently directly connected with the temperature minimum.

Inasmuch as water vapor is one of the main factors determining the heat conditions in the stratosphere, it is quite obvious that information on the vertical distribution of water vapor at high altitudes is necessary in order to understand the thermal structure of the stratosphere. At the same time, the data on the moisture content in the stratosphere can also help determine the origin of silver clouds.

On the other hand, these data should answer the question of whether there is enough moisture in the stratosphere to ensure that the silver clouds are due to the presence of water. These data should also explain whether a connection exists between the temperature minimum and the vertical distribution of the water vapor in the stratosphere. The recently published result of measurements of the fluxes of long-wave (thermal) radiation of the atmosphere at 20–25 km altitude offers evidence that, in spite of the hitherto prevailing notions, water vapor can play an important role in the thermal conditions of the stratosphere, at least in the 30–50 km layer.

Data on the distribution of water vapor in the atmosphere, while serving as an indicator of the air

*The mesosphere layer is frequently also called in the literature the "upper stratosphere."

masses, can also render appreciable help in the investigation of the general circulation of the atmosphere, particularly in the study of such thermodynamic phenomena as the frequently observed radical variations of the temperature of the stratosphere, amounting to tens of degrees within short time intervals.

In addition to the foregoing, there are other problems in atmospheric physics (for example, the question of the source of atomic hydrogen in the ionization layer), and also purely practical problems, connected with high-altitude flight of airplanes, rockets, and artificial earth satellites; to solve these problems it is necessary to have data on the distribution of water vapor in the upper layers of the atmosphere. This is apparently the reason why many researchers have paid attention to the question of the content of water vapor in the stratosphere during the last decade. The data contained in the foreign literature offer evidence that various methods have been employed in these researches, thus ensuring sufficient reliability of the results obtained.

As is well known, the vertical distribution of the water vapor is measured in a network of meteorological stations with the aid of standard apparatus, in the course of the aerological sounding of the atmosphere at altitudes up to 20–25 km. In particular, such measurements were carried out regularly in different regions of the earth during the International Geophysical Year (IGY). The moisture characteristic directly measured by radiosonde flights is the relative humidity u , that is, the ratio of the water vapor pressure to the vapor pressure in the saturated state. At the same time, measurements are made of the temperature and pressure of the air, so that it is possible to determine with the aid of psychrometric tables and the equation of state of an ideal gas the partial pressure (vapor tension) e , the density (absolute humidity) ρ_w , and the relative concentration of water vapor (specific humidity) $q = \rho_w/\rho$ (where ρ is the air density). In many cases one measures directly the dew point (or the frost point), that is, the temperature at which vapor of given tension becomes saturated over a surface of water (or ice).

A study of the radiosonde materials, including the IGY material stored in the World Center for IGY Data in Moscow, has confirmed the well known fact that relative humidity data obtained with standard apparatus are quite dubious for those altitudes where the temperature of air is low, and apparently do not reflect the real variation of the humidity.

As can be seen from Fig. 1, which shows for the sake of illustration the result of individual measurements in several points of the globe, the values obtained for the relative humidity in the stratosphere are quite high, and correspond to tremendous specific and absolute humidities. In all probability, owing to the unreliability of these data in some cases (particularly, in many American materials on stan-

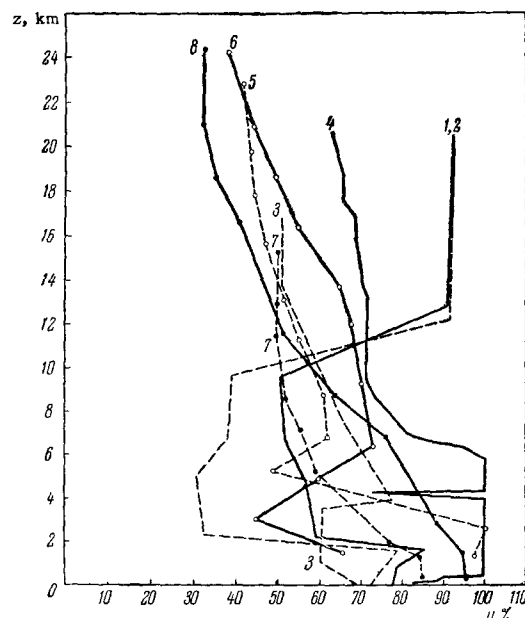


FIG. 1. Vertical distribution of relative humidity, from IGY data. 1—January 22, 1958; 2—August 18, 1958, Ceylon (Colombo); 3—August 1, 1956; 4—February 12, 1956, Mirnyĭ; 5—o—o—December 13, 1958; 6—o—July 17, 1958, Dolgoprudnaya; 7—•—•—December 18, 1958; 8—•—July 5, 1958, Heiss Island.

dard measurements during the IGY period), no data are given on the relative humidity above the tropopause. The reason for such unreliability lies in the large errors inherent in hair and electrolytic hygrometers used in standard humidity measurements at high altitudes, where very low temperatures are observed (on the order of 200–220° K) and the concentration of the water vapor is very low (reaching 10^{-6} g per gram of air). As shown by V. D. Reshetov^[1], the time constant of a hair hygrometer increases strongly and its sensitivity decreases when the air temperature and water-vapor pressure decrease. The theory proposed in^[1] for introducing corrections to the readings of hair hygrometers at low temperatures has been applied only in airplane measurements of humidity in the lower layer of the troposphere (to 5 km).

The vertical distribution of water vapor in the lower troposphere has been investigated more thoroughly. In the mean, the concentration of water vapor decreases with altitude, but this decrease is not monotonic. Quite frequently layers are encountered in the troposphere in which the humidity increases and can reach saturation. These humidity inversions are closely related with the temperature stratification of the atmosphere, and also with its dynamic state. The vertical currents and advective transport of air contribute greatly to the creation of a stratified structure in the distribution of humidity with altitude. The formation of clouds in the atmosphere is a direct manifestation of this stratification.

Owing to the vertical mixing of the atmosphere, the layers are destroyed and simultaneously some average stratification of humidity is produced in the troposphere. This average humidity distribution is usually described by the Gunn formula (see, for example [2]), which has the following form for water vapor tension:

$$e = e_0 10^{-a_0 z}, \quad (1)$$

where e_0 is the water vapor pressure at the earth's surface and depends greatly on the point of observation, $a_0 = 0.159 \text{ km}^{-1}$ is a parameter, and z is the altitude in kilometers. If we assume that the air pressure p decreases exponentially with altitude (the barometric formula), then Gunn's formula for the specific humidity q is written in analogous fashion:

$$q = q_0 e^{-\alpha_0 z} \quad (2)$$

(q_0 is the specific humidity at the earth's surface, $\alpha_0 = 0.23 \text{ km}^{-1}$).

Formula (2) is used frequently in calculations of the specific humidity at higher altitudes, where it yields values of q that are highly exaggerated. Therefore the value of the parameter α_0 is usually chosen somewhat larger than indicated above.

A. Kh. Khrgian [2] has obtained, on the basis of numerous airplane measurements of the humidity at altitudes up to 4–5 km, a second empirical relationship:

$$q = q_0 e^{-(bz + cz^2)}, \quad (3)$$

where it was found that the parameters b and c vary with the season. This formula cannot be extrapolated to high altitudes.

Owing to the aforementioned unreliability of the standard humidity measurements at high altitudes, the aerological measurement data cannot be used to obtain information concerning the vertical distribution of water vapor above 5 km.

In recent high-altitude measurements of humidity, specially constructed hygrometers have been used, capable of determining the dew point (or the frost point) with sufficiently high accuracy. This apparatus can measure very small amounts of water vapor at altitudes up to 30 km. In addition, presently developed spectroscopic methods of measuring humidity have also made it possible to obtain some information concerning the water-vapor content in the higher layers of the atmosphere. In the present article we review and analyze the results of these researches on the vertical distribution of water vapor in the stratosphere, and compare these results with data obtained from some indirect estimates of the moisture contents at high altitudes.

2. LOCAL MEASUREMENTS OF HUMIDITY

The first airplane measurements of humidity at altitudes up to 15 km were made in England by Dobson

Brewer, and Quilong [44]. The humidity was measured with Dobson's instrument, based on determination of the dew point or the frost point. The measurements were based on the principle that the temperature of the bottom of the chamber through which the investigated air passes rises or drops until equilibrium is established between the water vapor and the condensate.

The main measuring element of the instrument is a small aluminum chamber, comprising a dish with a polished, anodized, and blackened internal surface. The bottom of the dish can be cooled from the outside with liquid nitrogen or heated electrically. When the temperature of the bottom drops to the saturation temperature, ice crystals grow on it. The bottom of the chamber is illuminated and as little as 10^{-6} g of ice on its surface can be observed with a magnifier. The temperature at which the amount of ice remains constant is taken to be the frost point.

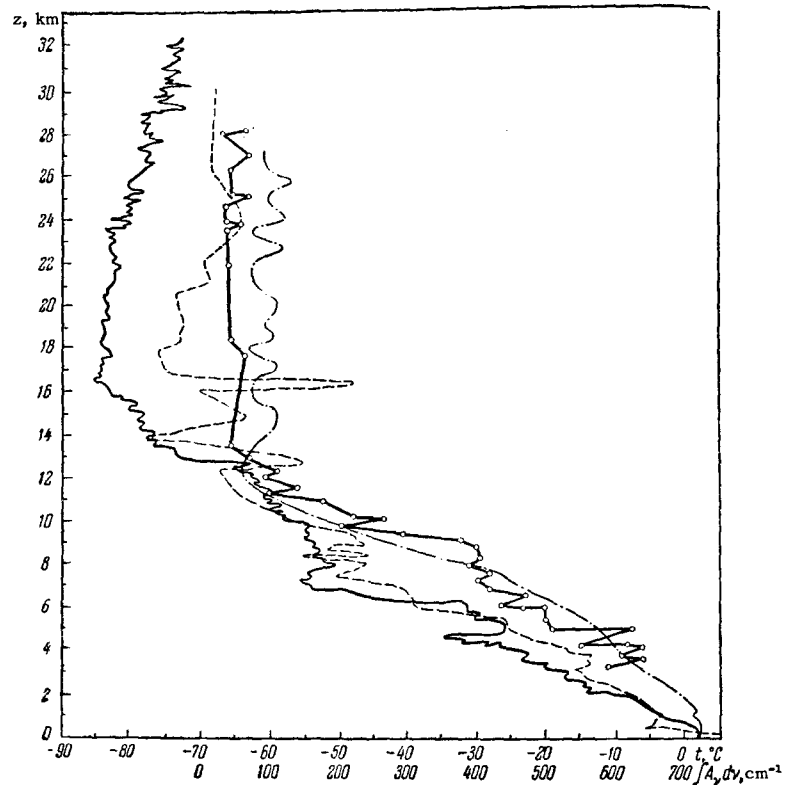
The errors of the method depend little on the measured humidity and are determined essentially by the error in the determination of the frost point. It can be shown that if the equilibrium temperature T is determined with an error $\pm 2^\circ \text{ K}$ the relative error in the measurement of humidity at $T = 200^\circ \text{ K}$ is $\sim 25\%$. This error is very small, since the measured water vapor tension can vary by a factor 10^4 . Dobson's instrument is suitable for measurements by this procedure at temperatures not lower than 180° K , at which temperature the ice settles at the bottom of the dish in the form of an invisible transparent layer, and not in the form of individual crystals.

Airplane measurements of the frost point [3, 4] have shown that there is a very little water vapor in the lower stratosphere. The frost point at 15 km fluctuates, in accordance with these data, between -80 and -85° C , corresponding to a specific humidity on the order of 10^{-3} g/kg (0.002 – 0.005 g/kg) and a relative humidity of 2–5%. In those cases when the measurements could be carried out at several levels above the tropopause (for example, in the measurements of Murgatroyd, Goldsmith, and Hollings [41]), the specific humidity had a tendency to remain constant with altitude.

The use of automatic hygrometers to measure the frost point has made it possible to raise the sounding ceiling to 30–35 km and to trace the vertical distribution of water vapor up to these altitudes. The automatic hygrometer used for these measurements is an improved Dobson instrument; the description of the automatic hygrometer will be given below.

Barret, Herndon, and Carter [5], who were the first to measure humidity with the aid of balloon-borne automatic hygrometers over the USA, obtained for the frost point in the lower layers of the stratosphere values 10 – 20° higher than obtained by airplane soundings over England. From the analysis of three vertical water-vapor distribution curves given in [5] for two summertime and one wintertime flight, it follows,

FIG. 2. Variations of the frost point, air temperature, and the absorption function. — Frost point [4], June 27, 1960; --- frost point [5], July 1, 1949; - - - - air temperature [7]; ○ — absorption function $\int A_{\nu} d\nu$ from spectroscopic data [17], June 19, 1959, 6.3μ band.



first of all, that in the mean the frost point in the stratosphere changes little with altitude. However, as shown in Fig. 2, appreciable fluctuations in the frost point were noted in flight, thus evidencing the existence of both relatively humid layers and very dry air. These fluctuations exceed by many times the possible measurement errors and show apparently the real distribution of moisture with altitude.

The question of the origin of the fine structure of the vertical distribution of water vapor in the stratosphere is of independent interest. However, the lack of the necessary material, such as would be obtained by all-inclusive observations of the various meteorological elements, does not permit at present any rigorous analysis of the stratified structure of the stratosphere. We note merely that measurements of the air temperature at these altitudes, made with a low-inertia acoustic thermometer [7], have disclosed analogous altitude variations of the temperature. It can be supposed that the thermal stratification of the stratosphere and the stratification in the water vapor distribution are interrelated.

If the frost point is recalculated in terms of specific humidity, which characterizes the concentration of water vapor at a given level*, the peculiarities of the

vertical distribution of the moisture manifest themselves most pronouncedly. As shown by these first measurements of water vapor at high altitudes, the specific humidity varies little (in the mean) with altitude. Near the lower stratosphere (up to 15 km) it turns out to be one order of magnitude higher than the specific humidity as measured in [3,4] (0.01–0.05 g/kg). With increasing altitude there was observed in the summertime an increase in the specific humidity, which reached 0.1–0.15 g/kg at levels on the order of 30 km (Fig. 3).

In wintertime measurements of the vertical distribution of the moisture [5], the specific humidity also turned out to be one order of magnitude higher than in [3,4], but in this case a weak decrease with altitude was observed, up to 25 km; above this level, the specific humidity could be regarded as constant (0.03 g/kg). We note that the temperature of the air did not increase in this case, as is usual, with altitude and turned out to be quite low.

Even higher values of water-vapor concentration were obtained at these levels by Bracefield [6] with the aid of a hygrometer (the so-called hygristor).

The main element of the instrument is an electrolyte film (saturated solution of LiCl) on a glass base. The film and a thermistor, which measures the temperature of the film, are placed between two metal electrodes. Let us assume that the vapor pressure over the film of the electrolyte is smaller than in the surrounding medium. If we apply approximately 25 V ac to the electrodes, current will flow through the film and its temperature will rise. In addition, if the vapor

*In the foreign literature the concentration is usually characterized by the so-called mixing ratio $r = m_w/m$ (m_w and m are the masses of water vapor and dry air, respectively), which under atmospheric conditions is close to q :

$$q = 0.622 \frac{e}{p - 0.378e}, \quad r = 0.622 \frac{e}{p - e}.$$

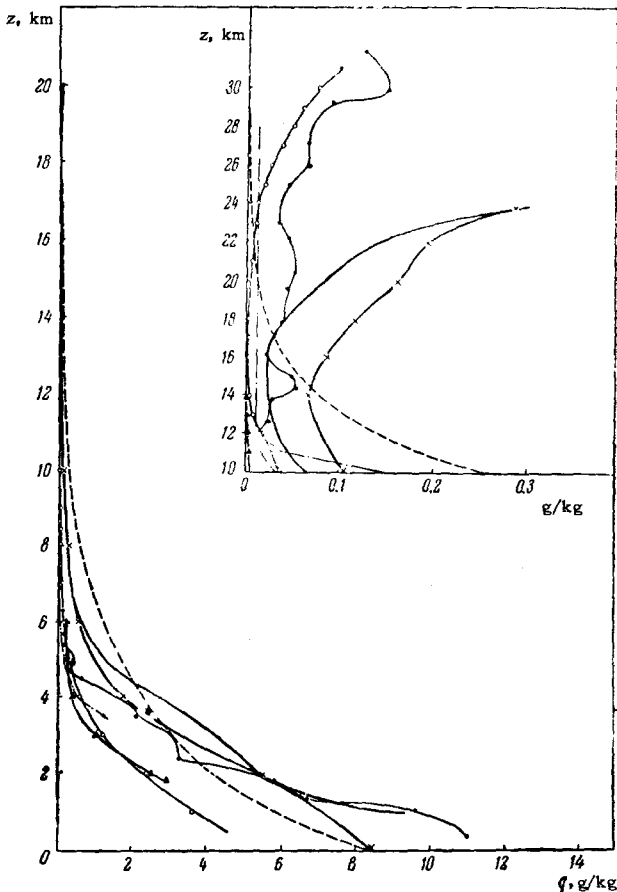


FIG. 3. Vertical distributions of the specific humidity obtained by different methods. ● - 26 August, 1949^[5]; ○ - 16 September 1953^[6]; ▲ - 6 July 1954^[4]; × - 16 July 1958 (IGY); ○ - 27 June 1960^[8]; --- exponential law (Gunn), $q_0 = 8.4$ g/kg, $\alpha = 0.23$ km⁻¹; - · - · - 19 June 1959^[17], 6.3μ band.

pressure over the film of the electrolyte is smaller than in the surrounding medium, the film will absorb water, its resistance will decrease, and the current through it will increase. This will lead to a further increase in the film temperature. The film temperature will rise until the vapor tension above the electrolyte becomes equalized with the vapor tension in the surrounding medium.

If the film temperature rises above the equilibrium temperature, water will evaporate from the film, increasing the resistance and decreasing the temperature. Measurement of the water-vapor tension by this instrument reduces to a measurement of the temperature of the electrolyte film at equilibrium.

If the film equilibrium temperature is known, the vapor tension over its surface is determined from tables.

On the calibration curve given in ^[6], the film temperature is linearly related with the dew-point temperature down to -50°C (for lower temperatures the calibration plot was extrapolated).

The use of the instrument is severely limited. First, the calibration curve does not apply when the

air velocity relative to the instrument is high. Owing to the strong ventilation, the instrument may undervalue the humidity at altitudes below 5 km. Above this level, the effect of ventilation is insignificant because of the low air density. Second, it has been observed that irreversible changes ("aging") occur in the film and lead to excessively large values of the humidity. Third, the instrument registers the dew point only if the relative humidity of the air exceeds 15%. Finally, the film temperature can only be raised; no provision is made for lowering the film temperature in the instrument. This may cause the instrument to give exaggerated humidity readings if raised into cold clouds, in view of the slow cooling. It would apparently be more advantageous to measure humidity not with a rising hygrometer, but with a descending one.

In view of these shortcomings in the procedure used to measure the water-vapor content with the aid of the hygrometer, the results of ^[6] cannot be regarded as reliable, particularly at high altitudes (above 20 km), where the values attained for the specific humidity are tremendous, greatly in excess of the data of ^[5] and close to the data obtained by standard measurements of the humidity during the IGY (the data obtained by one such measurement are shown in Fig. 3). We therefore do not present in Fig. 3 the distribution obtained in ^[6] for the specific humidity in the 25-33 km layer.

The vertical distribution of water vapor has in accordance with the data of ^[5,6] the following interesting peculiarity: Whereas in the low levels of the stratosphere there exist along with the moist layers also layers with low water-vapor concentration (0.003-0.004 g/kg), that is, with the same concentration as given by ^[3,4] in the higher layers the specific humidity is always high and its variation is considerably weaker.

The existence of high water-vapor concentration in the stratosphere, claimed in ^[5,6] and in particular the growth of relative concentration with altitude, were doubted, and with good justification, since the procedure used in these experiments made possible large errors, which were not estimated in these investigations.

Recently systematic measurements of the humidity in the stratosphere by more perfect methods have started in the USA ^[8,9,49]. The improvement consisted of introducing in the Dobson instrument an optical-electronic-thermal servomechanism to maintain the temperature of the mirror, on which the moisture is precipitated, at the saturation level (Fig. 4). The servomechanism controls the temperature of the mirror within the limits of $+50$ to -85°C , with accuracy $\pm 0.2^\circ\text{C}$. The mirror is a polished silver cylinder, 0.15 cm in diameter, coated with rhodium. A coil which serves as the heating element is wound on the mirror. The end of the cylinder is lowered into a bulb with liquid nitrogen, the pressure of which is regulated

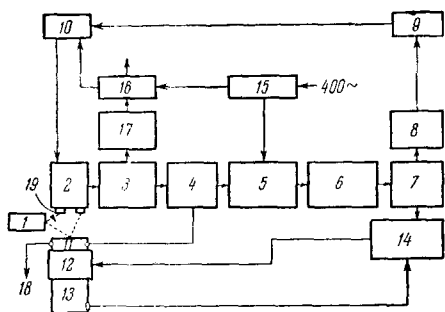


FIG. 4. Functional diagram of the servomechanism. 1—lamp; 2—balanced photocells; 3—buffer amplifier and 400 cps filter; 4—variable-gain stage; 5—demodulator; 6—corrector; 7—amplifier for mirror heater; 8—cooling switch; 9—time relay; 10—stability control; 11—mirror; 12—heating coil; 13—cooling unit; 14—heater switch; 15—reference-voltage amplifier; 16—demodulator; 17—error-signal amplifier; 18—output of mirror temperature indicator; 19—comparison photocells.

by the servomechanism. When the pressure decreases the nitrogen boils and the temperature of the rod rises.

Figure 5 shows the functional diagram of the hygrometer. Light from the lamp strikes two photocells, one after reflection from the mirror and the other directly. The amount of light reflected from the mirror depends on the thickness of the frost film. The two photocells are balanced with the aid of a bridge circuit at a fixed thickness of the film on the mirror. An increase in the film thickness turns on the heater, while a decrease opens a valve to the bulb with the liquid nitrogen. The water vapor contained in the chamber is thus constantly in equilibrium with the solid or with the liquid phase. Laboratory measurements have shown that at -50°C a single measurement of an abrupt change in humidity from zero to saturation lasts about 10 seconds. At higher temperatures the time constant decreases to 2–3 seconds. This limits the ascent and descent rates. The descent

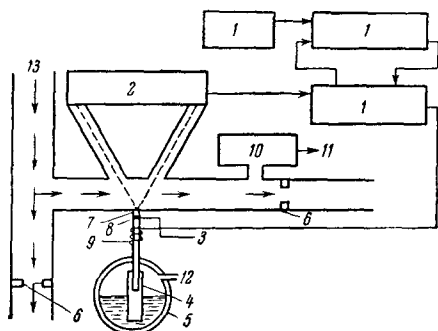


FIG. 5. Functional diagram of hygrometer. 1—servomechanism shown in Fig. 4; 2—photoelectric system; 3—output of mirror temperature indicator; 4—copper rod; 5—liquid nitrogen; 6—diaphragm; 7—mirror; 8—thermistor; 9—heating coil; 10—pressure converter; 11—output of pressure converter; 12—pressure regulator in vessel with liquid nitrogen; 13—output opening of measuring chamber.

of the instrument is slowed down either by a large parachute or by a system of two balloons, one of which becomes detached at a specified altitude.

In spite of these precautions, a difference exists between the pressures in the atmosphere and in the measuring chamber, owing to the fact that the instrument moves (this is probably characteristic of all types of chamber hygrometers). To take this error into account, the pressure inside the tube and outside the instrument is determined to calculate the correction.

To increase the accuracy, the latest models of the instrument have been provided with a ventilating unit, to completely remove at will the condensate from the measuring chamber. The need for such a device is dictated by the fact that large amounts of condensate on the mirror make it difficult for the compensation system to operate. It is possible that the absence of this unit from the earlier automatic hygrometers used by Barret et al [5] is the reason for the higher humidity values obtained in the lower stratosphere.

In determining the saturated water-vapor pressure from the equilibrium temperature it is assumed that the condensate consists of frost. Since the saturated vapor tension is different over water and over ice, this assumption can lead to an error if water condenses in place of ice. At -40°C the error can reach 4°C .

A temperature drop or spontaneous freezing of the supercooled water can be mistaken for a fluctuation in the humidity. This error is partially excluded by comparing data obtained as the instrument ascends and descends. If nearly equal vertical distributions of the frost point are obtained in the two cases, the obtained data can be regarded as close to reality. The only water-vapor distributions given in [8] are those satisfying this criterion.

As follows from Fig. 2, Mastenbrook and Dinger [8] observed frost-point variations analogous to those previously obtained by Barret et al [5]. These variations exceeded the measurement errors and, in addition, were observed during descents, thus unconditionally confirming the layered structure of the stratosphere. Mastenbrook and Dinger [8] give curves obtained for the specific humidity during one summertime and two springtime ascents of the hygrometer to 30 km (in one of the springtime cases the data obtained were only up to 24 km). In the summertime and that particular springtime case the specific humidity had a minimum (0.002 g/kg) at the 14–16 km level, in good agreement with the data of [3,4], but this minimum value was lower than in [5,6]. With increasing altitude, the specific humidity increased by almost two orders of magnitude, reaching 0.08–0.1 g/kg at 30 km, also in agreement with [5,6].

In the third springtime case, the specific humidity changed relatively little with altitude (in the range 0.025–0.05 g/kg) decreasing in the lower layer of the

Table I. Maximum and Minimum Values of the Specific Humidity q (g/kg)* and the Content of Water Vapor in the Stratosphere $w(\mu)$ ** at the Mean Value of q .

Author	Date of observation	Coordinates of the point of observation	Center of the band, microns	q_{\min}	q_{\max}	w ($z > 25$ km)	
Hygrometric method							
Barret et al ^[5]	1. VII 1949	} 45° N 92° W		0.0034 (15 km)	0.154 (30 km)	27	
	26. VIII 1949			0.0120 (12 km)	0.150 (30 km)	28	
Murgatroyd et al ^[4]	7. I 1950	} 39° N 90° W		0.0300 (27 km)	0.090 (16 km)	21	
	6. VII 1954			0.0020 (15 km)	—	1	
Bracefield ^[6]	16. IX 1953	} 40° N 72° W		0.0218 (16 km)	0.292 (24 km)	54	
	18. XI 1953			0.0104 (15 km)	0.097 (24 km)	20	
Mastenbrook et al ^[8]	27. VI 1963	} 39° N 76° W		0.0024 (16 km)	0.096 (31 km)	17	
	28. IV 1959			40° N 105° W	0.0260 (27 km)	0.050 (24 km)	13
	8. IV 1960			39° N 76° W	0.0017 (16 km)	0.020 (24 km)	7
Barclay et al ^[10]	2. V 1958	52° N 0°30' W		—	0.037 (27 km)	13	
Spectroscopic method							
Neporent et al ^[12]			1.38	—	—	50 (12 km)	
Murcray	3. VIII 1956	33° N 106° W	1.37	—	—	35 (14 km)	
Brooks et al ^[14]	3. VIII 1956	33° N 106° W	1.38	—	—	94 (14 km)	
Gates, Murcray, et al ^[15]	22. VI 1955	33° N 106° W	1.87	—	—	10 (13 km)	
Houghton et al ^[16]	10. XII 1957	} 50° N 02° W		0.003 (14 km)	—	5.2 (14 km)	
	12. VI 1958			—	0.015 (12 km)	—	35 (12 km)
Murcray et al ^[17]	19. VI 1959	33° N 106° W	6.3	0.005—0.012 (12—28 km)	0.015—0.35 ($z > 28$ km)	20 (28 km)	
*The figures in the parentheses next to the values of q are the altitudes corresponding to the given concentrations.							
**The figures in the parentheses next to the values of w are the lower limits of the air column in which w was estimated.							

stratosphere and increasing in the upper one. It can thus be assumed in practice that in this case the water vapor has been uniformly distributed, in the mean, up to 30 km.

An increase in water-vapor concentration with increasing altitude was obtained also by Brown [6], who indicated that at 23 km the specific humidity was 38 times higher than at the 9 km level.

Recently Barclay, Elliott et al constructed in England a water-vapor trap capable of gathering air samples at altitudes up to 30 km. The investigated volume of air (up to 10⁴ meters) is blown through a tube cooled with liquid nitrogen, where the water vapor freezes out from the air together with the carbon dioxide. A measure of the investigated volume of air is the amount of frozen carbon dioxide, the content of which in the stratosphere is assumed to be constant at 0.025%. A shortcoming of the instrument is that it can determine the humidity only at one level, or to sum the moisture content in a definite region of the stratosphere. In [10] the specific humidity was measured at 27 km and found to be close to that obtained in [5, 6, 8] (q = 0.037 g/kg).

In Table I are summarized the minimum and maximum values obtained by different authors for the specific humidity in the lower layer of the stratosphere and at the limiting sounding levels. The data display some scatter, inasmuch as the measurements were made over different regions, in different seasons, and with different apparatus. The presently available material on the measurements of the vertical distribution of the water vapor and simultaneous measurements of the temperature and pressure profiles is insufficient to determine the reasons of these variations. For this purpose it would obviously be necessary to have a more extensive set of observations, which must include measurements of the characteristics of the horizontal and vertical mixing of the atmosphere and measurements of the air temperature with the aid of low-lag thermometers. However, even at the present stage we can conclude that two types of vertical distributions of water vapor exist in the stratosphere (up to 30–35 km).

First type. An abrupt drop in the water-vapor concentration above the tropopause to the very small

values of specific humidity, on the order of 10⁻³–10⁻² g/kg, which are reached in the lower layer of the stratosphere (to 16 km). Above this layer the water vapor concentration increases, and at the limiting altitudes z_{max} at which the observations have been made (See Table I), the specific humidity increases by one or two orders of magnitude.

This type of vertical distribution of the specific humidity can be approximated by the function

$$q(z) = \begin{cases} q_0 e^{-\alpha_0 z}, & 0 \leq z \leq z_1, \\ q_1 e^{\alpha_1(z-z_1)}, & z_1 \leq z \leq z_{\max}, \end{cases} \quad (4)$$

where z₁ is the height of the minimum of specific humidity, while q₀ and q₁ are the values of the specific humidity at the levels z = 0 and z = z₁, respectively.

In Table II are listed the values of q₀, q₁, α₀, α₁, and z₁ corresponding to different measurements of the specific humidity, and also the values of these parameters in the exponential Gunn's formula (2). As shown in Fig. 6, formula (4) describes satisfactorily the vertical distribution of the specific humidity. As expected, the parameter α₀, which characterizes the decrease in the specific humidity with altitude, turned out to be larger (approximately twice as large) than in Gunn's formula (2), and leads thus to an overestimate of the humidity in the troposphere and in the lower stratosphere and to an underestimate in the higher layers. What is striking is that α₀ and α₁ do not vary very strongly, except in the summertime measurements of Barret et al [5], for which too high values of α₀ were obtained. This circumstance is obviously connected with the large variations of the humidity in the troposphere, obtained in [5], so that the data of [5] cannot be satisfactorily approximated by means of the function (4). Indeed, if we first smooth out the curves of [5], then the values of α₀ (the numbers in the parentheses) will be the same as in the other cases.

The variations of q₀ and q₁ turned out to be significant. This means that the absolute values of the moisture concentration are always variable while the concentration variation with altitude is quite stable.

Second type. The water vapor is in the mean uniformly distributed in the stratosphere, and there is a relatively high concentration of moisture in the entire

Table II. Values of the Parameters in Formula (4).

Author	Date of observation	q ₀ , g/kg	α ₀ , km ⁻¹	q ₁ , g/kg	α ₁ , km ⁻¹	z ₁ , km	γ _{TP}	γ _{str}
Barret et al [5]	26.VIII 1949	12.2	0.84 (0.52)	0.012	0.08 (0.06)	12	0.88	0.34
	1.VII 1949	11.5	0.73 (0.49)	0.006	0.23 (0.22)	14	0.87	0.09
	7.I 1950	1.6	0.43	0.05	0	8	0.81	0.50
Bracefield [6]	18.IX 1953	5.1	0.42	0.01	0.27	15	0.81	0.07
	28.IV 1959	17.5	0.56	0.04	0	10	0.84	0.50
Mastenbrook et al [8]	8.IV 1960	3.6	0.51	0.002	0.18	14	0.83	0.16
	27.VI 1960	5.7	0.48	0.002	0.24	16	0.82	0.10
Gunn's formula		5.3	0.23	—	—	—	0.73	0.50
Murcra et al [17] (6.3μ band)	19.VI 1959	2.6	0.55	0.002	0.31	14	0.84	0.05

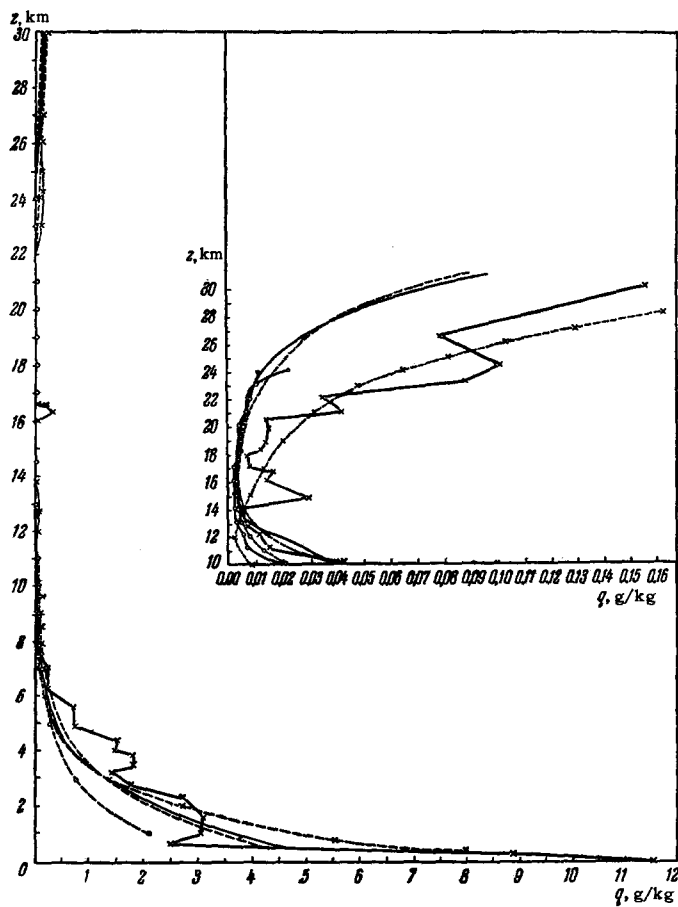


FIG. 6. Approximation of the experimental specific-humidity data in accordance with formula (4). — 27 June 1960^[6]; ○ — April 8, 1960^[6]; × — July 1, 1949^[5]; — experimental curves; --- in accord with formula (1).

investigated layer (the specific humidity varies within the limits 0.01–0.05 g/kg). To describe the vertical profile of this type we can use (4) in which we put $\alpha_1 = 0$ (See Table II).

The formation of these types of vertical water vapor profiles is apparently connected with the character of the mixing in the atmosphere and the thermal conditions of the stratosphere. For example, in the wintertime case mentioned above, where the specific humidity decreased slowly with altitude^[5] and which could be considered to some degree as a deviation from the second type, the temperature of the stratosphere did not increase with altitude as in the other cases, but, to the contrary, decreased somewhat, and reached low values (-59°C).

It is also of interest to estimate the total content of water vapor $w = \int_z^\infty \rho_w dz$ in the entire column of the atmosphere, 1 cm^2 in cross section, situated above a certain level z . Usually w is expressed in centimeters of precipitated water, numerically equal to the number of grams of water vapor in this column of air. Inasmuch as local measurements of the humidity en-

able us to determine the concentration or the density of the water vapor up to the maximum height reached by the hygrometer, the water vapor content above this level has been estimated by assuming that q is constant in the entire layer above this level and is equal to some average value which ranges between the minimum and maximum values of q indicated in Table I. In the calculations of w the lower limit of the air column was taken equal to 25 km, while the distribution of the air density ρ was assumed standard (the ARDC 1956 model). In those cases when the measurements yielded only the minimum values in the lower stratosphere, the obtained value of w characterized the content of water vapor in the dry stratosphere.

Estimates of the total content of water vapor in the stratosphere, given in Table I, are in satisfactory agreement with the values of w measured by the spectroscopic method.

3. SPECTRAL MEASUREMENTS

As is well known, water vapor has intense absorption bands in the infrared, centered about 0.94, 1.14, 1.38, 1.87, 2.7, 3.2, and 6.3μ , and also a series of rotational bands in the interval from 20μ to the microwave region.

By measuring the sun's spectrum at different heights in the atmosphere and at various resolutions it is possible to determine the water vapor content in the layer located above the level at which the measurement is carried out.

As is well known, the absorption bands of water vapor have a complex line structure. To resolve the absorption line in these water-vapor bands, the resolution required is on the order of 1 cm^{-1} . High resolution makes it possible to determine more accurately the positions and intensities of the absorption line and the distribution of the solar radiation of the spectrum.

The first measurement of the sun's infrared spectrum aimed at determining the contents of water vapor in the stratosphere were made on an airplane by Yarnell and Goody^[45] in England, and also on balloons by B. S. Neporent et al^[12,13] in the U.S.S.R. and by D. Murcray et al^[14] in the U.S.A.

In these investigations both prism and grating instruments were used; the image of the sun was guided onto the spectrometer slit with a heliostat. The optical resolutions of prism spectrometers do not exceed 100 cm^{-1} at 3μ , making it possible to determine the absorption over the entire band only.

An example of a prism optical hygrometer is the instrument of D. Murcray et al^[14]. The instrument is based on a Littrow type spectrometer with an LiF prism and a supercooled PbSe detector.

In order to compensate for the decrease in the energy incident on the slit with increasing wavelength, the Littrow mirror was connected with potentiometers, which changed the gain of the output stage of the amplifier in accordance with a prescribed program. The

signal from the output of the amplifier was fed to a galvanometer and registered on a 70 mm photographic film. The American investigators prefer to record on photographic film, since this eliminates additional errors ensuing from transmission and reception of the signal by radio. The presence of such errors was pointed out in the paper by B. S. Neporent et al.^[12], where an optical hygrometer with a diffraction grating is described for the first time. This instrument is based on a vacuum monochromator with diffraction grating having 300 lines per millimeter and measuring 50 by 70 mm, which separates five successive regions in the sun's spectrum (1.24, 1.4, 1.5, 1.8, and 2.2 μ). The wavelengths 1.24, 1.5, and 2.2 μ are between the absorption bands. By extrapolating the absorption over the altitude it is possible to determine the intensity of the sun's radiation incident on the top of the atmosphere. The linear dispersion of the instrument is 100 $\text{\AA}/\text{mm}$, and the width of the entrance and exit slits of the instrument is 1.5 mm. The receiver is a lead-sulfide cooled photoresistance. The signal from the amplifier is transmitted to the ground by radio-telemetry.

A better spectrometer with a resolution of approximately 2 cm^{-1} was proposed by Houghton, Moss, and Chamberlain^[46]. The optical system of the instrument is shown in Fig. 7. A $102 \times 76 \text{ mm}$ grating with 300 lines per millimeter, covering the region 1–5.5 μ , is used in the instrument. The fluorite prism of the preliminary monochromator has an angle of 15° . Rotation of the prism is synchronized with the rotation of the grating with the aid of an eccentric. The width of the intermediate and the exit slits is set by means of a micrometer, while the width of the entrance slit is constant at 1 mm. Preceding the entrance slit is a mechanical modulator, which interrupts the light flux at a frequency of 750 cps. To increase the stability of rotation, the modulator motor is fed from a special generator which is set with the aid of a tuning-fork circuit. The radiation receiver in the 2–3 μ region is a lead-sulfide photoresistance, while for longer wavelengths a tellurium-sulfide photoresistance is used, cooled with liquid nitrogen. A germanium filter is used to cut off radiation with wavelength shorter than 1.8 μ . The detector signal is amplified with a tuned amplifier.

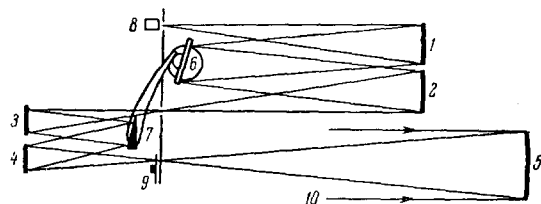


FIG. 7. Optical system of the hygrometer. Spherical mirrors: 1, 2—9 x 8 cm, focus 50 cm; 3, 4—6 x 6 cm, focus 25 cm; 5—15 x 5 cm, focus 70 cm; 6—grating 102 x 76 mm; 7—fluorite prism with 15° angle; 8—detector; 9—shutter; 10—light flux from heliostat.

One of the main difficulties when working with optical hygrometers is the development of a low-inertia heliostat. Particularly stringent requirements on the heliostats are imposed in the case of airplane measurements, owing to vibration of the airplane; if an instrument weighing 40–50 kg is raised by means of a bundle of balloons, the heliostat requirements are less stringent. The heliostat in the instrument of Houghton, Moss, and Chamberlain^[46] projected the sun's image, 6 mm in diameter, on the entrance slit with the aid of a system of mirrors. Four phototransistors were located on the edges of the light beam 5 cm ahead of the slits, and the light did not strike these phototransistors when in normal position. Any deflection of the beam caused one of the phototransistors to be illuminated and turn on through an amplifier a motor coupled to the mirror. Tests showed that the heliostat had a time constant of the order of 0.5 sec, making it possible to keep the sun's image on the slot satisfactorily if the beam deviations were not too fast ($\lesssim 10^\circ/\text{sec}$).

The instruments described above were used for measurements in the near infrared region of the spectrum only, up to 3 μ . Only recently did D. Murray and others^[17] get to 10 μ and measure the absorption of water vapor in the intense 6.3 μ band by raising a prism optical hygrometer to an altitude of 28 km. The instrument was based on a single-beam monochromator of the Littrow type with a NaCl prism and a bolometer receiver. The incident radiation was interrupted mechanically at a frequency of 80 cps, and the signal from the detector was amplified with a narrow-band amplifier and recorded on a four-track magnetic tape together with the reference signals. The spectrum from 1 to 10 μ was scanned within 18 seconds.

Thus, only a few of the above-described spectral instruments yield the sun's spectrum with sufficiently high resolution. The interpretation of the spectra is therefore made difficult in most cases by the fact that the spectral measurements are burdened with errors determined by the apparatus function. If the measured quantities are integrated over some sufficiently broad spectral interval, then, as shown by Nielsen, Thornton, and Dale^[47], the apparatus error becomes insignificant. However, other difficulties arise in the interpretation of integrated absorption characteristics.

As is well known, for monochromatic radiation the intensity I_ν (ν is the frequency) of the radiation passing through a layer of water vapor w is connected with the intensity I_ν^0 of the radiation incident on this layer by Bouguer's law:

$$I_\nu = I_\nu^0 e^{-k_\nu w}, \quad (5)$$

where k_ν is the absorption coefficient. If the absorption line has a Lorentz contour, then

$$k_\nu = \frac{S}{\pi} \frac{\alpha}{(\nu - \nu_0)^2 + \alpha^2}, \quad (6)$$

where S is the intensity of absorption of the line $\nu = \nu_0$ and α its half-width. All the calculations are usually carried out for the absorption function A_ν , which characterizes the fraction of the radiation absorbed in the layer w , and is determined by the relation

$$A_\nu = 1 - e^{-k_\nu w} \tag{7}$$

However, it is impossible to use expression (5) or (7) for a practical determination of w , owing to the above-mentioned apparatus errors. On the other hand,

the dependence of the integral absorption $\int_{\nu_1}^{\nu_2} A_\nu d\nu$ on w , which makes it possible to eliminate the apparatus error, is under real conditions quite complicated (ν_1 and ν_2 are the limits of the spectral interval over which the integration is carried out). It can be shown that in the case of an isolated line, for which relation (6) is satisfied, we have

$$\int_{\nu_1}^{\nu_2} A_\nu d\nu = S w \tag{8}$$

in the case of weak absorption (that is, small S or w) and

$$\int_{\nu_1}^{\nu_2} A_\nu d\nu = 2(S\alpha w)^{1/2}, \tag{9}$$

if the radiation absorption is large.

Analogous expressions are obtained also for the sets of nonoverlapping lines of the form (6). Inasmuch as the half-width of the line depends on the air pressure and the temperature, namely

$$\alpha = \alpha_0 \frac{p}{p_0} \left(\frac{T_0}{T} \right)^{1/2}$$

(α_0 is the half width at normal pressure p_0 and temperature T_0), the integral absorption also depends on p and T . For real conditions, when the line shape deviates from (6) and the lines overlap, the dependence of $\int_{\nu_1}^{\nu_2} A_\nu d\nu$ on w , p and T has not yet been sufficiently studied.

Usually to interpret the measurement of the amount of water vapor by the spectroscopic method, one uses

the dependences of the absorption function on w , p and T , obtained under laboratory conditions. Recently such investigations were carried out by Howard, Burch, and Williams [11,27] for the absorption bands of water vapor and carbon dioxide. By measuring the absorption of radiation in a multi-passage cuvette, with varying content of absorbing gas, temperature, and total and partial pressure, the authors of [11] ob-

tained empirical dependences of $\int_{\nu_1}^{\nu_2} A_\nu d\nu$ on w , p and T (the dependence of $\int_{\nu_1}^{\nu_2} A_\nu d\nu$ on T turned out to be extremely weak).

In the case of weak absorption, i.e., small w , p , or e , or a weak band, a good approximation is

$$\int_{\nu_1}^{\nu_2} A_\nu d\nu = c w^{1/2} (p + e)^k, \tag{10}$$

where c and k are certain empirical constants, characterizing a given absorption band or the spectral interval $\Delta\nu = \nu_2 - \nu_1$. In the case of strong absorption, that is, when w , p , and e are large or the band is intense, the analogous relations have the form

$$\int_{\nu_1}^{\nu_2} A_\nu d\nu = C + D \log w + K \log (p + e), \tag{11}$$

where C , D , and K are constants characterizing $\Delta\nu$.

To choose the proper formula, a limiting value of \tilde{A} is given in [11] for each band. If $\int_{\nu_1}^{\nu_2} A_\nu d\nu < \tilde{A}$, then formula (10) is used, and if $\int_{\nu_1}^{\nu_2} A_\nu d\nu > \tilde{A}$ formula (11).

The values of all the necessary parameters obtained in [11] are listed in Table III, which also indicates the range of the variation of w , p , and e for which formulas (10) and (11) were obtained. As indicated by the authors of [11], the error with which the experimental data are approximated by formulas (10) and (11) does not exceed 3%.

It is possible to neglect the partial pressure of water vapor in formulas (10) and (11), for under atmospheric conditions it amounts to not more than 5% of the total pressure. Comparing formulas (10) and (11) with formulas (8) and (9) we can readily verify

Table III. Values of the Parameters in Formulas (10) and (11) and the Limits of Their Applicability

Centers of H ₂ O bands, μ	Limits of variation			Limits of bands, ν_1, ν_2 , cm^{-1}	\tilde{A} , cm^{-1}	c	k	C	D	K
	w, cm	e , mm Hg	p, mm Hg							
0.94	0.09-1.0	5-28	10-740	10100-11500	200	38	0.27	—	—	—
1.1	0.05-1.8	3-28	10-740	8300-9300	200	31	0.26	—	—	—
1.38	0.03-3.8	3-28	5-740	6500-8000	350	163	0.30	202	460	198
1.87	0.01-3.8	3-29	3-740	4800-5900	275	152	0.30	127	232	144
2.7	0.02-2.4	2-25	2-760	3340-4400	200	316	0.32	337	246	150
3.2	0.02-2.4	2-25	2-760	2800-3340	500	40.2	0.30	—	—	—
6.2	0.01-1.5	2.5-25	2.25-742	1150-2050	160	356	0.30	302	218	157

that the dependence of the absorption function on w and p is different for a real spectrum than in the case of nonoverlapping lines with a Lorentz contour. For example, the square-root law is satisfied for an idealized spectrum in the case of large absorption, whereas under real conditions this law is satisfied for weak absorption. These differences are obviously connected with the effect of line overlap and the real absorption spectrum of water vapor.

It must be noted here that formulas (10) and (11) have definite application limits since they have been obtained only for a certain range of variation of w and p . In particular, for large w and p , formula (11) can yield absurd results: The mean value of the absorption function

$$A = \frac{1}{\nu_1 - \nu_2} \int_{\nu_1}^{\nu_2} A_\nu d\nu \quad (12)$$

may become larger than unity.

Formulas such as (10) and (11) are used for the interpretation of spectral measurements of humidity in the atmosphere. However, these formulas are obtained under laboratory conditions for values of w and e which are much larger than the content and partial pressure of the water vapor in the stratosphere (see Table III). Therefore, when using formula (11) for the interpretation of the spectral measurements in the stratosphere, one must unavoidably extrapolate the absorption function to the region of small w , e , and p . In addition, formulas (10) and (11) have been derived for values of p and e that are constant, albeit different, over the entire radiation path, whereas under real conditions the pressure changes over a wide range along the path of the sun's rays.

These circumstances make it very difficult to determine the vertical profile of the water vapor from spectral measurements. However, the main difficulty lies in the fact that the spectroscopic method itself has so far not yielded reliable data on the absorption of solar radiation at high altitudes. This is connected primarily with the use of the 1.3, 1.87, and 2.7 μ water-vapor bands. At low water-vapor contents, the absorption of solar radiation in these bands is less than the threshold sensitivity of the receiving apparatus, so that the measurements can be carried out only up to 12–15 km. These data therefore yield the distribution of water vapor only in the troposphere, and also yield estimates of the total moisture contents in the column of air above the level to which the spectrometer has risen.

As was already mentioned, Murcay et al.^[17] obtained data on the absorption of solar radiation not only in the near-infrared bands, but also in the intense 6.3 μ band, so that the sounding ceiling could be raised to 28 km. In Table IV are listed the results of the measurements of the absorption function in three water-vapor bands, obtained in^[17] and recalculated by formula (12). Inasmuch as the balloons drifted for

some time at high altitudes, several values of the absorption function were obtained for equal or slightly differing zenith distances of the sun.

What is striking are the substantial variations in the absorption function, which greatly exceed the measurement errors, which in turn are close to 10%. These variations of the absorption function are in qualitative agreement with the above-mentioned variations of the frost point in the case of local measurement (see Fig. 2). The reason of the variations of the absorption function can be the localized condensations of water vapor ("clouds"), which appear along the path of the solar radiation. Nor is the possibility excluded that the indicated variations are connected with inaccuracy of the heliostatic equipment (more about this later). Measurements of the frost point were carried out simultaneously with the spectral measurements. However, the data on the frost point were obtained only up to 12 km in^[17] and up to 15 km in^[16], so that the spectroscopic and local measurements of the humidity can be compared only up to these altitudes. The corresponding curves of the vertical distribution of the water-vapor density in Fig. 8 offer evidence of the relatively small differences up to 6 km; above that, the discrepancy becomes noticeable. It must be borne in mind here that the spectroscopic method cannot as yet yield a reliable altitude distribution of the water vapor, as follows from Table IV, which lists the total water vapor content as determined from simultaneous measurements of the absorption functions^[14] corresponding to different bands.

The appreciable differences in the values of w can be attributed to many circumstances.

1. Interpretation errors connected with an account of the changes in the air pressure and partial pressure of the water vapor along the path of the sun's rays to the radiation receiver. Since the value of the pressure at a given level cannot be used in this case in formulas (10) and (11) for the interpretation of the spectral measurements, account is taken in these formulas of the so-called "effective pressure." The effective pressure \bar{p} is obtained usually as the mean weighted pressure, the weight being some vertical

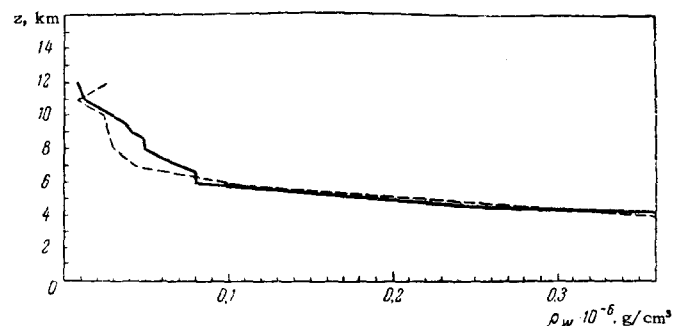


FIG. 8. Water-vapor density distribution obtained in^[17] by direct and spectroscopic measurements. — spectroscopic data, 6.3 μ band; - - - direct measurements.

distribution of the density of the water vapor $\rho_w = \rho q$, for example, by means of formula (4), that is,

$$\bar{p} = \frac{\int_0^z p q_w dz}{\int_0^z q_w dz}.$$

It is interesting to note that if we obtain the effective pressure from the water-vapor distribution given by (4), we get for the troposphere

$$\bar{p}_{tr} = p_0 \frac{\alpha_0 + \beta}{\alpha_0 + 2\beta} = \gamma_{tr} p_0, \quad (13)$$

and for the stratosphere

$$\bar{p}_{str} = p_1 \frac{\alpha_1 - \beta}{\alpha_1 - 2\beta} \frac{e^{(\alpha_1 - 2\beta)(z_2 - z_1)} - 1}{e^{(\alpha_1 - \beta)(z_2 - z_1)} - 1} = \gamma_{str} p_1 \quad (14)$$

(here β is the exponent in the barometric formula $p = p_0 e^{-\beta z}$, p_1 is the pressure at the level z_1 ; z_2 is some maximum altitude up to which $q(z)$ still increases). In Table II are given the values of γ_{str} and γ_{tr} for the corresponding distribution of water vapor and for $\beta = 0.13 \text{ km}^{-1}$.

The obtained values of p_{tr} and p_{str} turned out to be close to effective pressure values assumed in [15] ($\bar{p}_{tr} = 7p/9$ and $\bar{p}_{str} = p/2$, where p is the pressure at the given level).

In the calculations of w in [14], the effective pressure was assumed to be $\sim 60\%$ of the pressure at each given level.

The choice of any particular effective pressure does not influence noticeably the calculated value of w in the troposphere. However, in the interpretation of the spectral measurements in the stratosphere with the aid of formulas (10) and (11), the choice of the effective pressure indicated above leads even to absurd results. Let us consider, for example, the interpretation of the data obtained in [17] on the absorption function in the 6.3μ band. As can be seen from Table IV, this function decreases on the average very slowly with altitude, although the air pressure decreases strongly. Even if we disregard the problem of determining the content of water in the stratosphere from these data, the very fact that absorption decreases weakly with decreasing pressure is of tremendous interest, since it offers evidence that when the air pressure and water vapor tension vary along the path of the beam, the absorption of the radiation is described by relations other than of the type (10) and (11). In particular, the weak decrease obtained in [17] for the absorption function can be explained on the basis of the derivations of Sec. 2. If the first type of water-vapor distribution prevails in the atmosphere, then the main absorbing mass of water is concentrated in the layer above 25–30 km, and the value of w does not decrease in practice up to this altitude. Consequently, no matter how rapidly the air pressure decreases

altitude, the absorption function changes little. Moreover, it has been shown above that the relative concentration of the water vapor increases with altitude. Consequently, its pressure can also increase. Inasmuch as the absorption increases with increasing partial pressure of the water vapor, this can lead to a growth in the absorption function with altitude. Apparently the variations of the absorption function observed at all altitudes can also be attributed to the presence of regions with increased water-vapor concentration on the path of the ray.

All the foregoing gives some reason for stating that if the air and water vapor pressure vary along the path of the ray, the dependence of the absorption function on these factors will be described by formulas different from (10) and (11). In this connection notice must be taken of the fact that analogous results were obtained by A. M. Brounshtein [14] in measurements of the absorption function in the air layer next to the earth. For equal values of w but different water-vapor concentrations, the absorption function increased with increasing water-vapor concentration. A. M. Brounshtein attributed this to the fact that the effective diameter for the water-vapor molecule collisions exceeds by several orders of magnitude the effective diameter for the collisions between the water-vapor and air molecules. Therefore if a layer with increased water-vapor concentration exists in the stratosphere, this effect, which contributes to the broadening of the absorption line (and consequently to an increase in the absorption function), will compensate for the decrease in the absorption function with decreasing air pressure.

This conclusion can be of importance for the investigation of the temperature profile in the stratosphere, and particularly for an explanation of the existence of a maximum of the temperature layer at 40–50 km and a minimum at 80 km, inasmuch as the layer of air, which is saturated with water vapor, plays an important role in the radiant heat exchange. In particular, it can screen the radiation from the lower layers of the atmosphere.

On the other hand, as indicated above, the use of formulas of the type (10) and (11) for the determination of w can lead to serious errors (particularly in the stratosphere), since the air pressure and the partial pressure of the water vapor influence in different manners the absorption of radiation, and consequently cannot play comparable roles in the formulas for the absorption function. Indeed, the calculation of w made in [17], from data on the absorption of solar radiation in the 6.3μ band, with the aid of formula (14) for the same value of effective pressure as used in [14], yields the absurd result that the total content of water vapor has a minimum at a certain level and then increases with altitude (Table IV). In this connection, apparently, the calculations in [17] pertain not to the vertical variation of w as obtained from data on the absorption in

different bands, but to the so-called reduced water-vapor content

$$w_e = \int_z^{\infty} q_w \left(\frac{p}{p_0} \right)^{\varepsilon n} dz, \quad (15)$$

making it possible to dispense with an assumption concerning the value of the effective pressure. (Here p is the normal pressure and n is the exponent in formula (11), which is used as before for the calculation of w_p ; z is the attained altitude.)

To calculate w_e the stratosphere is divided into two layers (12–28 km and > 28 km), in each of which the concentration of water vapor is assumed constant).

Under this assumption it is found that the 12–28 km layer of the atmosphere contains on the average 10–12 μ of precipitated water, while in the layer above there is not less than 20 μ . The specific humidity in each layer depends on the choice of the exponent n (the authors of [17] choose two values of n , namely 0.3 and 0.5) within the following limits: in the 12–28 km layer $q = 0.005$ – 0.03 g/kg, and above 28 km we have $q = 0.15$ – 2.2 g/kg. Thus, these results, which approximate to some degree the real water-vapor distribution, agree sufficiently well with the data on the direct measurements of the humidity [5, 6, 8]. At the same time they show that in order to determine the total content of water vapor in the stratosphere from spectral-measurement data, principal significance attaches to a correct account of the dependence of the absorption function not only on the moisture content w itself, but also on the concentration of water vapor and the air pressure, which are variable along the path of the ray.

2. Differences in the values of w obtained from data on the absorption of solar radiation in different bands.

They can also be explained by the fact that the scanning of the spectrum consumes approximately 18 seconds, and in addition by the fact that the instrument rises at a rate of 20 m/min. Consequently, the absorption in the different bands can pertain to air masses with different water-vapor contents. It is likewise possible that the measured absorption of solar radiation is due to liquid water or to ice crystals, which appear from time to time on the path of the ray. Unfortunately, the absorption spectrum of water drops or of ice crystals has not yet been investigated sufficiently to be able to judge the error it introduces in the determination of the water-vapor content. However, if it is assumed that the water drops are absorbed in the same way as a film of liquid water, then in spite of the relatively low content of water aerosol in the atmosphere, the contribution to the absorption of solar radiation in the region of the water-vapor absorption bands will be noticeable. This is connected with the very strong absorption of water in the near infrared.

Inasmuch as the absorption of water increases with increasing wavelength in these spectral intervals

calculations of the water-vapor contents based on the absorption of the solar radiation in the 1.87 μ band can be expected to give exaggerated values of w . Nonetheless the values of w calculated in [14] from the absorption in the 1.38 μ band were systematically higher than the values of w corresponding to the 1.87 μ band. The authors of [14] do not indicate the causes of this result, but analogous calculations of w , which we carried out using the same data on the absorption functions given in [17] for the three bands 1.38, 1.87, and 6.3 μ , showed no systematic increase in w with decreasing wavelength bands (see Table III), although a large scatter in the values of w is obtained as before.

Nor can this excess be attributed to the influence of scattering of solar radiation in different parts of the spectrum on the calculation of w . As shown by the results of [13], when attenuation of radiation due to scattering is eliminated by extrapolation of the radiation intensity measured outside the band to the entire absorption band, no appreciable error in the calculation of w is introduced.

We are thus admittedly dealing with a systematic error connected with the very method used for the specific measurements in [14].

3. Accurate heliostating in measurements of humidity with a spectrometer.

If the guidance system misses the sun partially during the scanning time for 0.1–0.2 sec (a time interval on the order of the time constant of the heliostat), then the detector registers fictitious absorption in some interval of the spectrum. The width of the "absorption band" depends on the scanning rate. If this random interval does not overlap the water-vapor absorption lines, this "absorption" can be readily detected. In the case of overlap, on the other hand, the fictitious absorption cannot be eliminated. Murcay et al [14] used only those spectrograms in which there were patently no false absorption bands, but one can hardly speak of complete elimination of the errors. In spite of the uncertainties in the spectroscopic method, these measurements nevertheless make it possible to check the result of local humidity measurements at high altitudes. As already noted, the distribution of water vapor in the stratosphere (up to 28 km) can be obtained from all the spectral measurements only from the spectroscopic data of [17]. In all other cases the distributions are obtained only in the lower layers of the stratosphere near the tropopause.

Table I shows that the specific humidities in these layers are close to the analogous values obtained with the aid of hygrometers which measure the frost point. We recall that the specific humidity was obtained in [17] with the water vapor assumed constant in each of the considered layers of the atmosphere, (12–28 and > 28 km). Since the values of the specific humidity obtained in these layers differed in fact by almost two orders of magnitude, the actual altitude distribution of

the water vapor is inhomogeneous. These data confirm, both qualitatively and in order of magnitude, the existence of a new type of vertical water-vapor distribution in the stratosphere, as mentioned above.

The existence of such a humidity profile is also confirmed by measurement of the absorption function in the 1.38μ band^[12]. Namely, the weak variation in the absorption in the 12–17 km (maximum ascent) atmospheric layer obtained in this investigation is treated as being due to a rather slow change in the content of water vapor in this layer, although the authors of^[12] specially stipulate that these measurements are not sufficiently reliable.

In concluding this section, we make the following two remarks. Attempts were recently made in the USA to determine the content of water vapor in the stratosphere by spectral measurements in the interval $25\text{--}38 \mu$. To this end, a balloon-borne high-resolution spectrograph was used (a resolution of 1.5 cm^{-1} was reached in practice). As indicated in^[13], the 1960 tests were unsuccessful and it was proposed to repeat them in 1961.

An investigation of the content of water vapor in the stratosphere was also carried out^[20] by determining the absorption of L_α solar radiation. The coefficient of absorption of L_α radiation by water vapor is four orders of magnitude higher than the coefficient of absorption of dry air (390 and 0.044 cm^{-1} , respectively). This makes it possible to determine even very small moisture contents in the atmosphere. On the basis of these measurements, the concentration of water vapor in the upper stratosphere is found to be of the order of 10^{-1} g/kg .*

4. INDIRECT METHODS OF ESTIMATING HUMIDITY

In addition to the above results of direct and spectroscopic measurements of humidity, some information (essentially qualitative) on the water vapor content in the stratosphere can be obtained from an analysis of measurements of the integral fluxes of thermal radiation of the atmosphere (spectral region $> 3 \mu$), which have been carried out by many investigators^[21-24] at various altitudes up to 28 km. Figure 9 shows the results of measurements of vertical distribution of this radiation with the aid of special radiometers, with the receiving surface of one oriented downward and that of the other upward^[22,23]. The first radiometer measured the atmospheric radiation arriving from the upper hemisphere (descending radiation flux F^\dagger), while the other measured the radiation from the lower hemisphere (ascending radiation flux F^\ddagger).

It is easy to see that above the tropopause (12 km) the ascending flux is practically constant up to the 20 km level, and then a small increase in flux is ob-

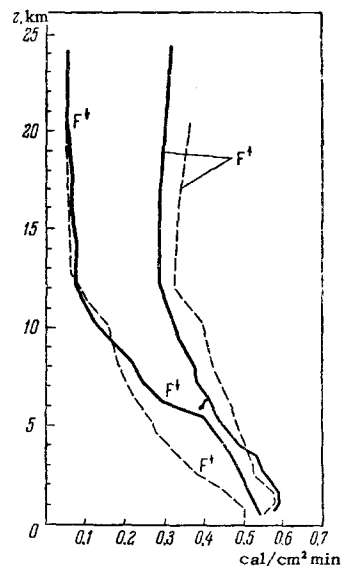


FIG. 9. Vertical distribution of fluxes of long-wave radiation. — July 30, 1958^[23]; July 7, 1958^[22].

served (by $0.02 \text{ cal/cm}^2 \text{ min}$). On the other hand, the descending flux changes very little in this layer. These peculiarities in the vertical distribution in the radiation fluxes offer evidence that the contribution of the lower layer of the stratosphere (10–20 km) to the thermal radiation is small, and the main radiating layer lies above 20–25 km.

The foregoing results of the measurements of the vertical distribution of the water vapor, and also the existence of relatively high temperatures in the stratosphere, give grounds for assuming that this layer is the 30–50 km mesosphere layer. It is assumed for the time being, that, unlike the troposphere where the ascending radiation is determined principally by the water vapor, the main contribution to the radiation of the atmosphere is made at these altitudes by the carbon dioxide and ozone. Unfortunately, direct measurements of the radiation of the atmosphere in the carbon-dioxide bands centered about 15μ ($12\text{--}18 \mu$) and the ozone bands centered at 9.6μ ($9\text{--}10 \mu$) at these altitudes, capable of disclosing the role of each of the three absorbing substances, are not available at present. Nonetheless, some estimates of the contribution from CO_2 and O_3 to the radiation of the high layers can be obtained by calculating the fluxes of descending radiation, if we specify the average distributions of these substances and of the temperature with altitude, and also their maximum possible variations in the earth's atmosphere. The results of such calculations made by Plass^[25,26] for two levels are listed in Table V. Inasmuch as an important role is played in these calculations by the account of the absorption spectrum, we have made analogous calculations using the latest data on the absorption of CO_2 and O_3 in the 15 and 9.6μ bands^[26,28] which were found to be close to the result of^[25,26].

*Recently Friedman^[43] obtained data on the absorption of L_α radiation, which makes these estimates doubtful (see below).

Table V. Fluxes of incident radiation (cal/cm² min).

Altitude, km	Measured fluxes $F \downarrow$ [22,23]	Calculated fluxes [25,26]			$\Delta F \downarrow = F \downarrow_{H_2O}$
		$F \downarrow_{CO_2}$	$F \downarrow_{O_3}$	$F \downarrow_{CO_2+O_3}$	
10	0.07	0.03	0.003	0.033	0.037
20	0.05	0.01	0.003	0.013	0.037
11	0.10	0.03	0.003	0.033	0.067
28	0.06	0.005	0.001	0.006	0.054
12	0.08	0.02	0.003	0.023	0.057
25	0.05	0.006	0.002	0.008	0.042

Table V lists also the values of the descending radiation fluxes $F \downarrow$, obtained by direct measurement in three cases [22,23] at two levels in each case, and consequently including the radiation of all the atmospheric gases. The total flux of radiation due to CO₂ and O₃ ($F \downarrow_{CO_2} + F \downarrow_{O_3}$) is everywhere much lower than the total descending flux ($F \downarrow$) (one order of magnitude lower at high altitudes), and the difference $\Delta F \downarrow = F \downarrow_{H_2O} = F \downarrow - (F \downarrow_{CO_2} + F \downarrow_{O_3})$ varies slowly with altitude, thus confirming the existence of the aforementioned main radiation layer in the mesosphere. This difference is obviously the contribution of the water vapor to the descending flux, and perhaps also the contribution of other substances (aerosol).

Some estimates of the content of water vapor w in the stratosphere layer lying above the level of the measured fluxes can be obtained from this difference, that is, from the value of the descending flux due to the water vapor ($F \downarrow_{H_2O}$)*.

As is well known (see, for example, [29]), this flux is connected with w by the relation

$$F \downarrow_{H_2O}(w) = \int_0^w B[T(u)] dP(w-u), \quad (16)$$

where $B(T) = \sigma T^4$ is the Stefan-Boltzmann function and T is the air temperature, regarded as a function of the water-vapor content, the latter being measured from the top of the atmosphere ($w = 0$ when $z = 0$); $P(w) = 1 - A(w)$ is the integral transmission function for water vapor. In general, the determination of w from (16) is not a simple matter, all the more since the dependence of the integral transmission function P on w and on the other parameters has not yet been sufficiently investigated. However, it is easy to estimate the value of w from the available data on $P(w)$. Indeed, if we take the value of B_{av} averaged over the stratosphere outside the integral sign in (16), then

$$F \downarrow_{H_2O}(w_{av}) = B_{av} [1 - P(w_{av})], \quad (17)$$

hence

$$P(w_{av}) = 1 - \frac{F \downarrow_{H_2O}(w_{av})}{B_{av}}. \quad (18)$$

If in addition we take the minimum and maximum values B_{min} and B_{max} , we get in perfect analogy

$$P(w_{min}) = 1 - \frac{F \downarrow_{H_2O}(w_{min})}{B_{max}}, \quad (19)$$

$$P(w_{max}) = 1 - \frac{F \downarrow_{H_2O}(w_{max})}{B_{min}}. \quad (20)$$

Using the dependence of the transmission function on w as calculated by Yamamoto [30], we can obtain from (18)–(20) respectively the average, minimum, and maximum values of w in the layer of the atmosphere located above the level of measurement of $F \downarrow_{H_2O}$.

If we take at the 20-km level $F \downarrow_{H_2O}(w) = 0.4$ cal/cm²min, $B_{av} = 0.3$ cal/cm²min ($T_{av} = 250^\circ$ K), $B_{min} = 0.2$ cal/cm²min ($T_{min} = 220^\circ$ K), and $B_{max} = 0.5$ cal/cm²min ($T_{max} = 280^\circ$ K), then $P(w_{av}) = 0.87$, $P(w_{max}) = 0.80$, and $P(w_{min}) = 0.92$. Consequently $w_{av} = 30 \mu$, $w_{min} = 6 \mu$, and $w_{max} = 60 \mu$, which is in sufficiently good agreement with the data given above on the local and spectroscopic measurements of the water-vapor content in the stratosphere.

5. POSSIBLE SOURCES AND SINKS OF MOISTURE IN THE STRATOSPHERE

The main conclusion, confirmed by different independent experiments, is that the water vapor distribution over the height of the stratosphere is such that makes it possible to distinguish between two layers; the lower stratosphere (10–20 km), in which the concentration of water vapor is low ($q = 10^{-3}$ g/kg), and the layer above, where the concentration increases by one or two orders.

By what physical mechanism does the concentration of water vapor increase in the mesosphere? To answer this question, and also to identify the sources and sinks of water vapor in the stratosphere, let us turn to results of optical soundings aimed at studying the structure of the high layers of the atmosphere.

As shown by projector soundings of the stratosphere [31], and also by direct measurements of the angular dependence of the intensity of the scattered light and the aerosol content [32–34], the layer up to 20–30 km contains a sufficiently large amount of aerosol, so that the scattering differs greatly from Rayleigh scattering. The appearance of mother-of-pearl clouds at these levels under certain meteorological conditions is also evidence of the presence of aerosol.

*The idea of this method of estimating was communicated to the authors by V. G. Kastrov.

logical conditions^[35] is evidence that this aerosol is of aqueous origin.

The reason for the aerosol being located at such high altitudes is apparently the turbulent and convective mixing of the atmosphere, and also possible volcanic eruptions and other gigantic explosions on the earth's surface, which carry away much moisture from the atmosphere, in the form of liquid or vapor, into the upper layer. In the lower stratosphere, where low temperatures predominate, the water vapor condenses or freezes, thus explaining the relative dryness of this layer of the atmosphere. In the mesosphere layer, however, where high temperatures (on the order of 270–300°K) are observed at low air pressures, the water vapor cannot condense^[36]. Moreover, the water drops or crystals reaching there from the lower layers of the atmosphere evaporate. Thus, the mesosphere layer at 30–60 km is a unique trap in which water vapor can accumulate.

The high concentrations of water vapor and the high temperature in the mesosphere are apparently related. Indeed, the absorption of thermal radiation from the earth by the water vapor is generally one of the causes of the heating of this atmospheric layer. The concept of increased water-vapor concentration in the mesosphere is in agreement with two other geophysical phenomena observed in the upper stratosphere.

The first is the already mentioned formation of silver clouds at the 80 km level; at present there are two different hypotheses (dust and condensation) concerning the origin of these clouds^[36-38]. According to the dust hypothesis the silver clouds are accumulations of dust particles of terrestrial or cosmic origin. Certain arguments are advanced in favor of this hypothesis in^[36].

According to the condensation hypothesis, the silver clouds consist of ice crystals. To confirm this hypothesis is first necessary to have a definite concentration of water vapor and sufficiently low temperatures at the level of formation of the silver clouds, so as to ensure formation of ice crystals. For example, if the relative concentration of water vapor at the 80 km level is of the same order as in the lower stratosphere (10^{-3} g/kg), then the temperature at which this vapor becomes saturated and sublimation is possible (the frost point) should be 145°K. Such a temperature, according to available rocket sounding data^[39,40] is extremely rarely encountered at the 80 km level. On the other hand, if the relative concentration of water vapor is one or two orders of magnitude higher, then the vapor becomes saturated at 150–160°K. Such temperatures were observed by rocket soundings in the region of the minimum (Fig. 10). If we assume that the values of the water-vapor concentration obtained by different authors (see Table I) remain constant up to the mesopause, then we can estimate the frost point at different altitudes, as was done in^[41]. Comparing the altitude variation of the

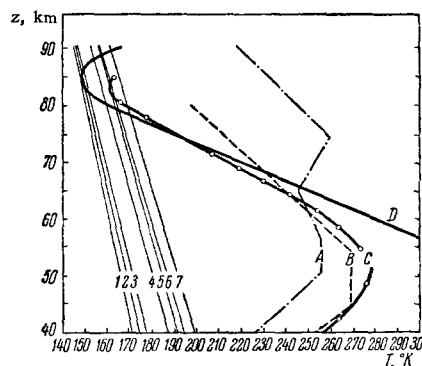


FIG. 10. Extrapolation of the frost point to large altitudes and vertical profile of the temperature, obtained by rocket measurements. Temperature: A – Nordberg et al^[46]; B – Nordberg et al^[46]; C – Stroud et al^[39]; D – Mikhnevich^[40]. Frost point: 1 – Murgatroyd et al^[4]; 2 – Houghton et al^[46]; 3 – Gates et al^[15]; 4 – Barclay et al^[10]; 5 – Barret et al^[5]; 6 – Mastenbrook et al^[6]; 7 – Murcay et al^[7].

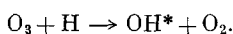
frost point with the vertical distributions of the air temperature, we can readily verify that the most favorable conditions for the formation of silver clouds can exist at the 80–82 km level (See Fig. 10).

On the other hand, in examining the difference between the frost point and the air temperature obtained at these altitudes in different latitudes and in different seasons, we obtain a conclusion of importance to the condensation hypothesis, namely that the conditions for the formation of silver clouds in high altitudes are more favorable in the summer, when this difference is minimal.

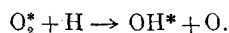
Thus, the existence of a mesosphere layer with relatively high concentration of water vapor as a source of moisture at high altitudes is an argument in favor of the condensation hypothesis of the origin of silver clouds. This layer can also make another contribution to the formation of silver clouds, in that long-wave radiation can be beamed from its upper boundary out into space, thus causing, as pointed out by I. A. Khvostikov^[38], cooling of the lower layers of the atmosphere and consequently the appearance of very low temperatures at the upper boundary of the stratosphere. The freezing of the moisture accompanying the formation of silver clouds is one of the possible sinks of water vapor which accumulate in the 30–60 km mesosphere layer. When the clouds become scattered, perhaps owing to dynamic phenomena (wind, vertical mixing) or to the precipitation of the particles, only some of the ice crystals will drop into the warmer layers of the atmosphere and evaporate there. The evaporation can occur also at the silver-cloud level, owing to the heating of the cloud by advection of heat. Some part of the water vapor (or ice crystals) will remain in one way or another at the 80 km altitude and above. It is well known (see for example,^[42]) that at these altitudes the water vapor already dissociates under the influence of the ultraviolet radiation from the sun. The dissociation of the molecules of water

and the escape of the hydrogen atoms outside of the earth's atmosphere is the second possible water-vapor sink.

We have arrived at the second phenomenon in which the important part is played by water vapor, namely the hydroxyl emission of the night sky. At the present time there are two hypotheses concerning the origin of the hydroxyl emission^[42]. According to the so-called ozone-hydrogen hypothesis of Bates, Nicolet and Herzberg, the excited OH* hydroxyl is the result of the reaction



According to the oxygen-hydrogen hypothesis proposed by V. I. Krasovskii^[42], it is assumed that the hydroxyl radiation occurs when the vibration-excited oxygen molecules interact with the hydrogen atoms



Without touching upon the question of the advantages of any one of the theories, we note only that to produce any of the foregoing reactions it is necessary that the atomic-hydrogen contents be appreciable at the levels where these reactions occur. According to the estimates in^[42], the ratio of concentration of hydrogen atoms to the total number of particles necessary to ensure the oxygen-hydrogen reaction at altitudes on the order of 80 km should exceed 10^{-4} – 10^{-5} . This quantity agrees with the data given above on the content of water vapor in the stratosphere.

Recently Friedman^[43] obtained rocket-measurement data on the absorption of solar radiation in the L_α line at 134–163 km, which make it possible to estimate the total content of hydrogen in the column comprising the atmosphere and the interplanetary gas between the sun and the measurement level. According to these data, the average concentration of the hydrogen atoms in this column is three or four orders of magnitude lower than the quantity mentioned above. It must be noted that such low average concentrations of atomic hydrogen as obtained in^[43] cannot cast doubts on the existence of increased local concentrations of atomic hydrogen at lower levels, particularly on the upper boundary of the stratosphere. A discussion of this question, however, is beyond the scope of the present article.

6. CONCLUSION

The described direct measurements of the vertical distribution of water vapor in the stratosphere indicate that there exists a layer of mesosphere, lying above 25–30 km, in which the relative moisture concentration increases with altitude by 1–2 orders of magnitude over the value in the lower stratosphere. Indirect estimates of the water-vapor content, based on measurements of the flux of long-wave radiation in the stratosphere, and also based on an analysis of

the condition for the formation of silver clouds and hydroxyl radiation, are in agreement with these data.

Satisfactory agreement was observed between the results of the measurements, which make it possible to describe the vertical distribution of the water-vapor concentration by means of a simple exponential function (4), the parameters of which have a scatter that can be readily attributed to the difference in the conditions of the individual measurements. Nonetheless, the conclusions derived must be regarded as most tentative, inasmuch as the observational material is still sporadic and in many cases insufficiently reliable.

To obtain more reliable information on the moisture content in the stratosphere it is necessary to carry out simultaneous measurements with the aid of different methods.

As can be seen from the foregoing, it is extremely important to measure the distribution of the water vapor at high altitudes. In this connection, one of the most promising is the spectroscopic measurement method with the aid of rockets and artificial earth satellites. However, this method calls for the development of reliable methods of interpreting the measured data.

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